



Earthquake-enhanced permeability – evidence from carbon dioxide release following the M_L 3.5 earthquake in West Bohemia



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ABSTRACT

The West Bohemia/Vogtland region is characterized by earthquake swarm activity and degassing of CO_2 of mantle origin. A fast increase of CO_2 flow rate was observed 4 days after a M_L 3.5 earthquake in May 2014 in the Hartoušov mofette, 9 km from the epicentres. During the subsequent 150 days the flow reached sixfold of the original level, and has been slowly decaying until present. Similar behavior was observed during and after the swarm in 2008 pointing to a fault-valve mechanism in long-term. Here, we present the results of simulation of gas flow in a two dimensional model of Earth's crust composed of a sealing layer at the hypocentre depth which is penetrated by the earthquake fault and releases fluid from a relatively low-permeability lower crust. This simple model is capable of explaining the observations, including the short travel time of the flow pulse from 8 km depth to the surface, long-term flow increase and its subsequent slow decay. Our model is consistent with other analyses of the 2014 aftershocks which attributes their anomalous character to exponentially decreasing external fluid force. Our observations and model hence track the fluid pressure pulse from depth where it was responsible for aftershocks triggering to the surface where a significant long-term increase of CO_2 flow started 4 days later.

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1. Introduction

Creation of new springs following earthquakes has been observed for millennia. Thales regarded the Earth as a disk swimming on water that erupts due to earthquake shaking (Oeser, 1992). Many observations of spring discharge anomalies have been collected (Muir-Wood and King, 1993; Montgomery and Manga, 2003; Cappa et al., 2009; Wang and Manga, 2015), which help to improve our understanding of processes related to earthquake preparation and the postseismic response of the crust. Various explanations have been proposed that include changes in crustal static or dynamic strain resulting in migration of the deep pressurized water due to dilation of pores in the crust (Muir-Wood and King, 1993) and release of groundwater from nearby mountains (Rojstaczer and Wolf, 1992; Wang and Manga, 2015). The coseismic discharge of deep fluids is reported by Cappa et al. (2009) in the case of uplift and increased spring water discharge during the 1965–1967 Matsushiro earthquake swarm in Japan. Similarly, the

2014 volcano-tectonic swarm in Lassen, California (Ingebritsen et al., 2015) triggered an increased outflow of hydrothermal fluids already present and equilibrated in a local hydrothermal aquifer. The fault-valve model (Sibson, 1981, 1992) explains situations where active faults transect suprahydrostatic gradients in fluid pressure. Fault rupture, which is associated with transient permeability increase, promotes upward discharge of overpressured crustal fluids along the newly created, high-permeability channel. Subsequent hydrothermal deposition results in resealing of the fault and recovery of the suprahydrostatic regime beneath the ruptured fault portion. Similar behavior is also predicted at hydrostatic conditions when a dilating fault drains water from the surrounding rock (Stanislavsky and Garven, 2003). Fault-valve behavior has been documented based on petrology of ancient faults (Cox, 1995; Sibson and Rowland, 2003) at structural levels corresponding to the base of the seismogenic zone (Sibson, 1990) and by increased post-seismic spring discharge and river flow (Muir-Wood and King, 1993). Aftershock activity has been successfully explained by fluid diffusion models for the 1997 Umbria–Marche sequence (Miller et al., 2004), the 2009 L'Aquila earthquake (Chiarabba et al., 2015), and the 2008 Wenchuan M_S 8.0 (Liu et al., 2014) and 2011 Tohoku-oki M_W 9.0 (Terakawa et al., 2013) earthquakes. To our

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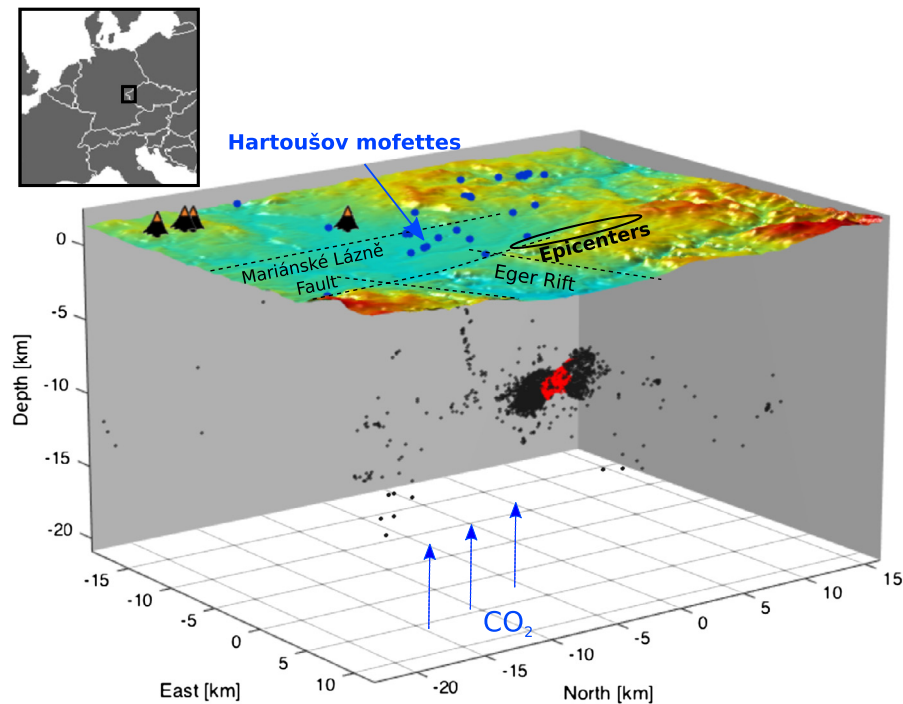


Fig. 1. Perspective 3D view of the earthquake swarm region of West Bohemia/Vogtland. Topography ranges from 400 to 800 m. CO₂ mofettes and springs are indicated by blue circles, blue arrows indicate presumed CO₂ migration from depth, black dots show earthquake hypocenters between 1991 and April 2014 and red dots earthquakes that occurred during the 2014 sequence (May–August). Cones indicate Quaternary volcanism. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

knowledge, there have been no observations of a coseismic and long term increase of discharge of mantle-derived fluids, in particular CO₂, and simultaneous seismic activity with spatiotemporal characteristics consistent with fluid triggering models.

On 24 and 31 May 2014 two (M_L 3.5 and 4.4) earthquakes occurred in West Bohemia (Central Europe) on a subvertical fault at depths of 8.5 and 8 km, in the same region and at similar depths as earthquake swarms within the last 30 years (Fischer et al., 2014). Both earthquakes were followed by aftershock series with unusually high activity rates and clear migration patterns. Based on statistical modeling of the sequences, Hainzl et al. (2016) explained the anomalous character of the aftershock series by an exponential aseismic forcing, which points to fluid intrusion due to mainshock rupture as the most probable triggering mechanism. In addition to persistent seismic swarm activity, the area of interest (Fig. 1) is typified by Quaternary volcanism active more than 300 ka ago (Mrlina et al., 2009) and strong discharge of CO₂ of upper mantle origin in gaseous mineral springs and dry mofettes (Kämpf et al., 2013) whose emission rates and isotopic signatures of He and C show some correlations with seismic activity (Heinicke and Koch, 2000; Bräuer et al., 2014).

Since 2009 we have been monitoring CO₂ emission in a hydrologic borehole at the Hartoušov mofettes field (Nickschick et al., 2015) (Fig. 1). The original gas flow rate had decreased by a factor of 5 in spring 2014, but recovered rapidly to the original level starting four days after the first M_L 3.5 mainshock. This response coincides with groundwater level (GWL) variations measured in the same well, which scale with the bubble content and act as an independent verification of the anomaly. Here we analyze the measured data series and suggest the fault-valve mechanism as the most plausible model to explain the anomalies. We show by numerical modeling that an emission rate increase only four days after the earthquake requires high diffusivity in the upper part of the fault.

2. Data

Borehole VP8303 is situated in the mofette field of Hartoušov where about 60 tons of CO₂ are released daily over an area of 0.35 km² (Nickschick et al., 2015). The diameter of the plastic casing is 125 mm. The borehole penetrates Quaternary and Tertiary deposits of the Cheb Basin to a depth of 28.2 m, and its perforated part targets an artesian aquifer sealed by a clay layer at 7.6 m depth. Groundwater is saturated with carbon dioxide of magmatic origin (Bräuer et al., 2014) and exsolves at a depth of about 20 m creating ascending CO₂ bubbles. The massive gas flow results in strong oscillations of GWL, and, consequently, in eruptive gas flow during periods of higher CO₂ flow rate. The gas flow is vented through the closed wellhead and continuously measured since November 2009 by a Ritter gas flowmeter, along with gas temperature, atmospheric pressure and other parameters, at sampling interval of 5 min. Freezing and other environmental effects resulted in some gaps in data, and the flowmeter was also out of operation in the period from January 2012 to November 2013.

The daily average gas flow rate (Fig. 2) shows gradual decrease from 3.6 kg/h in late 2009 to 0.7 kg/h in early 2014. After five months of stabilized low flow, an abrupt rise of CO₂ flow commenced on 28 May 2014, four days after the first M_L 3.5 mainshock. The steady, almost linear increase lasted for about two months, reaching a level of almost 4.8 kg/h, and was followed by a slower decrease of flow rate down to about 2.4 kg/h in February 2016 that resembles a similar decrease from 2009 to 2011, after the 2008 earthquake swarm. Along with the flow rate, which suffers from data outages, we show the concentration of gas bubbles in water, which was derived from GWL measurements (see Methods). The flow rate strongly correlates with bubble concentration. Note the parallel decrease of both quantities in 2010 and 2011, their steady level in early 2014, and the fast increase that followed the onset of 2014 earthquakes. After a period of unreliable flowmeter operation in the second half of 2015, both curves show self-similar decay. Accordingly, below we use the term “flow

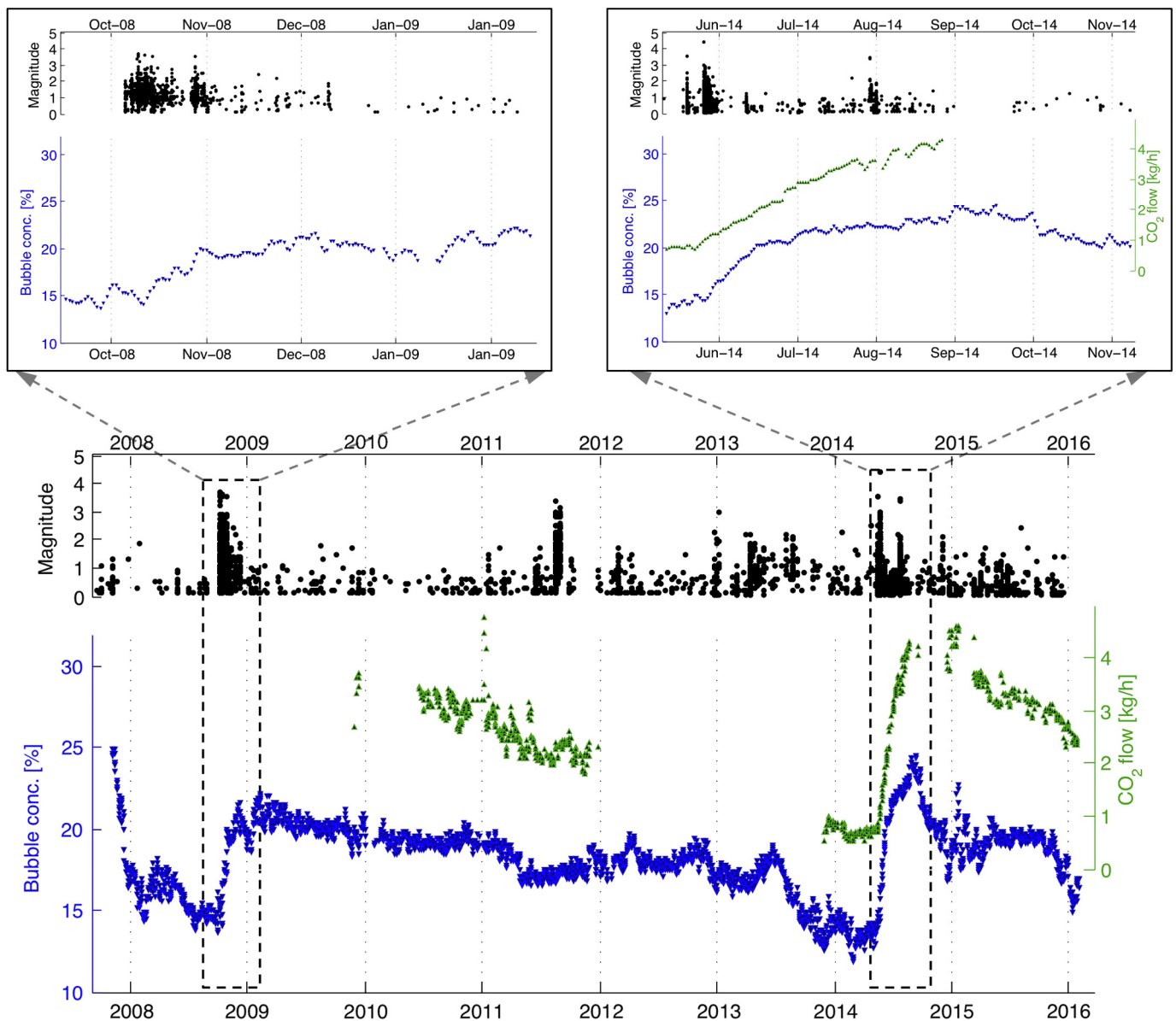


Fig. 2. Local seismic activity (event magnitudes indicated by black dots) and CO₂ degassing recorded in borehole VP8303 in the Hartoušov mofette field. CO₂ flow rate (green up-pointing triangles) and concentration of CO₂ bubbles in water (blue down-pointing triangles) are shown for the period from November 2007 to February 2016. The large scatter of bubble concentration curve is caused by the oscillating ground water level in the well. The periods gas flow rise related to seismic activity are zoomed in the two insets.

rate” to denote both CO₂ flow rate and CO₂ bubble concentration in the well.

A similar anomalous increase of CO₂ bubble concentration occurred in 2008, during an earthquake swarm that started on October 6. The left inset in Fig. 2 shows that the rise started on October 10, contemporaneous with the strongest M_L 3.7 event and 4 days after the inception of the swarm. Hence we find repeated behavior of the same kind: the degassing of CO₂ at Hartoušov mofette reacts to major earthquake sequences in the nearby (9 km distant) Nový Kostel focal zone by (A) fast postseismic increase of gas flow with a delay of about 4 days after the onset of the sequence. The flow rate increase lasts for more than two months and is followed by (B) a more gradual flow decrease. During the investigated period three major seismic sequences with magnitudes of largest events exceeding M_L 3.5 took place. Two of these (2008 and 2014) were followed by flow anomalies; the 2011 swarm was not visibly followed by similar CO₂ flow anomaly.

3. Model

Various models can be invoked to fit the observed postseismic anomalies of CO₂ flow. Here we consider three of them. The first is a model of gas bubble nucleation and growth in groundwater (Crews and Cooper, 2014) initiated by passage of seismic waves. This model was proposed to explain a 34 cm increase of GWL that occurred less than 15 min after the Landers 1992 earthquake in the Bourdieu Valley well, 435 km from the epicenter. The water level reached a peak after ~36 h and then slowly declined; no degassing was reported. This scenario differs significantly from our observations, namely gas flow data are absent, the epicentral distance is large (>400 km vs. 10 km) and time scale is shorter (1.5 vs. 80 days of rise time).

Another possible model (Bräuer et al., 2003) consists of an impermeable cap rock that overlies the focal zone in Nový Kostel (Fig. 1) and traps ascending mantle fluids, preventing them from rising further. This model requires that the cap rock, with a max-

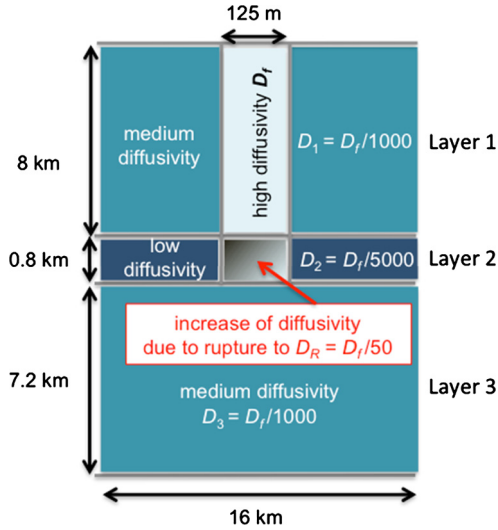


Fig. 3. Conceptual three-layer model. The uppermost Layer 1 contains a very thin high-diffusivity channel (its thickness in the figure is exaggerated). The middle Layer 2 represents a seal that can be ruptured by seismic activity below the channel. The lowest Layer 3 is thick enough that we may suppose constant pressure p_b at its bottom. Diffusivity ratios of the optimum model are indicated.

imum expected depth limited by the smallest hypocenter depth of about 5 km, is broken by the passage of seismic waves and subsequently releases accumulated fluids. This scenario appears more plausible in the case of shocks of higher magnitudes like the M_L 3.5 mainshock at the beginning of the 2014 sequence. Its applicability to the weak M_L 2 events that started the 2008 swarm (the first M_L 3+ events occurred only four days later) is however questionable. It is well accepted that oscillations in stress can increase permeability of geologic environment (Manga et al., 2012). According to Elkhoury et al. (2006) rock permeability increases during earthquakes depending on the seismic-wave peak ground velocity (PGV) in the range of 0.21–2.1 cm/s. The lower boundary of this interval is however well above ground velocities observed during West Bohemia earthquakes; while PGV of M_L 1.0 events range up to 10 $\mu\text{m/s}$ it exceeds 1 mm/s only for the strongest $M_L > 3$ events at the closest stations (Fischer, 2005).

A third possible scenario is the fault-valve model (Sibson, 1981), which is supported by the fact that we observe two full cycles: (i) low flow rate in 2007–2008, consistent with self-sealing of the fault; (ii) a rapid flow rate increase after 2008 swarm, consistent with fault rupturing; (iii) continuous decay of the flow rate from 2009 to 2014, consistent with self-sealing of the fault; (iv) fast flow rate increase in 2014 consistent with fault rupturing; and (v) flow rate decay in late 2014 consistent with repeated self-sealing of the fault. The fluid-valve mechanism plays an active role in the seismogenic process and results in the observed periodicity of seismic activity. It does not require large earthquakes that would break a distant cap rock, because even small earthquakes may increase fault permeability locally and open a pathway for the gas to reach the surface (Wang and Manga, 2010). Our choice of the fault-valve model is also consistent with the recent study of Hainzl et al. (2016) who analyzed in detail the 2014 seismic activity and attributed the anomalous character of the aftershock series to an exponential aseismic forcing, which points to fluid intrusion following mainshock rupture as the most probable triggering mechanism.

Our observations show four characteristic parameters: (i) delay time of about 4 days between the onset of seismic activity and start of the gas flow growth, (ii) increasing flow for about 150 days, (iii) a ratio of about 6 between the flow maximum and its stationary level prior to the earthquake and (iv) characteristic decrease

of the flow after reaching a maximum. We use numerical simulation of two-dimensional fluid diffusion to test whether the fluid pressure pulse and changes of flow rate are consistent with Sibson's (1992) fault-valve model composed of a simple three-layer crustal structure (Fig. 3). The uppermost Layer 1 contains a high-diffusivity vertical channel corresponding to highly permeable fault zone surrounded by less diffusive ambient crust material.

An intermediate, thin Layer 2 represents a low-diffusivity seal for the gas flow providing basic stationary flow in the model. The fault crosses this layer, whose diffusivity is periodically increased by fracturing during earthquake swarms. The lowermost Layer 3 is a CO_2 reservoir characterized by a relatively low diffusivity equal to that of upper crust. We assume constant pressure at the bottom of the model.

Our scenario starts with a mainshock that breaks the seal layer, increasing the permeability of the seal and allowing flow of over pressurized fluids into the overlying fault zone. The fluids trigger aftershocks (Hainzl et al., 2016) and migrate through the open fracture zones to the surface, where they are observed four days later in the form of a rapid increase of CO_2 discharge in the mofettes.

4. Methods

4.1. Conversion of GWL to bubble concentration

To derive the bubble concentration we used GWL measurements recorded by pressure gauges compensated to atmospheric pressure. The groundwater level in a borehole tapping an aquifer depends on the water density in the borehole. Because pervading gas bubbles decrease water density, a GWL increase compensates for the hydrostatic pressure in the aquifer. Provided that other hydrologically driven GWL changes are negligible, the average density $\rho(t)$ of the water–gas mixture can be determined using the GWL measurements (Fig. 4a) as

$$\rho(t) = (p_e - p_m(t)) / g(d_e - d_m), \quad (1)$$

where p_m is measured at depth d_m , and d_e is the depth and p_e the pressure of bubble formation, which depends on the temperature, salinity and surface tension of the liquid (e.g. Elhajj et al., 2014). These were determined by simultaneous water conductivity and pressure logs, employing the fact that the conductivity log oscillates when bubbles are present and stationary elsewhere. We measured $p_e = 145$ kPa and $d_e = 20.5$ m at the point where the conductivity log changed from constant to oscillating. The exsolution depth varies with time and was further corrected for hydrological GWL variations using a nearby well which is free of CO_2 degassing (Fig. 4b). Water temperature was quite stable, showing seasonal variations from 6 to 10 °C.

The average density of water–gas mixture is then converted using density of water ρ_0 to average volumetric fraction (concentration) C of bubbles as

$$C(t) = (\rho_0 - \rho(t)) / \rho_0. \quad (2)$$

GWL changes due to hydrologic effects directly influence the exsolution depth d_e . We used the GWL record from well VP8302 in Hrzín (Fig. 4b), which is situated 8 km upstream in the same river catchment, to correct for hydrologic effects. This reference record is free of preseismic decrease and postseismic rise of GWL in 2008 and 2014. Due to its upstream position, we expect that the long-term GWL decrease in the reference well was larger than that in the monitored well; we used a weight of 0.5 to correct the GWL at Hartoušov using GWL at Hrzín.

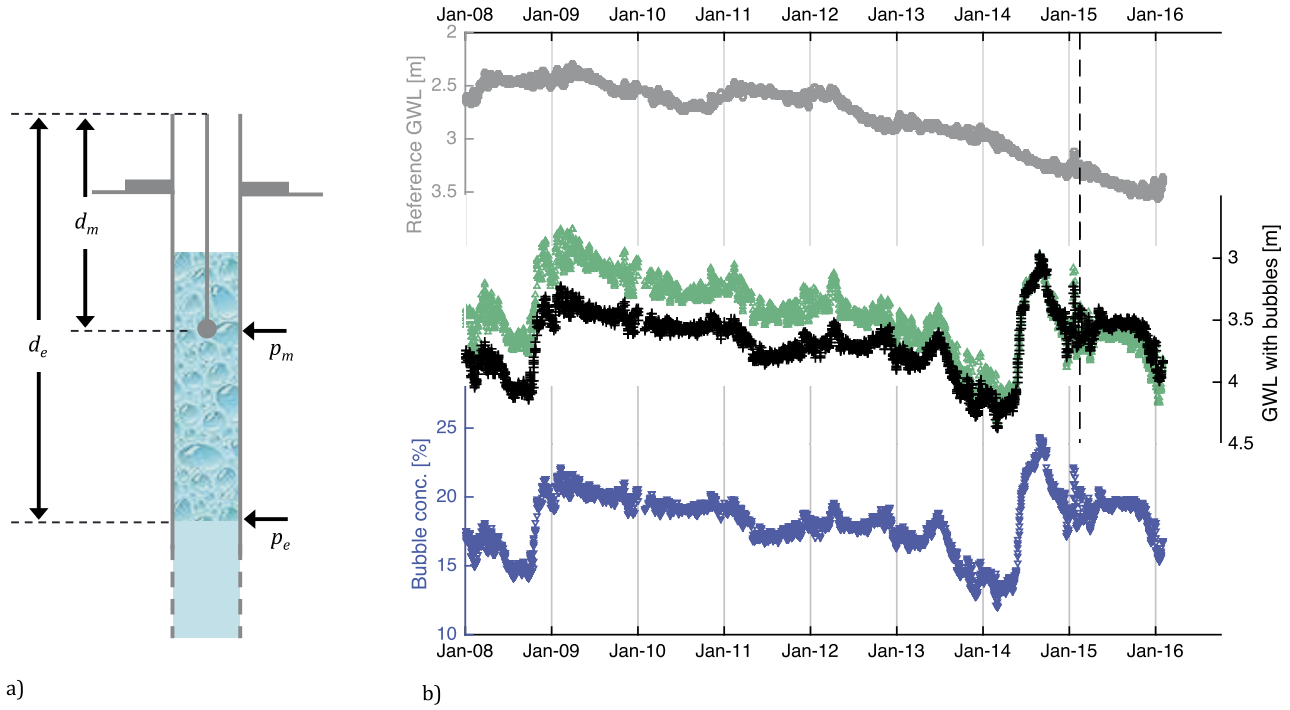


Fig. 4. (a) GWL in a borehole with degassing of CO₂. (b) Depth of GWL in the reference well (grey, first curve from top) and in the monitored Hartoušov well before (green, second from top) and after (black, third from top) correction for hydrologic GWL changes; concentration of bubbles (blue, lower curve) was determined by Eq. (2). The vertical dashed line shows the reference date of 26-Feb-2015 used for correcting the GWL in Hartoušov using the GWL in Hrzín as reference (gray) and for determining the depth d_e and pressure p_e of bubble evasion in Hartoušov.

4.2. CO₂ flow modeling

We simulate a propagation of a pressure pulse from the seismogenic depth to the surface by a simple diffusion equation. Based on the CO₂ degassing in mofettes we suppose that the fluid is probably a mixture of two phases (water and carbon dioxide), dominantly CO₂, which changes phase during its ascent to the surface from supercritical fluid to liquid (at about 700 m depth) and finally to gas (at about 20 m depth). Because properties of supercritical CO₂ are similar to water (Colina et al., 2003) we model the flow, similarly to Miller et al. (2004) and apply a single-phase fluid model. Lacking knowledge of temperatures, viscosity and rock porosity, we use diffusivity, rather than permeability. Our 2-D Cartesian model consists of several homogeneous blocks as drawn in Fig. 3. Then the time evolution of the fluid overpressure p is described by (e.g. Chen et al., 2006)

$$\partial p / \partial t = \text{div}(D \text{grad}(p)). \quad (3)$$

On the surface (top of the model) $p = 0$ and we consider constant pressure magnitude p_b at the model bottom. No horizontal flow is allowed at the sidewalls, which corresponds to the assumption that horizontal flow is negligible at larger distances from the channel. A finite difference scheme is applied to solve equation (3) with these boundary conditions for the layered structure shown in Fig. 3. The flow simulation is started from a steady-state calculated using a uniform low value of diffusivity throughout the seal (Layer 2). Then, at time $t = 0$ the seal-layer diffusivity below the 125-m-wide channel is increased and the time-dependent solution to equation (3) is calculated. We tested several models to obtain a good fit to the time delay of the initial increase of the surface flow after the mainshock, the time interval of the flow increase, the ratio of the maximal to minimal flow and the characteristic decrease of the flow after reaching its maximum. To avoid numerical problems with resolution, we consider here a rather wide chan-

nel (125 m; 0.8% of the model height). Numerical tests show that characteristic times of channel flow are not very sensitive to the channel width due to the very high diffusivity contrast between the channel and the ambient crust.

The diffusivities of the optimal model are $D_1 = D_3 = D_f/1000$ (Layers 1 and 3 in Fig. 3), $D_2 = D_f/5000$ and $D_R = D_f/50$ (Layer 2 – seal before and after earthquake) where D_f is the diffusivity of the channel, which is considered to be a free scaling parameter of the model. The total flow q of the fluid is given by Darcy's law $q = -(k/\mu) \text{grad}(p)$, where k is the permeability of the medium and μ is the fluid dynamic viscosity. Note that $D = k/(\mu\Phi c)$, where Φ is the porosity of the medium and c the total CO₂ compressibility. The flow amplitude is proportional to p_b , which is considered as another free scaling parameter. Accordingly, we solve eqn. (3) for unit depth of the model, unit pressure at the bottom and unit diffusivity of the channel and find two scaling constants a and b such that $a(\partial p / \partial z)$ at the surface yields the best fit with the observed CO₂ flow data after scaling time by the constant b .

Note that in the borehole we measure total mass of CO₂ per unit time, i.e., our data correspond to the product $S\rho q = -S\rho(k_c/\mu) \text{grad}(p)$, where S is the borehole cross-section, ρ is the CO₂ density at atmospheric pressure and k_c is the channel permeability. Taking into account that $\text{grad}(p) = (p_b/L) \text{grad}^*(p^*)$, where stars denote dimensionless quantities, we obtain the interpretation of the constant a as $a = S\rho k_c p_b / (\mu L)$. Similarly, it is clear from eqn. (3) that the time is scaled by the factor $b = L^2/D_f$, i.e., $t = L^2/D_f t^*$. Our optimal fit of data yields $a = 108 \text{ kg/h}$ and $b = 245 \text{ days}$. The depth of our model is $L = 16 \text{ km}$ and the diffusivities are thus scaled by $L^2/b = 12 \text{ m}^2/\text{s}$.

The obtained diffusivity lies in at the upper margin of the range of fault-zone diffusivities. For comparison, by analyzing more than 90 case histories of induced seismicity, Talwani et al. (2007) determined hydraulic diffusivity of fractures associated with seismicity to lie between 0.1 and 10 m²/s.

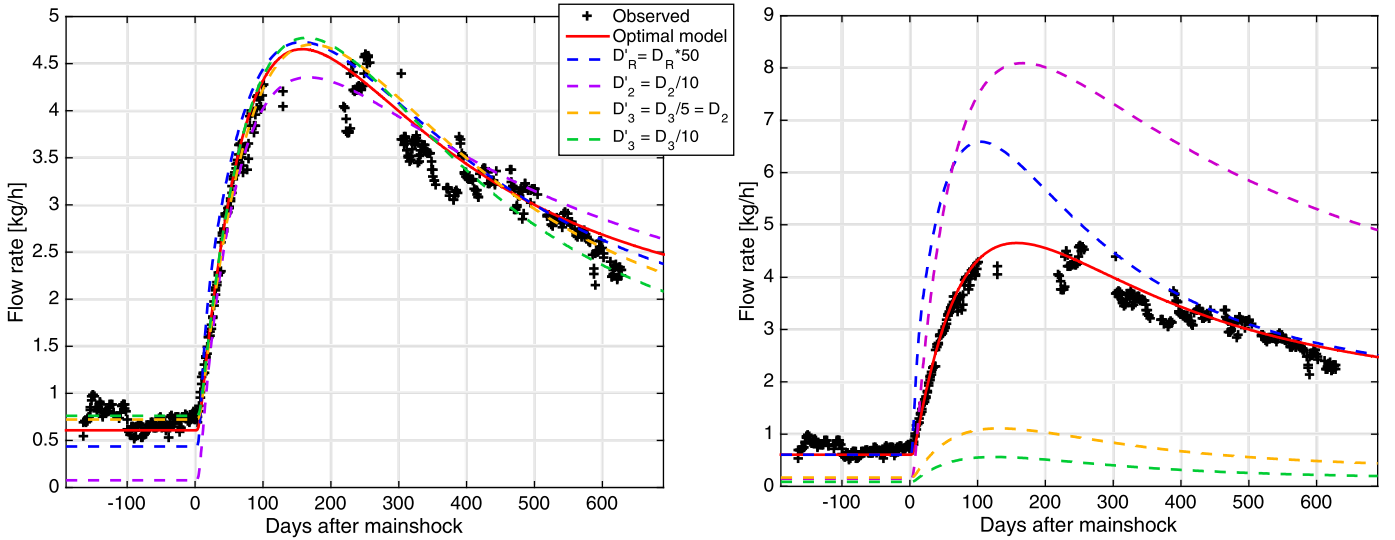


Fig. 5. Comparison of measured (black dots) and modeled CO₂ flow rate (lines) at the central surface point of the high-permeability channel for models with different diffusivities (Fig. 3). The dimensional flow and time was obtained by rescaling the dimensionless calculated flow by factors a and b , see Methods. The red line shows the result for an optimal model (see Fig. 3 with $D_f = 12 \text{ m}^2/\text{s}$; factors $a = 108 \text{ kg/h}$ and $b = 245 \text{ days}$). The dashed lines refer to different diffusivities of the rupture zone, Layer 2 and Layer 3: rupture zone diffusivity D_R increased 50 $\times$ that corresponds to elongation of the high-permeability channel; Layer 2 diffusivity D_2 decreased 10 $\times$; Layer 3 diffusivity D_3 decreased 5 $\times$ and 10 $\times$ (diffusivity of this layer is then the same or even lower than that of Layer 2 and represents models without sealing) compared to the optimum model. The left panel is obtained by scaling each model to yield its best fit with data after rupture; the right panel shows the case when the same scaling parameters a and b are used for all models. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

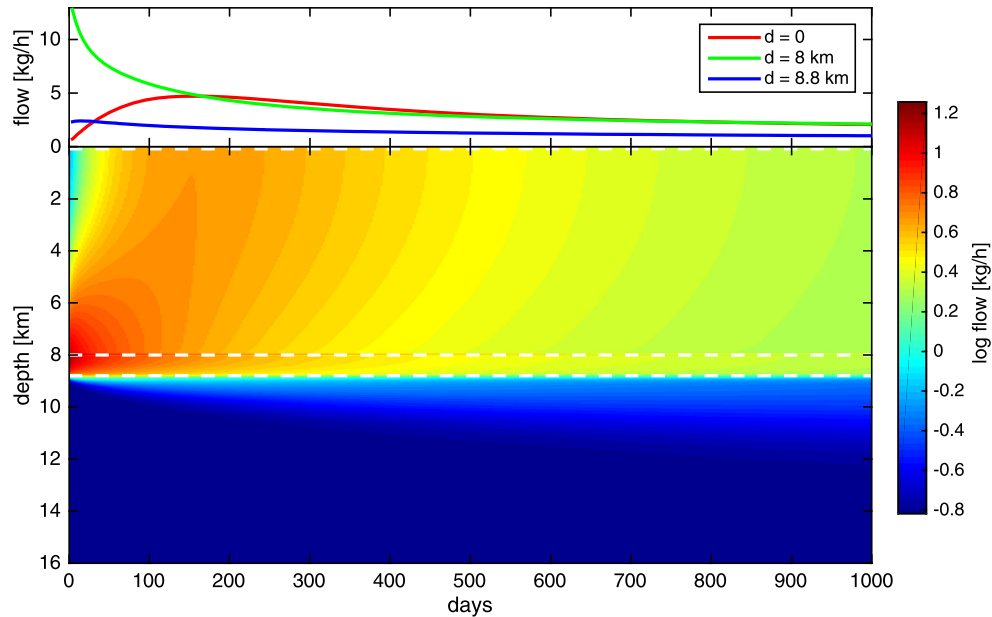


Fig. 6. Flow rate in the optimal model after the earthquake. Bottom color image shows time and depth dependence of the flow rate along the centerline of the model in logarithmic color scale. The top diagram shows flow rate at the Earth's surface (red curve), at the top (green curve) and bottom (blue curve) of the seal layer; the depths ($d = 0, 8$ and 8.8 km) of the curves are indicated by dashed white line. For more details see caption of Fig. 5. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

5. Results and discussion

The result for an optimal model with $D_f = 12 \text{ m}^2/\text{s}$ is shown in Fig. 5 (red solid line). Since flow is linearly dependent on the magnitude of pressure p_b at the bottom of the model, the flow maximum can be matched to measured data by rescaling p_b . However, the ratio of flow maximum to initial value and other flow changes are independent of this boundary condition. The sensitivity of flow characteristics to changes of diffusivities is documented by the dashed lines. A higher diffusivity of the ruptured part of the seal (blue line) and lower diffusivity of Layer 2 (violet line) yields

steeper growth of CO₂ flow rate and a higher maximum/minimum flow ratio. Models with low diffusivity in the bottom Layer 3 (orange and green lines), such that the sealing effect of the middle layer is removed (i.e. initial pressure gradient in the Layer 3 is the same or higher than the initial pressure gradient in the Layer 2) show faster flow decay, but can be rescaled to yield satisfactory fit to the data. Note that we also tested changes in diffusivity of the uppermost crust (Layer 1) surrounding the channel, but the results exhibited very small sensitivity. The optimum seal-layer model (red line) and non-seal model (orange line) appear almost equivalent, which indicates that the robust component of

the model is a permeable channel through the crust and breakage of a low-permeability zone.

At early times the flow rate field along the centerline of the model (Fig. 6 bottom) shows a steep depth gradient at the bottom of the broken seal due to fluid release in Layer 3. Maximum upward flow occurs in the lower part of the channel. The time dependence of the flow (Fig. 6 top) reflects gradual release of the fluid, with transient onset and a delayed flow rise at the Earth's surface (red curve). The flow at the top of the ruptured seal layer (green curve) is higher than the flow at its base (blue curve) because of additional horizontal inflow from the seal.

Propagation of the pore pressure increase in the fault zone may be responsible for earthquake triggering. Analysis of the aftershock sequence of the strongest (M_L 4.4) mainshock of May-31 (Hainzl et al., 2016) has shown that the unusually high rate of aftershocks, their slow decay and migration from a point can be explained by exponentially decaying external forcing that started at the time of the mainshock. Accordingly, those authors suggested that the mainshock opened fluid pathways from a finite fluid source and that fluid flow contributed to triggering of aftershocks. Our data complement this coseismic fluid pulse activity in the focal zone by describing postseismic CO_2 flow increase in a mofette 8 km above and 9 km to the south of the hypocenters and suggest that these two events probably share the same origin – mainshock rupture. The CO_2 flow increase at the mofette was observed 4 days after the first mainshock of May-24 whereas the statistical modeling of Hainzl et al. (2016) focused on the second mainshock of May-31, which probably also contributed to the surface CO_2 flow increase. We tested a numerical model with two ruptures 7-days apart and found that the difference between one and two ruptures is not significant in terms of the error of the flow rate measurements, hence we favor the simpler model with a single rupture.

The excess flow rate after valve opening amounts to an average of about 2.9 kg/h giving the total of 44 t for the modeled period of 630 days. By comparing the average CO_2 flow in the monitored well of about 2.8 kg/h and the estimate of the daily CO_2 production in the Hartoušov mofette field of 60 t (Nickschick et al., 2015) one gets that the monitored well accounts for about one thousandth of the total CO_2 flow in the mofette field. Hence, the excess CO_2 mass released after the mainshock in the whole mofette field is about 44 000 t, which amounts to the CO_2 fraction of 2.4×10^{-8} in the bulk of the basement reservoir of the size $7.2 \times 16 \times 16$ km, which was released after valve opening. This appears as a lower estimate only regarding the fact that the CO_2 increase most probably occurred also in other neighboring mofette fields.

Our model does not invoke the sealing mechanism that is responsible for pressure build-up in Sibson's (1992) model as simulated by Zhang et al. (2015). To explain the observed decay of flow rate, the process of emptying a finite fluid reservoir is sufficient, and fault sealing is not necessary. Pressure build-up due to decreased fault diffusivity is, however, a likely mechanism for the repeated seismic activity observed in 2008 and 2014.

6. Conclusions

The CO_2 flow rate monitoring in a borehole within the Hartoušov mofettes in West Bohemia has shown a long-term decrease with an overall decay from nearly 3.6 kg/h in late 2009 to 0.7 kg/h in spring 2014. There was then a rapid increase in flow rate starting 4 days after the first event of a 2014 seismic sequence, and the flow rate exceeded the original level of 3.6 kg/h within 4 months. A similar anomaly occurred during the 2008 swarm sequence, which points to repeated fault-valve behavior. We use numerical simulation of gas flow in a two dimensional model to demonstrate that the delayed flow rate increase can be explained by breakage

of a sealing layer located at the hypocentral depth, which releases fluid from a relatively low permeability lower crust. Fluid involvement in seismic triggering is supported by an independent analysis of the 2014 aftershocks that attributes its anomalous character to exponentially decreasing external fluid forcing.

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