PRECURSORY GROUNDWATER LEVEL CHANGES

OF THE BOHEMIAN MASSIF (CENTRAL EUROPE) IN 2005

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ABSTRACT

We analyse continuous measurements of groundwater level in two deep wells VS-3 and V-28 at the experimental hydro-meteorological station situated on the NE margin of the Bohemian Massif, central Europe, characterized by the weak intraplate seismic activity. The aim of our study is to examine the relationships between changes in the groundwater level and earthquake occurrence. Based on the tidal and barometric response of the water level, we estimated selected elastic parameters of the observed aquifers: the shear modulus G, the Skempton ratio B, the drained matrix compressibility β *and the undrained compressibility* β_u*. Using these parameters and assuming the homogeneous poroelastic material, we derived the sensitivity of the wells to the crustal volume strain. During the observation period from November 1998 to December 2005 we detected in the VS-3 well two pre-seismic steps, related to August 10, 2005 (M = 2.4) and October 25, 2005 (M = 3.3) earthquakes. Amplitudes of the recorded precursory changes (+6 cm and +15 cm) are several times higher than the values predicted from the theoretical precursory crustal strain and the strain sensitivity of the well. Therefore, we presume that the observed pre-seismic water level steps can be attributed to heterogeneity of poroelastic material. We consequently propose the hypothesis of the origin of precursory events based on the presumption of a sensitive site, at which the well is situated.*

Key words: seismic activity, earthquake precursors, groundwater, crustal deformation, Earth tides

1. INTRODUCTION

Hydrogeological effects of seismic activity are the results of anomalous migration of fluids or pore-pressure variations in the Earth's crust due to seismo-tectonic processes. These effects are observable as fluctuations of various quantitative as well as qualitative hydrogeological parameters, such as water well level, springs discharge, chemical or isotopic composition, temperature, electric conductivity, turbidity and others. Seismic-induced fluctuations of groundwater parameters were described by many authors from different seismoactive regions of the world (see, e.g., *Thomas, 1988*; *Igarashi and Wakita, 1990*; *Roeloffs and Quilty, 1997*; *Grecksch et al., 1999*; *Gavrilenko et al., 2000*; *Chandha et al., 2003*).

This paper considers only seismic-induced changes in groundwater level in wells. These changes are generally believed to reflect stress variation in the Earth's crust (see, e.g., *Kümpel, 1992*). According to the origin time of an earthquake and groundwater level anomaly, we distinguish pre-, co-, and post-seismic groundwater level changes. From the point of view of earthquakes prediction, the most important are the pre-seismic anomalies which can serve as earthquake precursors. Many of the reported precursory anomalies were reviewed by *Kissin (1982)*, *Roeloffs (1988)*, *Kissin and Grinevsky (1990)* or *King et al. (2006)*. These papers summarize basic characteristics of pre-seismic groundwater level changes, like the size of the anomaly, lead time of occurrence and possible relations between the earthquake magnitude or epicentral distance and the amplitude of an anomaly. Reported amplitudes of the anomalies vary from several centimetres to several metres and, similarly, the precursor times range from less than one day to several months or even years. The epicentral distances at which the seismic-induced hydrogeological effects are observed reach max. several hundreds of kilometres. The analysis of the spatial distribution of groundwater level anomalies performed by *Montgomery and Manga (2003)* reveals a relatively clear dependency of this distance on the earthquake magnitude. To define the maximum distance at which the hydrogeological effects of earthquakes can be detected, the value of seismic-induced volume strain $\varepsilon > 10^{-8}$ was used by the authors. In general, groundwater level anomalies are associated not only with strong earthquakes. For example, *Kissin et al. (1996)* or *Leonardi et al. (1997)* reported pre-seismic and co-seismic well level changes induced by seismic events with magnitudes $M < 3$.

The present work has been carried out in the area of the Hronov-Poříčí Fault Zone (Bohemian Massif, Czech Republic) and describes the relationships between relatively weak intraplate seismic activity and groundwater level fluctuations recorded in two experimental deep wells. In the studied area, the hydrogeological effects of seismicity have never been systematically monitored. Their occurrence has been mentioned only in terms of an assessment of macroseismic effects of strong seismic events. The most significant changes in groundwater parameters were described by *Woldřich (1901)* in connection with the January 1901 earthquake ($M = 4.6$). Anomalous fluctuations of water level and turbidity were observed in dug wells as far as 60 km from the earthquake epicenter. The reported changes sustained for several hours or even days after the main shock.

In the following sections we deal with the determination of elastic properties of the observed aquifers and strain sensitivity of wells, based on the water level response to the

Earth tides and air pressure fluctuations. Then, we discuss the origin of the precursory water level changes recorded before two earthquakes in August and October 2005 with regard to the properties of observed aquifers and theoretical pre-seismic deformations. Finally, we propose a mechanism of the origin of the observed earthquake-induced water level anomalies.

2. GEOLOGICAL AND TECTONIC SETTINGS OF THE STUDY AREA

Bohemian Massif - one of the most prominent central European Variscan structures - belongs to areas with a relatively low seismicity. Magnitudes of the strongest events in this intraplate region do not exceed the value of local magnitude $M_L = 5$. Seismoactive zones of the Bohemian Massif are limited to its marginal parts, where young tectonic movements (Late Tertiary – Early Quaternary) caused the uplift of marginal crustal blocks and the generation of mountain chains, framing the almost aseismic central part of the massif (*Procházková et al., 1986*). The two most seismically active areas are situated on the NW and NE margins of the Bohemian Massif. The first one, known as the West Bohemia/Vogtland zone, is situated in the area of intersection of two prominent fault systems - the Eger Rift and the Mariánské Lázně Fault (see, e.g., *Bankwitz et al., 2003*; *Geissler et al., 2005*). This area is characterized by the occurrence of seismic swarms, which usually include several thousands of events (see, e.g., *Fischer and Horálek, 2000*; *Fischer, 2003*; *Fischer and Michálek, 2008*). The second most seismically active area is situated on the NE margin of the Bohemian Massif. It is represented by a zone of generally NW-SE orientation, approximately 40–60 km wide and 150 km long, which comprises a number of NW-SE and NNW-SSE-striking faults. This zone forms a SE termination of the important central European tectonic structure - the Elbe Fault system (EFS). The EFS (Fig. 1) has been described as a NW-SE-striking fracture zone extending from the North Sea to the eastern margin of the Bohemian Massif (see, e.g., *Scheck et al., 2002*; *Špaček et al., 2006*). It is assumed that the EFS represents a zone of crustal weakness of considerable age. According to *Scheck et al. (2002)*, crustal deformation along the EFS has taken place repeatedly since Late Carboniferous times. The most intense deformation took place during the Late Cretaceous – Early Cenozoic time, when the EFS responded to regional compression with an uplift of up to 4 km.

In comparison with the West Bohemia/Vogtland zone, the seismoactive region of the SE termination of the EFS is characterized by less frequent occurrence of seismic events. Smaller earthquake swarms in this area were reported by *Špaček et al. (2006)*, but these micro-swarms do not number more than 50 weak events ($M \leq 1.3$). The strongest earthquakes occur on the NW margin of this seismoactive region and are connected with movements along the Hronov-Poříčí Fault Zone (*Kárník et al., 1984*; *Procházková et al., 1986*; *Schenk et al., 1989*).

Hronov-Poříčí Fault Zone (HPFZ) is a system of parallel fractures, dividing two important structural units - the Intra-Sudetic Basin and the Krkonoše Piedmont Basin (Fig. 1). The NW-SE-striking fault zone is approximately 40 km long and up to 500 m wide. It is bounded by the Vrchlabí lineament in the north and by the Nová Paka lineament in the south. Both E-W-striking faults are supposed to be sinistral strike slips (cf. *Schenk et al., 1989*). The contemporary HPFZ is a result of complicated and long-

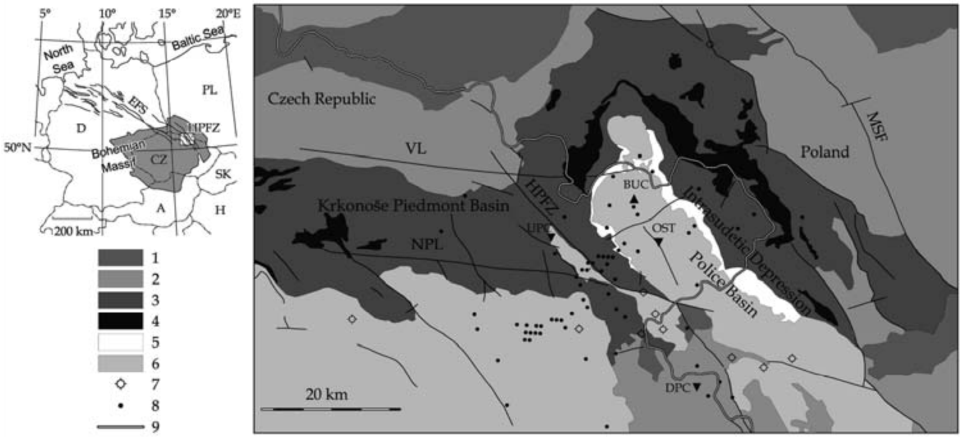

Fig. 1. Geology and tectonics of the study area (after *Biely et al., 1968*; *Jetel and Rybářová, 1979*; *Schenk et al., 1989*; *Scheck et al., 2002*; *Cymerman, 2004*). 1 - plutonic rocks (granites, granodiorites), 2 - metamorphites (gneisses, schists, granulites, migmatites), 3 - Permian and Carboniferous sediments, 4 - Permian volcanics, 5 - Triassic sediments, 6 - Cretaceous sediments, 7 – CO_2-rich mineral springs, 8 - epicentres of seismic events recorded from 1985 to 2005, 9 - state boundary. HPFZ - Hronov-Poříčí Fault Zone, MSF - Marginal Sudetic Fault, EFS - Elbe Fault System, VL - Vrchlabí Lineament, NPL - Nová Paka Lineament, BUC - hydro-meteorological station Bučnice, DPC, OST, UPC - seismic stations.

lasting evolution, which began in the late Paleozoic. Since then, several tectonic phases have taken place. The fault zone has been successively developed from an asymmetric anticline, whose steeply inclined SW arm was axially disrupted due to the regional compression by a reverse fault (*Tásler, 1979*). Along this fault the NE block was relatively uplifted. The main reverse fault is accompanied by parallel or oblique high-angle dislocations (normal or reverse faults), grouped under the term Hronov-Poříčí Fault Zone. The youngest (Cenozoic) tectonic movements in the area of the HPFZ are traceable by means of geomorphological methods. The study of drainage network evolution and perturbations in fluvial terraces revealed tectonic movements along normal faults dating back to Pliocene/Pleistocene to Early Pleistocene (see, e.g., *Stejskal et al., 2006*).

The relatively frequent local seismic activity is a proof of the present mobility of the HPFZ. Macroseismic effects of historical earthquakes in this area reached the epicentral intensity $I_o = 7°$ three times during the last 300 years (June 30, 1751, December 11, 1799 and January 10, 1901 - *Kárník et al., 1957*). The strongest historical earthquake of January 10, 1901 reached the magnitude of 4.6 and was felt over an area of 50 000 km^2 (*Woldřich, 1901*). The isoseists of local earthquakes are elongated mostly NW-SE, parallel to the orientation of the HPFZ. The depth of foci is mostly between 5 and 15 km (*Schenk et al., 1989*). Instrumental monitoring of seismic activity focused on local earthquakes began in the 1980s. Since 1984 were recorded in the area of HPFZ 78 earthquakes in the magnitude range $M = 0.0 – 3.4$.

Another significant proof of the increased endogenic dynamics of the study area is the occurrence of CO_2-rich mineral springs. A major role in the transport of the mantle-derived CO_2 is played by deep reaching faults with high vertical permeability like HPFZ or some other local fractures (*Jetel and Rybářová, 1979*). In general, the mineral springs in the area of the HPFZ belong to a larger zone called the Náchod-Klodzko mineral water region, which partly extends to the territory of Poland (Fig. 1).

A possible explanation of the present mobility of the HPFZ was given by *Schenk et al. (1989)*. According to this local geodynamic model, the HPFZ as a reverse fault balances the compression caused by the movements along the Nová Paka and Vrchlabí lineaments, bounding the HPFZ in the north and south. This presumption is supported by the analyses of repeated triangulation and precise levelling performed in the broader vicinity of the HPFZ by *Vyskočil (1988)*. Results of the repeated geodetic measurements indicate compression tendencies across the HPFZ. Moreover, the repeated precise levelling at two lines crossing the HPFZ revealed two anomalous uplifts across the fault zone, which preceded seismic events of May 7, 1984 (M = 3.4) and October 20, 1985 (M = 3.0). The more recent data on crustal deformation in the broader area of the HPFZ are available due to GPS measurements along the Marginal Sudetic Fault (MSF), running parallel to the HPFZ along the NE margin of the Bohemian Massif (Fig. 1). Preliminary results of the GPS monitoring indicate the NE-SW compression tendencies, perpendicular to the MSF and HPFZ (*Kontny, 2004*).

3. OBSERVATIONS

3.1. Monitoring of Seismic Activity

At present, three seismic stations are in operation in the broader area of the HPFZ: Úpice (UPC), Dobruška-Polom (DPC) and Ostaš (OST). The UPC and DPC stations belong to the Czech regional seismic network and are operated by the Institute of Geophysics (GfÚ) in Prague. Data from these stations are recorded in a continuous mode with sampling frequency of 20 Hz. The third station - OST - has been operated by the Institute of Rock Structure and Mechanics (IRSM) since 2005. This station was designed as a small aperture array containing three satellite short periodic sensors SM6b and a broadband sensor Güralp CMG-40T as a central point. Data are recorded in a continuous mode with sampling frequency of 100 Hz. All seismic data (i.e., co-ordinates, magnitudes and origin times of seismic events) used in this paper were taken from the Catalogue of regional seismic events, compiled by the GfÚ (the catalogue is published only in the electronic form at *http://web.ig.cas.cz/en/seismic-service/catalogs-of-regional-seismic-events/*).

3.2. Hydro-Meteorological Observations

Most hydro-meteorological data used in this study have been collected at the experimental station Bučnice, located approximately 9.5 km northeast of the surface trace of the HPFZ. The station is situated in the territory of an Upper Cretaceous sedimentary unit - the Police Basin. The basin extends in the NW-SE direction, parallel to the HPFZ and is a part of the higher-order structural unit - the Intra-Sudetic Basin (cf. Section 2). The maximum thickness of the Upper Cretaceous sediments in the Police Basin is

450–500 m (*Krásný et al., 2002*), and the underlying strata consist of Triassic, Permian and Carboniferous deposits.

The hydro-meteorological station Bučnice has been operated since 1963 by the Water Research Institute (WRI) in Prague. There are measured basic meteorological parameters, groundwater level in observation wells, and discharge of the two nearby streams - the Metuje River and the Zdoňov Creek. Since 1998, most meteorological parameters have been recorded by an automatic meteorological station. At present, we dispose of time series of the following meteorological data: precipitation amounts, air temperature and humidity, snow cover thickness, solar radiation, wind speed and wind direction.

The main purpose of the hydro-meteorological monitoring at the Bučnice station is the data acquisition for: (1) development and testing of the methods of modelling the local hydrological balance (i.e., relations between precipitation, evapotranspiration, runoff and groundwater discharge); (2) assessment of the long-term changes in water resources related to the changes in climatic parameters; (3) protection of the surface water and groundwater resources. With respect to the relatively frequent occurrence of earthquakes in the area of the HPFZ, we analyse the relations between the occurrence of seismic events and fluctuations of the groundwater level in two deep wells VS-3 and V-28 situated in the area of the experimental station.

Observation Wells and Water Level Measuring Technique

The deep observation wells VS-3 (305 m) and V-28 (300 m), were drilled in the valley of the Metuje River, only 540 m from each other. Both wells tap water bearing horizons in Upper Cretaceous sediments; nevertheless, they differ in the vertical range of screens (i.e., the open parts of the casing). The VS-3 well is opened at depths of 38.38–207.06 m and 216.65–260.13 m, and the V-28 well at a depth of 157.2–211.75 m. Thus the total length of screens of the VS-3 and V-28 wells is 212.96 m and 54.55 m, respectively. If we compare the vertical sections of the two wells (cf. Fig. 2), we can see that the V-28 well is opened only to the Lower Turonian aquifer, whereas the VS-3 well is opened also to aquifers in the Middle Turonian and Cenomanian sediments. The uppermost member of the Cenomanian - the chert series tapped by the VS-3 well at the depth of 234 m - represents the most important water-bearing sedimentary unit of the Police Basin. The aquifer consists of silt- and sand-dominated silicites (cherts) with fracture porosity; its thickness reaches maximum 15 m. It is characterized by the highest permeability (hydraulic conductivity $k = 1 \times 10^{-4}$ to 1×10^{-3} m/s) and a hydraulic continuity over a large area of the northern part of the basin, as proved by pumping tests (*Krásný et al., 2002*).

The groundwater level measurements in the VS-3 and V-28 wells are fully automatic from November 1998 on. Both wells are equipped with water level sensors comprising the measuring unit (pressure transducer) and the data storage unit with a capacity of 2 MB. Resolution of the measurements is 1 cm. Sampling period was initially set to 2 hours and changed for 1 hour on October 2005. Additional measurements of air pressure, which considerably influences the groundwater level fluctuations, are carried out at the OST seismic station, located approximately 7.5 km SE of the Bučnice station. Data are recorded automatically with a sampling period of 10 minutes. Resolution of the measurements is 0.001 hPa.

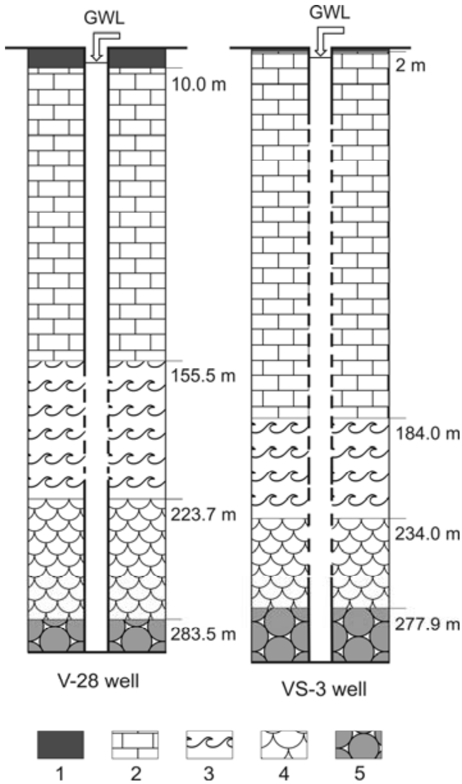

Fig. 2. Observation wells VS-3 and V-28. 1 - Quaternary alluvial deposits. 2 - Middle Turonian marlstones and silty sandstones. 3 - Lower Turonian marlstones. 4 - Cenomanian silicites, marlstones and sandstones. 5 - Triassic sandstones. Broken line - open parts of the well casing, GWL - groundwater level.

4. WATER LEVEL DATA PROCESSING

Water level data processing is based on decomposition of the initial measured signal (i.e., directly observed groundwater level) into four components: barometric response (*BAR*), tidal response (*TID*), low-frequency component (*LFC*) and high-frequency non-tidal residuals (*HFR*). The observed water level *OBS* is decomposed as the following form:

$$
\begin{aligned}
OBS &= BAR + NBAR, \\
NBAR &= LFC + HFC, \\
HFC &= TID + HFR, \\
OBS &= BAR + TID + LFC + HFR,
\end{aligned}
\tag{1}
$$

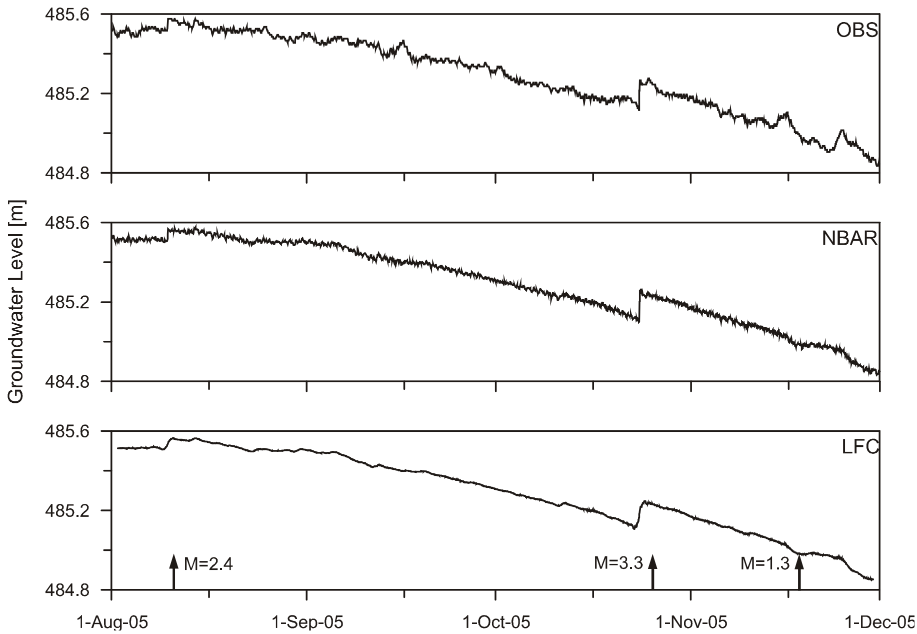

Fig. 3a. Decomposition of the observed groundwater level in the VS-3 well: *OBS*, *NBAR* and *LFC* components. Period: August 1, 2005 to December 1, 2005. Arrows indicate the times of local earthquakes (*M* - earthquake magnitude). Note the pre-seismic steps recorded before August 10 (M = 2.4) and October 25 (M = 3.3) earthquakes.

where *NBAR* is the groundwater level after removing the effects of air pressure and *HFC* is the high-frequency component of *NBAR*. The low-frequency component *LFC* (below 0.8 cpd) and the high-frequency component *HFC* (above 0.8 cpd) are separated by numerical filtering. For this purpose we use the *Pertsev (1959)* low-pass filter of 50 hours length. The decomposition of the observed well water level is shown in Figs. 3a,b. The technique of the estimate and the character of the tidal and barometric response is discussed in more detail in the two following sections (4.1. and 4.2.).

4.1. Barometric Response of Groundwater Level

We determined the barometric response of the groundwater level using the method elaborated by *Lyubushin (1994)* and *Lyubushin and Latynina (1994)*. This method is based on nonparametric estimate of frequency-dependent complex transfer function $H(f)$ from atmospheric pressure to water level data. The transfer function $H(f)$ is estimated within each position of moving time window of a certain length L samples which is shifted by one sample from left to right direction of the time axis. The estimate of $H(f)$ is given by:

$$H(f) = Sxy(f) / Syy(f) , \qquad (2)$$

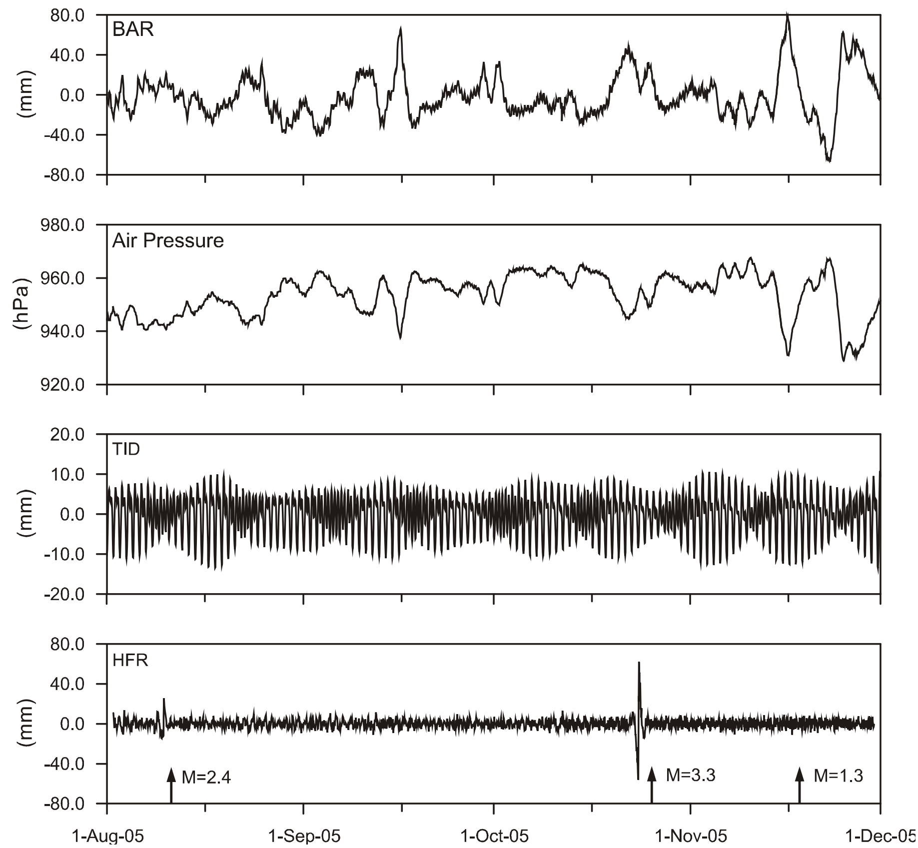

Fig. 3b. Decomposition of the observed groundwater level in the VS-3 well: air pressure and *BAR*, *TID* and *HFR* components. Period: August 1, 2005 to December 1, 2005. Arrows indicate the times of local earthquakes (*M* - earthquake magnitude).

where $Sxy(f)$ is an estimate of a complex cross-spectrum between the water level and the air pressure, $Syy(f)$ is an estimate of power spectra of the air pressure. The barometric response BAR of the groundwater level is found as the following inverse Fourier transform:

$$BAR = \frac{1}{F}\left[H\left(f\right)AY\left(f\right)\right], \tag{3}$$

where $Y(f)$ is the Fourier transform of the air pressure time series. The absolute values $\left|H\left(f\right)\right|$ of the transfer function represent the amplitude response of the groundwater level to the air pressure as a function of frequency. An almost constant amplitude response is typical for both the VS-3 and V-28 wells at low and intermediate frequencies (Fig. 4). Nevertheless, at higher frequencies (between approximately 3.0 and 12.0 cpd) the

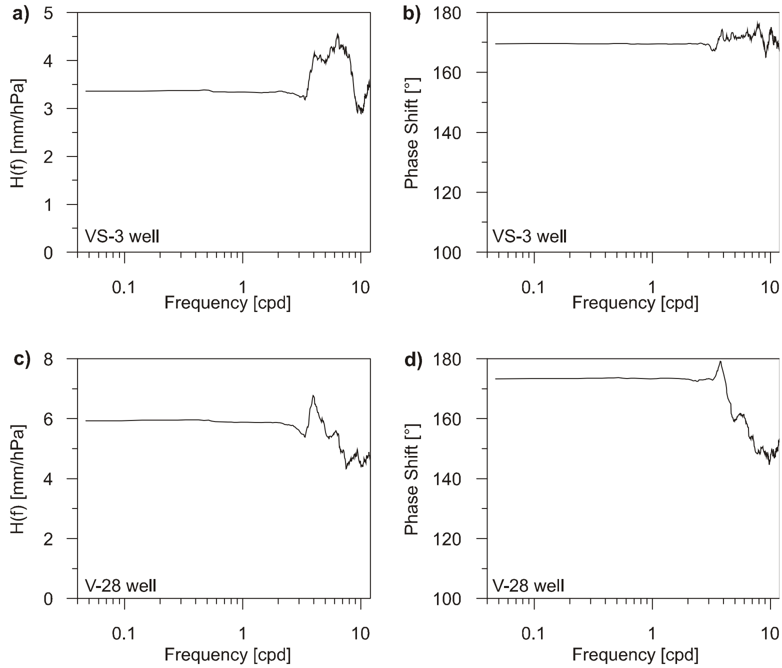

Fig. 4. The amplitude response (**a** and **c**) and phase response (**b** and **d**) of the water well level to the air pressure as a function of frequency.

amplitude response is considerably less stable. For the VS-3 well it is generally higher, with maximum values between 4.0–4.5 mm/hPa in the frequency range 4.0–8.0 cpd. For the V-28 well the amplitude response reaches the maximum of 6.8 mm/hPa at 3.9 cpd and then it gradually decreases.

The phase lags between the groundwater level and the air pressure variations as a function of frequency (Fig. 4) are given by the argument $\arg\left(H\left(f\right)\right)$ of the transfer function. For the VS-3 well, the phase lags are relatively stable at different frequencies, reaching the value $170 \pm 6°$. For the V-28 well a general increase is observed in the frequency range 3.5–12.0 cpd. At frequencies lower than 3.2 cpd, the phase lag is almost constant, reaching the value of $173 \pm 1°$. The relative stability of the amplitude and phase response beginning at subtidal frequencies indicates the high degree of confinement of the observed aquifers with the overlying rocks. Lower values of phase lags at higher frequencies (up to 3.5 cpd) in the V-28 well can be explained as inertial effects of groundwater flow from/to the well (see, e.g., *Quilty and Roeloffs, 1991*; *Igarashi and Wakita, 1991*).

4.2. Tidal Response of Groundwater Level

The tidal constituent of the groundwater level signal was determined from the tidal analysis of the high-frequency component of the air pressure corrected data *HFC*. The tidal analysis was carried out for five main wave groups (O1, K1, N2, M2, S2) according to *Tamura's (1987)* development with 1200 waves, using a modified ETERNA 3.0 program (*Wenzel, 1993*). The global tidal model of an ellipsoidal, rotating, elastic and oceanless Earth "Wahr-Dehant-Zschau", considering the imperfect elasticity of the Earth's mantle, was used (for details see *Wahr, 1981*; *Dehant, 1987*; *Zschau and Wang, 1981*). Based on this model, theoretical values of the volume tidal strain were calculated in order to determine the amplitude factors and phase differences between the theoretical and observed tidal waves. The tidal constituent *TID* represents the sum of all tidal waves in the frequency range of 10.822–32.743 deg/h, whose frequencies are given by the used tidal model, and the amplitudes A_o and phases Φ_o are calculated using the formulae:

$$A_o = A_m F_A , \tag{4}$$

$$\Phi_o = A_m + D_f , \tag{5}$$

where A_m and Φ_m are the amplitudes and phases of the model values of volume tidal strain, F_A is the amplitude factor, and D_f is the phase difference calculated for each analysed wave group. The expression for the tidal constituent *TID* is then given by:

$$TID(t) = \sum_{i=286}^{1121} A_{oi} \left(\cos f_i t + \Phi_{oi} \right) , \tag{6}$$

where t is time in hours, f is the frequency of the wave and i is the order of the wave according to the Tamura's development. Since we use only the waves from the above mentioned frequency range in the *TID* constituent, index i in Eq.(6) ranges from 286 to 1121.

The results of tidal analyses are listed in Tables 1 and 2. Analogous to the barometric response, amplitudes of all the five tidal waves are somewhat higher in the V-28 well. The maximum variations are induced in both wells by the semi-diurnal M2 wave with the frequency of 28.984 deg/h. The phase differences of tidal groundwater level variations are relatively low compared to the model values of volume tidal strain; however, a relatively high error of the estimate must be taken into account (Tables 1 and 2). The phase differences fall within the range of approximately ±30° (i.e., ±2 hours). The negative values of the phase difference indicate a lag of the observed tidal variations of groundwater level behind the theoretical values of the relative volume tidal strain, whereas the positive values indicate that the observed values precede the theoretical ones.

5. DETERMINATION OF ELASTIC PROPERTIES OF THE AQUIFERS

According to *Rice and Cleary (1976)*, there are four main elastic parameters governing the response of the porous rock saturated with a compressible fluid to stress or pore pressure changes. These parameters are the shear modulus *G*, the Poisson ratio for drained (ν) and undrained (ν_u) conditions, and the Skempton ratio *B*. Comprehensive treatment of all these poroelastic variables can be found together with bellow stated Eqs.(7)–(13) in

Table 1. Results of tidal analysis of groundwater level measurements in the VS-3 well.

Tidal Constituent	Period [h]	Frequency [deg/h]	Amplitude		Amplitude Factor [mm/nstr]	Phase Difference [deg]
			Model [nstr]	Observations [mm]		
O1	25.819	13.943	6.794	2.789 ± 0.617	0.411 ± 0.091	25.709 ± 12.666
K1	23.934	15.041	7.113	3.799 ± 0.617	0.534 ± 0.087	13.913 ± 9.317
N2	12.658	28.440	1.325	0.770 ± 0.338	0.581 ± 0.255	–0.204 ± 25.152
M2	12.421	28.984	6.918	4.647 ± 0.338	0.672 ± 0.049	2.460 ± 4.167
S2	12.000	30.000	3.219	2.145 ± 0.338	0.666 ± 0.105	–14.087 ± 9.025

Table 2. Results of tidal analysis of groundwater level measurements in the V-28 well.

Tidal Constituent	Period [h]	Frequency [deg/h]	Amplitude		Amplitude Factor [mm/nstr]	Phase Difference [deg]
			Model [nstr]	Observations [mm]		
O1	25.819	13.943	6.794	3.260 ± 0.508	0.480 ± 0.075	18.019 ± 8.926
K1	23.934	15.041	7.113	4.785 ± 0.508	0.673 ± 0.071	–0.316 ± 6.087
N2	12.658	28.440	1.325	1.518 ± 0.335	1.146 ± 0.253	1.066 ± 12.629
M2	12.421	28.984	6.918	6.427 ± 0.335	0.929 ± 0.048	0.720 ± 2.983
S2	12.000	30.000	3.219	2.251 ± 0.335	0.699 ± 0.104	–30.450 ± 8.512

Wang (2000). Nevertheless, in this section we also refer to previously published, mostly original, papers.

When the fluid flow in the reservoir can be neglected, the pore pressure change relative to the volume strain can be estimated according to *Rice and Cleary (1976)* at:

$$\Delta p = (2GB/3)\left[(1+\nu_u)/(1-2\nu_u)\right]\Delta\varepsilon\,, \tag{7}$$

where Δp is the change in reservoir fluid pressure and $\Delta\varepsilon$ is the change in volume strain in the reservoir (compression is negative). The above stated Eq.(7) enables to determine the response of groundwater level to volume strain of any origin, including the tectonic strain.

Assuming a well to tap an aquifer which is infinite in extent and consists of porous, permeable rocks bounded below and above by impermeable rocks, elastic parameters of the aquifer can be estimated based on the response of the groundwater level to Earth tides and atmospheric pressure (see, e.g., *Rojstaczer and Agnew, 1989*; *Roeloffs, 1996*). The response of the groundwater level to Earth tides and air pressure variations is defined by the tidal strain sensitivity A_s, and the barometric efficiency E_b (cf., e.g., *Roeloffs, 1988*):

$$A_s = -\Delta h/\varepsilon_t\,, \tag{8}$$

$$E_b = -\Delta h/\Delta p_b\,, \tag{9}$$

where Δh is the water level change, ε_t is the volume tidal strain and Δp_b is the change in barometric pressure. As a value of the tidal strain sensitivity A_s we take, in accordance with *Roeloffs (1988), Rojstaczer and Agnew (1989)* or *Roeloffs (1996)* the amplitude factor F_A for the M2 wave, estimated by means of the tidal analysis (cf. Tables 1 and 2). This way, we obtain A_s = 0.67 mm/nstr for the VS-3 well and A_s = 0.93 mm/nstr for the V-28 well.

The barometric efficiency E_b was estimated based on the frequency-dependent complex transfer function $H(f)$ between atmospheric pressure and water level data (cf. Section 4). For the value of E_b used in Eq.(9), we adopted the maximum of the quasi-constant course of the amplitude response at lower periods. Hereby, we obtain E_b = 3.6 mm/hPa for the VS-3 well and 6.0 mm/hPa for the V-28 well. Uncertainty of these values expressed by the standard deviation yields 0.04 mm for the VS-3 well and 0.12 mm for the V-28 well.

Once we determined the tidal strain sensitivity A_s and the barometric efficiency E_b, we can obtain the Skempton ratio B according to *Rojstaczer and Agnew (1989)* or *Igarashi and Wakita (1991)* as:

$$B = \frac{\rho g A_s \beta}{1 + \rho g A_s \left(\beta - \beta_u\right)}, \tag{10}$$

where $\rho = 1.0 \times 10^3$ kg/m^3 is water density, g = 9.8 m/s^2 is the acceleration of gravity, β is the drained matrix compressibility and β_u the undrained compressibility. Adopting the values of $\nu = 0.25$ and $\nu_u = 0.3$, we derived β and β_u solving the system of the two following equations from *Rojstaczer and Agnew (1989)* and *Kümpel (1991)*:

$$\beta = \frac{1 - E_b}{E_b}\left(\frac{3}{2\rho g A_s \left(1+\nu\right)} - \beta_u\right), \tag{11}$$

$$\beta_u = \frac{\left(1 - 2\nu_u\right)\left(1+\nu\right)}{\left(1-2\nu\right)\left(1+\nu_u\right)}\beta \tag{12}$$

Eventually, the shear modulus G was obtained based on parameters β and ν (e.g., *Kopylova and Boldina, 2006*):

$$G = \frac{3\left(1-2\nu\right)}{2\beta\left(1+\nu\right)}. \tag{13}$$

Table 3. Elastic parameters of observed aquifers estimated from the response of the water well level to Earth tides and atmospheric pressure. Uncertainty of B, G, β and β_u parameters can be expressed using the $\Delta h_m/\Delta h_{obs}$ ratio (see Tables 4 and 5), which yields approx. 8% for the VS-3 well and 16% for the V-28 well.

Well	A_s [mm/nstr]	E_b [mm/hPa]	β 10^{-10} Pa^{-1}]	β_u [10^{-10} Pa^{-1}]	G [10^{10} Pa]	B
VS-3	0.67 ± 0.05	3.6 ± 0.04	1.370	1.054	0.438	0.747
V-28	0.93 ± 0.05	6.0 ± 0.12	0.587	0.452	1.022	0.476

Table 4. VS-3 well - comparison of the model values of tidal water level fluctuation Δh_m derived on the basis of Eq.(7) with the observed values of the tidal water level fluctuation Δh_{obs} derived from the tidal analysis. $\Delta\varepsilon_t$ is the theoretical volume tidal strain in the area of the observation wells.

Tidal Wave	$\Delta\varepsilon_t$ [nstr]	Δh_m [mm]	Δh_{obs} [mm]	$\Delta h_m/\Delta h_{obs}$
O1	6.794	3.001	2.789	1.076
K1	7.113	4.088	3.799	1.076
N2	1.325	0.828	0.770	1.076
M2	6.918	5.000	4.647	1.076
S2	3.219	2.308	2.145	1.076

Table 5. The same as in Table 4, but for the V-28 well.

Tidal Wave	$\Delta\varepsilon_t$ [nstr]	Δh_m [mm]	Δh_{obs} [mm]	$\Delta h_m/\Delta h_{obs}$
O1	6.794	3.772	3.260	1.157
K1	7.113	5.538	4.785	1.157
N2	1.325	1.757	1.518	1.157
M2	6.918	7.437	6.427	1.157
S2	3.219	2.605	2.251	1.157

The resulting elastic parameters, tidal strain sensitivity A_s and barometric efficiency E_b are listed in Table 3. Introducing now the parameters B, G and ν_u into Eq.(7), we obtain the resulting response of water level to the volume strain as Δh = 0.72 mm/nstr for the VS-3 well and Δh = 1.08 mm/nstr for the V-28 well.

Although the values of B, G, β and β_u are generally acceptable for sedimentary rocks, it must be taken into account that the obtained values are only rough estimates. Therefore, we tested the accuracy of the estimate of G and B by the calibration of Eq.(7), based on the substitution of theoretical values of the volume tidal strain as $\Delta\varepsilon$. Then, we compared the resulting model values of the pore pressure change converted to the water level change as $\Delta h_m = \Delta p/(\rho g)$ with tidal fluctuations of the groundwater level determined by the tidal analysis (Δh_{obs}). The calculated $\Delta h_m/\Delta h_{obs}$ ratios (see Tables 4 and 5) are surprisingly low, which indicates a satisfactory degree of accuracy of the estimate of G and B parameters. The $\Delta h_m/\Delta h_{obs}$ ratio yielded 1.076 for the VS-3 well and 1.157 for the V-28 well.

The differences between the model and observed tidal water level fluctuation can be explained by the fact that the real values of the volume tidal strain in the area of the observation wells are not known. In Eq.(7), only the theoretical values resulting from the global tidal model of the Earth are used as $\Delta\varepsilon$, and the local conditions are not considered. Nevertheless, it must be pointed out that there are some alternative methods of estimating the local Earth tide strain. For example *Masterlark et al. (1999)* provide a method of tidal strain estimate based on autoregression analyses of the pore pressure time series data.

6. EARTHQUAKE-RELATED GROUNDWATER LEVEL CHANGES

From 1998, when the automatic monitoring of the VS-3 and V-28 wells was started, were observed two distinct precursory groundwater level changes associated with two earthquakes, relatively strong for the given region. The earthquakes were recorded on August 10, 2005 at 18:54:34 UTC ($M = 2.4$) and on October 25, 2005 at 10:51:57 UTC ($M = 3.3$). Focal mechanisms of both earthquakes could not be estimated, due to the limited number of local seismic stations in the studied area. Both events were accompanied by several weak shocks (Table 6). Nevertheless, the first foreshock before the August 10 main event does not precede the water level anomaly, and the October 25 main event was not accompanied by any foreshocks, only by a series of five aftershocks. The observed anomalous groundwater level changes can be therefore considered to be the earthquakes' precursors. The precursory changes were recorded in the VS-3 well only, and had the character of a sudden step-like water level rise. These anomalous steps were not followed by any detectable co-seismic or post-seismic water level variations (Figs. 5 and 6).

Other earthquakes recorded between the years 1998 and 2005 (see Fig. 7 for positions of epicentres) were not accompanied by similar anomalous groundwater level response. Selected parameters of seismic events recorded during this period are listed in Table 6. These data indicate that the August 10, 2005 and the October 25, 2005 events are 1–4 orders of the seismic energy higher than the other seismic events, which explains the lack of the precursory water level change before the rest of earthquakes recorded since 1998.

The parameters of the observed precursory groundwater level changes listed in Table 7 indicate the existence of a possible relation between the earthquake magnitude and the amplitude and the lead time of the anomaly. The stronger October 2005 earthquake was preceded by an anomaly with the amplitude 2.5 times higher, and the lead time of the

Table 6. Parameters of seismic events recorded in the period 1998–2005 in the area of the HPFZ. N - number of seismic events, M - magnitude of the main event, d - distance between the epicentre of the main event and the VS-3 well, E - seismic energy of the whole group derived according to *Tobyáš and Mittag (1991)* as $\log E = 1.2 + 2.0\,M$, [N, E] - coordinates of the main event.

Group of Events	N	M	d [km]	E [J]	Coordinates of the Main Event
E1: Jun 24, 1999	1	2.2	14.9	3.98×10^5	50.49°N, 16.06°E
E2: Sep 5, 1999	1	1.0	12.6	1.58×10^3	50.51°N, 16.07°E
E3: Dec 2–5, 2003	15	1.7	21.7	4.33×10^4	50.43°N, 16.04°E
E4: Aug 10, 2005	24	2.4	11.3	1.05×10^6	50.59°N, 16.16°E
E5: Oct 25, 2005	6	3.3	16.8	6.31×10^7	50.47°N, 16.07°E
E6: Nov 7, 2005	1	1.3	12.6	6.31×10^3	50.50°N, 16.13°E

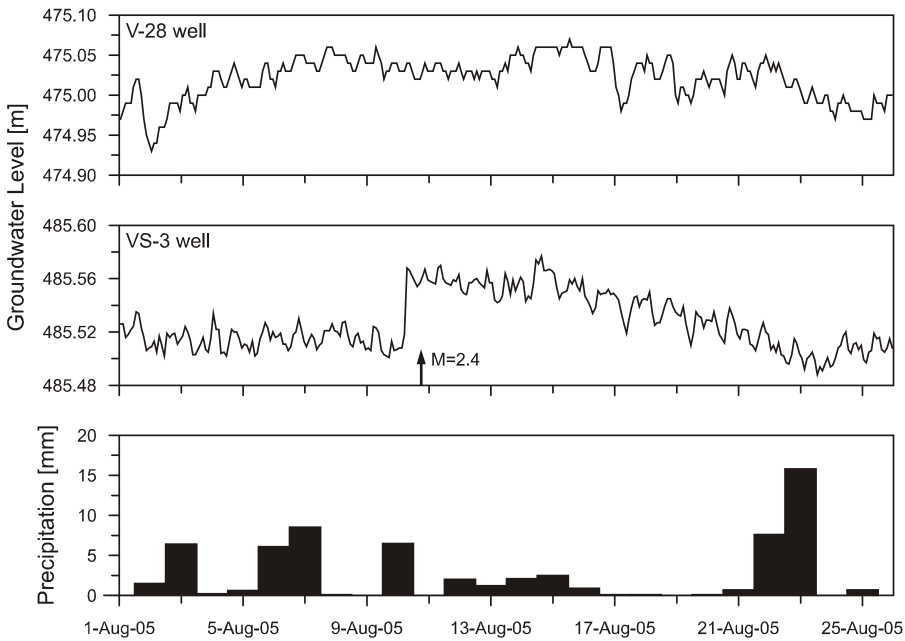

Fig. 5. Pre-seismic water level change recorded in the VS-3 well before the August 10, 2005 earthquake. The groundwater level is plotted against the precipitation daily amounts and the water level in the V-28 well. The air pressure-corrected water level data are shown.

anomaly occurrence was more than two times longer. These relationships have, however, no statistical significance for the presence, because we do not dispose of any other examples of the precursory groundwater level changes.

Assuming the fact that the precursory water level changes are the results of the pre-seismic volume crustal strain, we can calculate the magnitude of this strain in the area of the VS-3 well adapting Eq.(7) in the following way:

$$\Delta\varepsilon = -\frac{3}{2}\frac{\Delta p\left(1-2\nu_u\right)}{GB\left(1+\nu_u\right)} . \tag{14}$$

Since the precursory response has the form of a water level rise, the corresponding strain is considered to be a compression (see, e.g., *Roeloffs, 1988*). The values of the precursory strain derived from Eq.(14) and listed in Table 7 seem to be rather high. We therefore used the empirical formula of *Dobrovolsky et al. (1979)* to obtain the rough estimate of the precursory strain given only as a function of magnitude and epicentral distance. For earthquakes with magnitude $M < 5$, it holds:

$$\Delta\varepsilon_e = \left(\frac{10^{0.5M-3.06}}{d}\right), \tag{15}$$

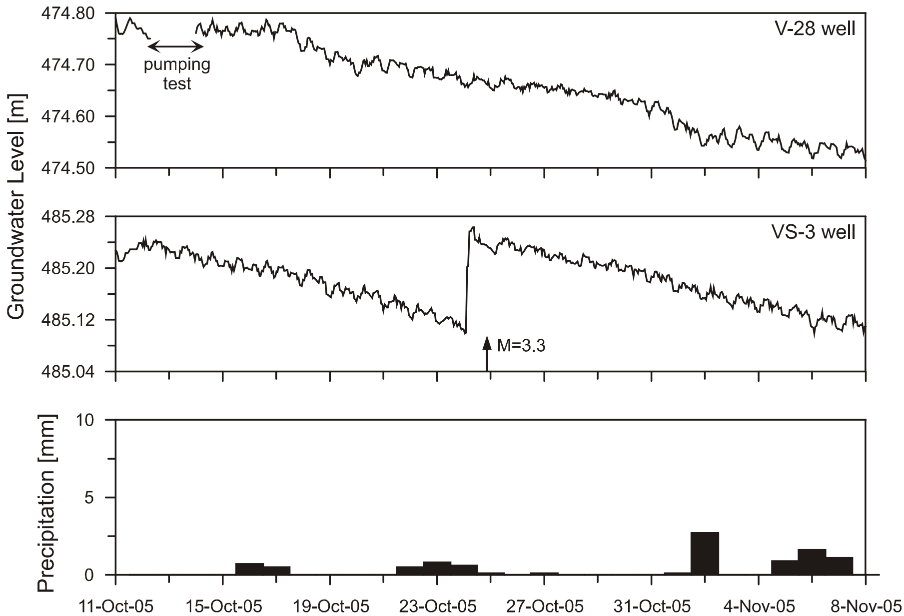

Fig. 6. Pre-seismic water level change recorded in the VS-3 well before the October 25, 2005 earthquake. The groundwater level is plotted against the precipitation daily amounts and the water level in the V-28 well. The air pressure-corrected water level data are shown.

where $\Delta\varepsilon_e$ is the precursory strain and *d* is the epicentral distance in kilometres. The theory of *Dobrovolsky et al. (1979)* formulates empirical relations between the precursory strain, the earthquake magnitude and the epicentral distance. It assumes that the earthquake precursors are detectable over an area, where the precursory strain exceeds the value of $\Delta\varepsilon_{t\,e} = 10^{-8}$. The relations resulting from this theory are used when studying the earthquake-related groundwater level changes, e.g., by *Kissin et al. (1996)* or *Montgomery and Manga (2003)*. Nevertheless, we have to point out that the estimate of earthquake precursory strain based on Eq.(15) doesn't take into account some important parameters like focal mechanism or heterogeneities in the Earth's crust.

The values of precursory strain calculated from Eq.(15) are considerably lower than the values derived from the precursory water level change (cf. Table 8). The precursory strain is $\Delta\varepsilon_{t\,e} = 1.823 \times 10^{-9}$ for the August 2005 earthquake and $\Delta\varepsilon_{t\,e} = 1.242 \times 10^{-8}$ for the October 2005 earthquake. Introducing these values into Eq.(7), we obtain the corresponding water level changes (cf. Table 8). For the October 2005 earthquake, the simulated water level changes in both wells are about 1 cm, which corresponds to the resolution of the used water level sensors. In August 2005, the simulated precursory changes are lower than 2 mm and thus potentially undetectable. In analogy with *Igarashi and Wakita (1991)*, *Kissin et al. (1996)* or *Grecksch et al. (1999)*, we come to the conclusion that the amplitudes of the observed earthquake-induced groundwater level anomalies are considerably higher than the amplitudes derived from theoretical volume

Table 7. Precursory groundwater level changes in the VS-3 well preceding the August 2005 and October 2005 seismic events; d - epicentral distance, Δh_e - preseismic water level rise, Δt_e - time of the origin of the anomaly before the earthquake, $\Delta \varepsilon_t$ - precursory deformation (compression) in the area of the VS-3 well estimated from the water level change.

Earthquake	d [km]	Δh_e [cm]	Δt_e [hours]	$\Delta \varepsilon_t$
Aug 10, 2005	11.3	+6.0	11–15	8.302×10^{-8}
Oct 25, 2005	16.8	+15.0	32–29	2.075×10^{-7}

Table 8. Model values of the water level response Δh_m to the precursory volume strain $\Delta \varepsilon_{t\,e}$ calculated on the basis of Eq.(15).

Seismic Event	$\Delta \varepsilon_{t\,e}$	Δh_m [cm]	
		VS-3	V-28
Aug 10, 2005	1.823×10^{-9}	0.13	0.20
Oct 25, 2005	1.242×10^{-8}	0.90	1.34

strain caused by the earthquake. This implies that the pre-seismic water-level steps recorded in the VS-3 well cannot be easily explained assuming the presence of homogeneous poroelastic material. Such an interpretation would not explain the lack of the pre-seismic response in the V-28 well, which has the strain sensitivity of approximately 1.5 times higher than the VS-3 well (see Section 5).

7. HYDROLOGICAL INTERPRETATION AND DISCUSSION

Considering the low theoretical precursory strain related to earthquakes accompanied by the pre-seismic water level anomalies, we are proposing the hypothesis of the origin of precursory events based on the presumption of the existence of a sensitive site, at which the VS-3 well is located. The existence of sensitive sites where unexpectedly high amplitudes of earthquake-related water level changes are observed is supposed by e.g., *King et al. (2006)*, *Kissin et al. (1996)* or *Kümpel (1992)*. In all the three studies, the sensitive sites are characterized as structurally weak zones surrounded by relatively more stiff material, which are often situated near tectonic faults.

We have already mentioned the connection of the VS-3 well with the Cenomanian chert aquifer, which is characterized by high permeability, fracture porosity and well documented regional continuity (see also Section 3.2.). Besides that, *Krásný et al. (2002)* suppose the presence of so-called preferential zones of groundwater flow within the chert aquifer. These zones are represented by a dense system of highly permeable fissures, permitting flow of considerably higher amounts of water, than in the less permeable surrounding material. The transport velocity of water within the preferential zones is also substantially higher. Using a numerical model of the groundwater flow, *Krásný et al.*

(2002) estimated the transport velocity at max. 15 m/day within the preferential zones and at 0.1 m/day outside the preferential zones - in less fractured silicites.

Formulating now the hypothesis on the VS-3 well as a sensitive site, we must presume the connection of the well with a preferential zone in the chert aquifer. Its presence in the area of the well can be a result of the more intensive fracturing of the aquifer in the neighbourhood of tectonic dislocations running near the well (see Fig. 7). Hereby, the observed pre-seismic anomalies can be considered to be a result of modification, most probably the compression, of preferential zones in the chert aquifer. The compression resulted in the lowering of the volume of fluid-filled fractures of the preferential zone and the consequent well water level rise. With respect to the regional continuity of the aquifer, a hydraulic interference can be presumed over longer distances, which enables the transfer of effects of larger deformations taking place closer to the earthquake epicenter. Regarding the low values of the theoretical pre-seismic strain derived for the earthquakes of August 10, 2005 and October 25, 2005, and the estimated strain sensitivity of the VS-3 and V-28 wells, we cannot expect the presence of precursory groundwater level changes

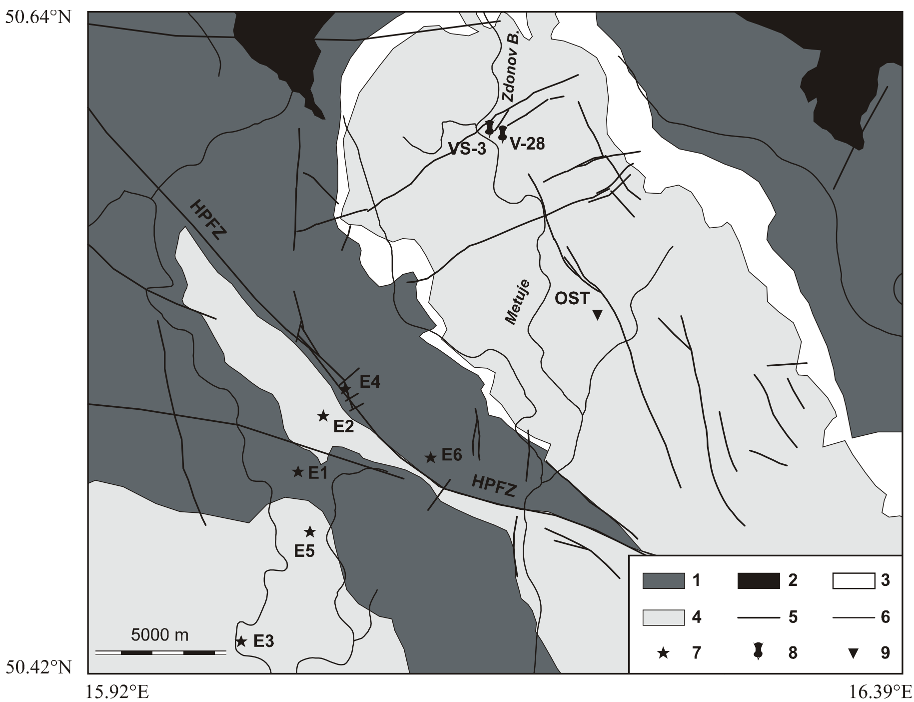

Fig. 7. Seismic events recorded in the area of the HPFZ during the period 1998–2005. In case of a sequence of seismic events, only the main event is displayed (geology and fault tectonics after *Schenk et al., 1989*; *Kněžek, unpublished results*; *Vejlupek, 1990* and *Tásler, 1995*). 1 - Permian and Carboniferous sediments, 2 - Permian volcanics, 3 - Triassic sediments, 4 - Cretaceous sediments, 5 - tectonic faults, 6 - streams, 7 - epicentres of earthquakes listed in Table 6, 8 - observation wells, 9 - seismic station.

unless the observation well is connected to a sensitive zone. Consequently, the V-28 well, which is not opened to the chert series, can be used as a suitable reference object for the identification of anomalous groundwater level changes induced by the weak local seismicity in the area of the Hronov-Poříčí Fault Zone.

Interpretation of the anomalous groundwater level changes observed before the August and October 2005 earthquakes considering the response of the fluid-filled fissure systems of the chert series to the compressive strain is consistent with the general conceptions of the stress conditions in the area of the HPFZ (see Section 2). The compressive strain tendencies perpendicular to the HPFZ are presumed by *Vyskočil (unpublished results)* or *Kontny (2004)*. The compressive strain can also explain the pre-seismic uplifts across the HPFZ, reported by *Vyskočil (1988)*. Relatively rare for the precursory groundwater level changes is the sharp step-like character of the observed anomalies. This type of short-time precursory phenomena recorded within a time span of hours before an earthquake is often explained by aseismic creep-like movements (see, e.g., *Rikitake, 1975*). Similarly *Lorenzetti and Tulis (1989)* anticipated that the pre-seismic strain increases steeply within a few minutes to a month before the earthquake, which is caused by accelerating aseismic slip. Some examples of the step-like precursory groundwater level changes were reported by *Kissin et al. (1996)*. The authors explain these changes by the aseismic movements in the near fault zone as well. In agreement with the above mentioned opinions, we regard the pre-seismic creep movements in the fracture system of the HPFZ as the primary cause of the precursory groundwater level changes recorded in the VS-3 well. Nevertheless, we do not dispose of any direct evidence of the pre-seismic creep movements along the HPFZ. To confirm the above proposed conception of the origin of the hydrologic earthquake precursors it would be necessary to correlate the water level records with the data providing direct information on fault displacement (see, e.g., *Roeloffs et al., 1989*; *Rudnicki et al., 1993*).

8. CONCLUSIONS

Analysis of continuous groundwater level data from two experimental wells situated in the area of weak intraplate seismicity on the NE margin of the Bohemian Massif yielded the following main conclusions:

1. Water level fluctuation in both wells exhibits, in spite of the relatively low instrumental resolution (1 cm), well apparent response to tidal and barometric loading. Both the tidal response and the barometric response reach somewhat higher amplitudes in the V-28 well. Based on the response of the groundwater level to tidal and barometric effects, selected elastic parameters of the observed aquifers were derived: the shear modulus G, the Skempton ratio B, the drained matrix compressibility β and the undrained compressibility β_u. Based on these parameters and assuming the homogeneous poroelastic material, we estimated the sensitivity of both wells to the changes in the crustal volume strain at 0.72 mm/nstr for the VS-3 well and 1.08 mm/nstr for the V-28 well using the method of *Rice and Cleary (1976)*.
2. In connection with the earthquakes of August 10, 2005 (M = 2.4, d = 11.3 km) and October 25, 2005 (M = 3.3, d = 16.8 km), we observed pre-seismic step-like water

level rises in the VS-3 well. The first anomalous precursory change reaching the amplitude of +6 cm was observed 11−15 hours prior to the August 10 seismic event. The second precursory change with the amplitude of +15 cm was recorded 32−29 hours prior to the October 2005 event.

3. Amplitudes of the recorded precursory changes are several times higher than the expected values corresponding to the theoretical precursory crustal strain derived by the method of *Dobrovolsky et al. (1979)*. We therefore presume that the pre-seismic water level steps recorded in the VS-3 well cannot be easily explained assuming the presence of homogeneous poroelastic material. Such an interpretation would not explain the lack of the pre-seismic response in the V-28 well, which has the strain sensitivity approximately 1.5 times higher than the VS-3 well.
4. The origin of precursory events recorded in the VS-3 well is herein explained by its connection with preferential zones of groundwater flow, which are represented by a dense system of highly permeable fissures in the Cenomanian chert aquifer. The pre-seismic water level rise can be then interpreted as a result of compression of these zones, which are, due to their regional continuity, supposed to transfer the effects of larger deformations taking place more closely to the earthquake's epicenter. Pre-seismic creep movements in the fracture system of the HPFZ can be regarded the primary source of the compression causing the precursory groundwater level changes.

The results of the analysis of the relations between groundwater level changes and the local seismicity in the area of the HPFZ show that the anomalous response to the weak earthquakes with magnitude $M < 3.5$ is observable only at sensitive sites, determined by the presence of structurally weak zones. Such an example is the VS-3 well, which is presumed to be connected with fracture systems of the preferential zones of groundwater flow in the chert aquifer. On the other hand, the V-28 well, which is not opened to the chert aquifer and therefore not expected to have any response to weak local earthquakes, can be used as a suitable reference object for the identification of anomalous seismic-induced groundwater level changes in the VS-3 well.

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