In situ stress variations at the Variscan deformation front — Results from the deep Aachen geothermal well

Ute Trautwein-Bruns a,⁎, Katja C. Schulze b, Stephan Becker a, Peter A. Kukla a, Janos L. Urai c

a Geological Institute, RWTH Aachen University, Wüllnerstraße 2, D-52056 Aachen, Germany
b GeoMechanics International, Inc., Emmerich-Josef Straße 5, D-55116 Mainz, Germany
c Structural Geology, Tectonics and Geomechanics, RWTH Aachen University, Lochnerstraße 4-20, D-52056 Aachen, Germany

A B S T R A C T

In 2004 the 2544 m deep RWTH-1 well was drilled in the city centre of Aachen to supply geothermal heat for the heating and cooling of the new student service centre “SuperC” of RWTH Aachen University. Aachen is located in a complex geologic and tectonic position at the northern margin of the Variscan deformation front at the borders between the Brabant Massif, the Hohes Venn/Eifel areas and the presently active rift zone of the Lower Rhine Embayment, where existing data on in situ stress show complex changes over short distances. The borehole offers a unique opportunity to study varying stress regimes in this area of complex geodynamic evolution.

This study of the in situ stresses is based on the observation of compressive borehole breakouts and drilling-induced tensile fractures in electrical and acoustic image logs. The borehole failure analysis shows that the maximum horizontal stress trends SE–NW which is in accordance with the general West European stress trend. Stress magnitudes modelled in accordance to the Mohr–Coulomb Theory of Sliding Friction indicate minimum and maximum horizontal stress gradients of 0.019 MPa/m and 0.038 MPa/m, respectively. The occurrence of drilling-induced tensile failure and the calculated in situ stress magnitudes are consistent with a model of strike-slip deformation.

The observed strike-slip faulting regime supports the extension of the Brabant Shear Zone proposed by Ahorner (1975) into the Aachen city area, where it joins the major normal faulting set of the Roer Valley Graben zone. This intersection of the inherited Variscan deformation grain and the Cenozoic deformation resulting in recent strike-slip and normal faulting activity proves the tectonically different deformation responses over a short distance between the long-lived Brabant Massif and the Cenozoic Rhine Rift System.

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1. Introduction

Understanding the stresses in the earth's crust is of great interest as the stress regime influences underground mining, fluid movements along joints and faults but also the failure of engineering structures and earthquakes. The first- and second order tectonic stress pattern in the crust are well known based on data in the World Stress Map database (Zoback, 1992; Reinecker et al., 2005; Heidbach et al., 2008), but for the purpose of planning drilling and underground mining projects (e.g. geothermal wells and underground storage) the understanding of third order patterns and local stress field variations is important.

Due to the tectonic position in the triangle between the Caledonian London Brabant Massif, the Variscan Ardennes/Eifel Mountains and the active rift zone of the Lower Rhine Embayment, the area of Aachen is in an interesting position for structural and in situ stress investigations (Fig. 1).

1.1. Geodynamic evolution

The present-day structure in the region of Aachen is formed by a complex geological history that extends back to Early Paleozoic times when the continental Gondwana-derived Avalonia terrane, including the London–Brabant Massif (Jager, 2007), had collided with Baltica and Laurentia within the Caledonian orogeny. This collision generated major sutures and faults, which have a NW–SE trend in the Netherlands (Dirkzwager et al., 2000).

During Devonian and Carboniferous times sedimentary rocks were deposited south of the London–Brabant Massif and were deformed during the Variscan Orogeny. NE–SW to ENE–WSW (in Belgium) trending folds and thrust faults mark the front of the Variscan Mountain Belt. The Aachen–Midi Thrust is the northernmost Variscan thrust fault that outcrops only a few hundred meters south of the Aachen borehole.
The so-called Bordiére Fault first described by Legrand (1968) is interpreted to follow the northern boundary of the Namur Synclinorium (Leynaud et al., 2000). This fault and the Aachen Midi Thrust are grossly in line with the North Artois dextral shear zone in the north of France, which is presumably connected to the Hainaut Shear Zone in Belgium (Leynaud et al., 2000). Ahorner (1975) termed this major seismoactive fracture zone of Central Europe the Belgian or Brabant Shear Zone, which reaches westward across Belgium to the Channel coast near Ostend.

The Lower Rhine Embayment is part of the recently active European Cenozoic Rift System of western and central Europe. It consists of several NW–SE trending graben structures, with the Roer Valley Graben as its SW-extension. The Roer Valley Graben is bordered by major normal faults (e.g. Peel Boundary Fault Zone and Feldbiss Fault Zone). Related to this rift system, the Aachen borehole is located on the SW flank of the Roer Valley Graben at its southernmost extension and is locally framed by the Laurensberger Fault system.

Bassin formation of the Lower Rhine Embayment started in the Late Paleozoic (Ziegler, 1992; Ziegler and Dézes, 2007) as foreland basin of the Variscan mountain chain (Ziegler, 1990). During the Mesozoic, the area was characterized by several periods of subsidence and inversion, reactivating Variscan structural trends (Michon et al., 2003; Worum et al., 2004). The opening of the European Cenozoic Rift System started during the Late Eocene and evolved under repeatedly changing stress fields (Michon et al., 2003), reflecting changes in the interaction of the Pyrenean and Alpine orogens with their forelands (Dézes et al., 2004).

Major fault structures occur in the vicinity of the drill site. The NE–SW trending Aachen–Faîlle Du Midi Thrust Fault, the northernmost Variscan overthrust, intersects the surface 500 m to the SW. Mesozoic and Cenozoic NW–SE trending faults, e.g. the Laurensberger Fault system, are connected to the active rift zone of the Lower Rhine Embayment and frame the borehole location to the SW and NE (Fig. 1).

1.2. Recently active stress field and natural seismicity

Recent tectonic movements are expressed by the seismic activity of the region. Natural seismicity is mainly concentrated in the border region between Belgium, the Netherlands, and Germany (Camelbeeck et al., 2007). After Leynaud et al. (2000) the region close to Aachen can be divided into four different seismic source zones: the zones of Liège, Haute–Fagnes and Limburg as well as the Roer Valley Graben. Camelbeeck and Eck (1994) also emphasize the complex tectonic deformation field in the small region extending from the Roer Graben to the city of Liège (40 km SW of Aachen).

The World Stress Map database information for the study area is based on earthquake focal mechanisms, with the maximum horizontal stress in SE–NW orientation (145° ± 25°, Pleneuf and Bonjer, 1997), and its magnitude decreasing with the distance from the alpine deformation front (Larroque et al., 1987). The Roer Valley Graben zone is the most active area of intraplate Northwest Europe (Camelbeeck et al., 2007). Under the present-day NW-directed compressional stress field extension occurs in NE–SW direction by normal faulting with a small strike-slip component (Dost and Haak, 2007). The two largest earthquakes within this seismotectonic zone are those of Roermond (M=5.4 on 13 April 1992) and Alsdorf (M=4.8 on 22 July 2002) (Camelbeeck et al., 2007). The hypocentral depths of the earthquakes in the graben range from 5 to 20 km, and indicate that the boundary faults continue into the lower crust (Dost and Haak, 2007).
In the eastern part of the Brabant Massif, three recent earthquakes for which reliable focal mechanisms have been calculated show normal dip-slip dislocations like in the Roer Valley Graben but with a N–S direction of extension suggesting a minimum horizontal stress rotation of 50° compared to the minimum horizontal stress direction of the Roer Valley Graben (Camelbeek and Eck, 1994).

West of the Roer Valley Graben (in the Limburg Zone) an anomalously shallow, swarm-like seismicity occurred at depths between 3 and 8 km in 1985 and 2000–2002, most probably connected to the E–W oriented Kunrade Fault, which shows a scarp at the surface south of Voerendaal (20 km NW of Aachen, Dost and Haak, 2007). For the Voerendaal events, different source solutions have been proposed by Houtgast (1991), who reports a normal fault of 270° strike and 80° N dip, and Camelbeek (1994), who reports strike-slip movement (Dost and Haak, 2007).

West of Aachen, the Liège zone includes the region between the cities of Liège in Belgium and Gulpen in the Netherlands, limited by the Nordière Fault to the North and the Aachen Midi Thrust to the South and is characterized by strike-slip earthquakes with thrust component, one of which is the 1983 Liège event (Leynaud et al., 2000). The Earthquake of Liège is characterized by strike-slip with a small thrust component along a NNE–SSW striking fault (Camelbeek and Eck, 1994).

Ahorner (1975) defined the whole region from the city of Cologne westward across Aachen to the North Sea (including the Liège area, the North Artois shear zone and the Hainaut Shear Zone) as Belgian or Brabant shear zone with right-lateral strike-slip deformation along NNW–ENE-trending crustal faults.

Southwest of the Roer Valley Graben notable seismic activity exists in the Belgian Ardennes and Eifel Mountains (Camelbeek et al., 2007). Several recent earthquakes with magnitudes up to 4.4 have been recorded within the Haute-Fagnes zone. The mechanisms range from pure normal faulting on NW–SE-striking faults to quasi-pure strike-slip faulting on NNW–SSE or ENE–WSW faults (Camelbeek et al., 2007). One of the most violent historical earthquakes having ever struck NW Europe was the 1692 Verviers earthquake (30 km SW of Aachen, Demoulin et al., 2005). It was related to the NNW–SSE-striking Hockai fault zone, already defined by Ahorner (1983). A seismic sequence along this fault zone in 1989–1990 suggests that activity is related to sinistral strike-slip mechanisms (Lecocq et al., 2008).

Present day uplift rates up to 1.6 mm/y have been observed in the southwestern Eifel Mountains and the Hohes Venn (Ahorner, 1983). Several authors have suggested that pre-existing basement faults striking NE–SW have been reactivated in transpression recently in the Rhenish Massif (e.g. Ahorner, 1975, 1983; Hinzen, 2003; Bourgeois et al., 2007). The 1978 Roetgen Earthquake (ML = 3.0) displays a thrust faulting regime in the uplift region of the Hohes Venn (Ahorner, 1983). Hydraulic fracturing experiments down to 400 m depth in the borehole Konzen (25 km SSE of Aachen) also indicate a reverse-faulting stress regime with the maximum horizontal stress direction at 120° ± 10° (SE–NW, Rummel and Baumgartner, 1985).

The RWTH-1 borehole is situated in a complex intersection of major active fault trends and associated changes in seismic activity. Its study therefore offers the unique possibility to investigate the local in situ stresses down to 2.5 km depth.

2. The RWTH-1 well

The 2544 m deep Aachen borehole was drilled in the year 2004 with the purpose to supply geothermal heat for the heating and cooling system of the new student service centre “SuperC” of the Aachen University. The borehole diameter decreases from 23 in. in the upper 240 m, to 17.5 in. between 240 and 1250 m to 8.5 in. down to total depth. Deviation accumulates to 15° between 1600 and 1700 m depth and decreases to 10° at total depth (Fig. 2). A total of 146 m core samples were cut in the depth intervals of 1392–1516 m, 2128–2143 m and 2537–2545 m. The 2 in. cores from the first and second core section were obtained by a coring while drilling system (CoreDrill™), while the core at total depth was cut by a conventional system and has a diameter of 4 in.

Beyond a thin Quaternary and Cretaceous cover, the borehole penetrates strongly compacted siliciclastic rocks (claystones, siltstones and sandstones) of Carboniferous and Devonian age, which were deposited in the Rhenohercynian basin and deformed during the Variscan orogeny. The Lower Carboniferous and Upper Devonian limestones (e.g. “Kohlenkalk” and “Massenkalk”), which form prominent outcrops SE of Aachen, are lacking in the drilled strata (Ribbert, 2006; Becker, 2008). The first core section (128 m of 146 m core) is characterized by intensive deformation and hydrothermal vein formation within a fault zone (Sindern et al., 2007).

An extensive borehole logging campaign was carried out from 250 to 2530 m. In addition to the standard logs it also includes the Cross Multipole Array Acoustic Log (XMAC™ — Baker Atlas), which provides the horizontal shear wave velocity anisotropy and the azimuth of the fast shear wave, and the Simultaneous Acoustic and Resistivity imager (STAR™ — Baker Atlas) below 1263 m. The latter was only run on the 8.5 in. section because of its limitation to small borehole radii. It delivered high-resolution image logs of the borehole wall which allow a detailed analysis of drilling-induced rock failure (see below) and natural fractures (Trautwein-Brüns et al., submitted for publication). Above 1263 m only hexagonal dipmeter data are available for structural interpretation.

Due to environmental and safety regulations at the inner city location of the borehole, no active radiation source was run in the borehole, therefore no density measurements are available. Density was then calculated from velocity using a density/velocity calibration from high-resolution core-logger measurements (Fig. 3). A reservoir characterization instrument (RCI, Baker Atlas™) which was run to collect fluid samples and to obtain pore pressures did not obtain suitable results due to the very low in situ porosity and permeability of the rocks encountered. For technical reasons direct measurements of minimum horizontal stress by minifract tests using a straddle-packer system failed. Comparable measurements such as Leakoff Tests or Formation Integrity Tests were not an option in this borehole.

Downhole measurements of gamma radiation, ultrasonic velocities and resistivity revealed layered and rather uniform siliciclastic compositions of the drilled sequence (Fig. 2). Gamma radiation ranges from 15 GAPI in the Lower Devonian quartzitic sandstones to 180 GAPI in Upper Carboniferous organic shales. P-wave velocity varies between 3.5 km/s in coal layers and organic shales and 7.0 km/s within the only limestone unit in the borehole at 1408 to 1438 m depth. The Lower Devonian siliciclastic rocks have a range between 4.5 and 5.7 km/s (Pechnig, 2005). Shear wave velocity ranges from 2.4 to 3.8 km/s. The resistivity log reaches values of up to 1000 Ωm in quartz-cemented sandstones, while organic shales have resistivities lower than 100 Ωm (Pechnig, 2005).

The generally high level of ultrasonic velocities reflects the strongly compacted character of the siliciclastic rocks. The rock cuttings and the core material show no visual porosity, and helium porosities are below 0.7. In order to determine reliable rock mechanical input parameters for stress modelling confined single stage failure tests were performed in the laboratory. The samples were chosen from the core material of the RWTH-1 well in order to cover the lithological range defined by different gamma radiation and velocities. The stress at failure is plotted as a function of confining pressure (Fig. 4) and the unconfined compressive strength is extrapolated from the intercept of the failure line corresponding to failure stress at zero confining pressure. The unconfined compressive strength reaches 336 MPa for the quartzitic sandstone of the core at bottom hole. The strength of the siltstones strongly scatters due to the range in lithology, depth and degree of fracturing. If a sample failed along an existing fracture plane (siltstones
and fractures reactivated) the strength is considerably reduced (40 MPa). The failure stress of the mudstones samples is comparable to that of the siltstones samples. This reflects the strong compaction of the rock material.

The dipmeter and image logs show that bedding in the upper part of the borehole generally dips to SE. This changes between 1300 and 1463 m towards gently NNE-dipping trends (dip 20°). Heterogeneous dip orientation, development of cleavage and increased dip angles are observed within fault and fracture zones sealed by veins which can be clearly depicted on the image logs. The uppermost core section shows the strongest structural disturbance of all imaged borehole sections (1263–2520 m). Structural core descriptions reveal fracture planes with several orientations, slickensides on bedding and fractures with different types of vein filling (Sindern et al., 2007; Loegering, 2008). Fluid inclusions investigated by Loegering (2008) attribute these structures to Variscan deformation.

The interpretation of large-displacement faults in the borehole is problematic due to the uniform lithology and missing stratigraphic marker horizons (carbonates and conglomerates), which would show stratigraphic duplications or gaps. However, several low angle fractures and fault planes of Variscan trend (NE–SW) have been observed in the resistivity image log. Similar fracture and fault zones have also been documented in outcrop studies (Becker, 2007) and in seismic data (seismic line BEB 8001 traverses only a few hundred meters NE of the borehole location, Becker, 2008), which documents thrust faults in the vicinity of the borehole with displacements between 300 and 1800 m (Becker, 2008).

Another set of fractures is observed in the acoustic as well as in the resistivity images (Fig. 5). These fractures trend NW–SE or NNE–SSW and basically dip towards the NE with varying dip angles. The most prominent fault zone of this set is observed at 1895 m. It is characterized by an increased caliper due to large washouts, open fractures, increased anisotropy and reduced velocity. This fault also marks the boundary between an upper clay-dominated and a lower sand-dominated part of the Lower Devonian and comes along with a change in the dominant rock colours from red–green to gray (Becker, 2008). No borehole breakouts occur in the vicinity of the fault and drilling-induced fractures are rotated (see below), which could indicate recent slip along that fault plane (Shamir and Zoback, 1992; Barton and Zoback, 1994).

Fig. 2. Composite log of the borehole caliper, the borehole deviation, the natural radioactivity from the gamma radiation log, the compressive and shear wave velocities and the resistivity measured in well RWTH-1. MD = measured depth.
3. Borehole failure analysis

Borehole failure occurs where the in situ stress exceeds the rock strength. In analysing borehole failure it is usually assumed that the three principal stresses at depth are the vertical stress $S_v$ and the maximum and minimum horizontal principal stresses, $S_{H_{\text{max}}}$ and $S_{H_{\text{min}}}$ (Zoback et al., 2003). In the RWTH-1 borehole a Simultaneous Electrical and Acoustic Imager of the borehole wall (STAR tool, Baker Atlas™) was run downward from 1263 m depth and delivered excellent, high-resolution images to study the occurrence and distribution of stress induced borehole failure, i.e. compressive borehole breakouts and drilling-induced tensile fractures.

3.1. Failure occurrence and distribution

Compressive borehole breakouts form in the area of maximum circumferential stress, which in vertical wells is found at the azimuth of $S_{H_{\text{min}}}$ (Fig. 6). The breakout width is defined by the angle that is span by the broken out zone at the borehole wall. Drilling-induced tensile fractures form at the azimuth of $S_{H_{\text{max}}}$.

All available data are used to constrain a geomechanical model for the location. A geomechanical model consists of the rock mechanical properties such as the rock strength (unconfined compressive strength and friction coefficient), the Poisson ratio, the Biot coefficient and the stress magnitudes, the orientation of the maximum horizontal stress and the pore pressure. The rock mechanical properties are derived from laboratory tests and logs and the vertical stress can be calculated from integrated densities. Thereafter the occurrence and type of failure is used to constrain the in situ stress field at the location.

Fig. 3. Correlation of neutron density and compressive wave velocity from high-resolution core-logger data of the first core section (1390–1515 m).

Fig. 4. Failure stress observed in triaxial single failure tests at different confining pressures. The unconfined compressive strength UCS (at zero confining pressure) has been extrapolated from the linear trend (numbers in plot). Existing fractures have been reactivated during the compression tests for the fractured siltstone samples.

Fig. 5. Stereographic Schmidt-net projection (lower hemisphere) of open fault and fracture planes picked in the acoustic amplitude images show two preferred trends (NW–SE and NNE–SSW). Most structures are steeply dipping to NE or ESE. The most prominent fault of these intersects the borehole at 1895 m.

Fig. 6. Theoretical sketch of borehole failure for a vertical well with respect to the principle horizontal stress $S_{H_{\text{min}}}$ and $S_{H_{\text{max}}}$. Compressive borehole breakouts form in direction of $S_{H_{\text{min}}}$ and $S_{H_{\text{max}}}$. Drilling-induced tensile fractures form at the azimuth of $S_{H_{\text{max}}}$.
On acoustic image logs breakouts appear as dark bands of low acoustic amplitude on opposite sides of the borehole wall. Fig. 7 shows typical examples of breakouts from the RWTH-1 borehole. Between 1440 and 1895 m the shape of breakouts often follows the planes interpreted as cleavage (Fig. 7a). Breakouts below 1895 m are generally smaller, presumably due to the sandier lithology and thus higher rock strength (Fig. 7b).

Drilling-induced tensile fractures open in direction of Shmin. Induced fractures were detected in the resistivity image and occur as low-resistive (dark) fractures, as they are filled with drilling mud. Fig. 7c shows an example of drilling-induced tensile fractures and drilling-enhanced natural fractures at 2325 m depth. Often it is not possible to distinguish between drilling-induced and drilling-enhanced natural fractures. Near-vertical features at opposite sides of the borehole are interpreted as drilling-induced fractures, while inclined, partly open, sinusoidal features are interpreted as drilling-enhanced natural fractures.

The breakouts are consistently oriented with a mean azimuth of 45 ± 14° (Fig. 8a), which indicates a NE–SW orientation of the minimum horizontal stress. Fig. 9 shows the correlation of breakout width with important petrophysical rock properties (GR, velocity and resistivity). As clearly shown by the dependence from GR the breakout width increases with increasing clay content. Smaller breakouts are formed in the lower borehole section due to the more sandy lithology (breakout width < 40°), while breakout in the upper part show widths of more than 32°. This is also manifested in the mean breakout width, which is 46 ± 18° for the breakouts between 1440 and 1895 m and 29 ± 16° for the breakouts below (Fig. 8b). The dependence from lithology is also evident in other correlations with petrophysical properties: the breakout width tends to increase with decreasing vp, vs and resistivity and increasing vp/vs ratio. For the RWTH-1 well, breakouts with larger breakout width also tend to have a larger caliper. While observed frequently in wells, this is not necessarily the case.

The azimuth histogram of the induced fractures shows two maxima, one close to 110° and the other at 125° (Fig. 8c). The rose diagram (Fig. 8d) summarizes the results concerning the orientation of the maximum horizontal stress. Considering all breakouts (breakout azimuth + 90°) and induced fractures from the imaged borehole section (1263–2520 m) yields a mean direction of Shmax of 126 ± 15°. The SE–NW direction is in accordance with the regional stress field.

Most breakouts occur between 1400 and 1895 m and correspond to the fine-grained mud- to siltstone sequences of lower Devonian age (Fig. 10). Below the fault zone and lithostratigraphic boundary at 1895 m the frequency of breakouts is clearly decreased and correlates to the lithofacies changes from shale-dominated to sand-dominated rocks. Between 1895 m and 2160 m as well as below 2350 m only a small number of breakouts are observed generally in association to fractures and fault zones. Clear breakouts are also formed in the depth interval between 2160 and 2350 m within the more sandy part of the Lower Devonian rocks.

The amount of drilling-induced tensile fractures increases with depth (see Fig. 10). The occurrence of near vertical induced fractures with a maximum inclination of 10° (mean value 1.7°) supports our assumption that one principal stress is vertical and nearly parallel to the borehole axis.

3.2. Rock mechanical properties

The clear correlation of breakout width to lithology dependent log properties (gamma radiation, resistivity and velocity, Fig. 9) shows the influence of rock properties on the rock strength. That means that the occurrence of breakouts depends on lithology, which can then be used for stress magnitude calculations based on breakout width. There is no tool to directly measure compressive rock strength in boreholes. Commonly empirical equations in combination with laboratory strength
test are used to derive rock mechanical parameters from downhole measurements. In this study UCS was estimated from the downhole measurement of shear wave velocity based on the empirical equation published by Urai (1995) and Ingram and Urai (1999):  

$$\log \text{UCS} = -6.36 + 2.45 \log v_s$$  \hspace{1cm} (1)$$

with UCS in MPa and vs in m/s. UCS measurements on RWTH-1 core samples range from 40 MPa to 336 MPa and agree with Eq. (1) (Fig. 11). The small number of UCS–vs data available from the RWTH-1 borehole doesn’t allow the derivation of a local relation of UCS from velocity. Therefore we use this empirical relation (Eq. 1), which was derived from a wide range of data on well consolidated, siliciclastic rocks comparable to those from the RWTH-1 borehole.

The average of log derived UCS values for the mudstones yields $124 \pm 16$ MPa and for sandstones $159 \pm 27$ MPa. P10 values (i.e. 10% of calculated UCS values are below or equal) are 94 MPa and 131 MPa and P90 (i.e. 90% of calculated UCS values are below or equal) values 162 MPa and 191 MPa for shale and sandy sections, respectively. The GR is used to distinguish between shaly and sandy sections.

The XMAC tool measures the horizontal anisotropy based on the principle of shear wave splitting. The tool provides the magnitude and orientation of the fast and slow shear waves. In the RWTH-1 borehole the anisotropy is increased at faults and within fracture zones. The velocity of the slow shear wave, which is lower in fracture zones, is taken for the calculation of UCS (Fig. 10) in order to account for the anisotropy and the reduction of rock mass strength due to fractures. The pore pressure coefficient ($\alpha$) is needed to calculate the effective stresses acting on the rock (effective stress = absolute stress − PorePressure × $\alpha$). It can take values between 0 and 1. Rocks with grain compressibility much smaller than the matrix compressibility, which are often compressive or highly porous rocks, tend to have values close to 1. Warpinski and Teufel (1992) state that samples rich in microcracks also tend to have values close to 1. For rock with low matrix compressibilitites or low porosity rock the Biot

$$\mu = \left( \frac{v_p^2 - 2v_s^2}{2(v_p^2 - v_s^2)} \right).$$  \hspace{1cm} (3)$$

Log ratios below 1250 m MD have mean values of $0.26 \pm 0.03$ and $0.19 \pm 0.04$ for mudstones and sandstones respectively. P10 values are 0.14 and 0.24, and P90 values 0.26 and 0.29, respectively. Laboratory values range from 0.10 to 0.23 with an average value of 0.18. The calculated values for the dynamic Poisson ratio are higher than suggested by lab test results. This may reflect the difference between dynamic and static values. On the other hand, Poisson ratios derived from lab tests are often ambiguous. We used a constant Poisson ratio of 0.25 for the stress modelling. Note, that the Poisson ratio has negligible influence on constraining the stresses as long as temperature effects can be neglected. The pore pressure coefficient was determined from the lab tests and used for the stress calculations.

Fig. 8. Statistical distribution of (a) the orientation of breakouts between 1440–1895 m and 1895–2520 m, (b) the width of breakouts 1440–1895 m and 1895–2520 m, (c) the azimuth of drilling-induced tensile fractures below 1440 m and (d) the orientation of Shmax based on borehole failure analysis (azimuth breakouts = 90° and azimuth induced fractures). Shmax trends SE–NW (126° ± 15°).

$$\text{IntFric} = \tan(\alpha \sin((v_p-1)/(v_p+1)))$$  \hspace{1cm} (2)$$

with vp in km/s. For shaly sections average values yield $0.89 \pm 0.03$, P10 0.85 and P90 0.96. The sandy section result in slightly higher friction coefficients with an average of $0.92 \pm 0.18$, and P10 and P90 values of 0.87 and 0.97, respectively. The log derived values are significantly higher than the lab result which implies that the relationship doesn’t work for this case. As the UCS has a significantly higher influence on the rock failure, we used a constant friction coefficient of 0.6 for the stress modelling.

The dynamic Poisson ratio $\mu$ is calculated from downhole compressional and shear wave velocities:

$$\mu = \frac{v_p^2 - 2v_s^2}{2(v_p^2 - v_s^2)}.$$  \hspace{1cm} (3)$$

The pore pressure coefficient was determined from the lab tests and used for the stress calculations.
A lower Biot coefficient is comparable with higher effective stresses while the differential stress remains the same. Higher confining stresses also imply that higher differential stresses are necessary to fail the rock. In other words, for lower Biot coefficients higher (absolute) stresses result in the same amount of failure.

3.3. Constraining stress magnitudes

The stress magnitudes are evaluated based on the observation that the breakout zone corresponds to the region where the state of stress exceeds the rock strength (e.g. Chang, 2004). Consequently, the combination of unconfined compressive strength and breakout width allows the estimation of in situ stress magnitudes (e.g. Zoback et al., 2003).

The overburden stress is calculated by integration of the mean overburden density of the overlying rock. Due to the absence of a density log a density–velocity relation was developed based on high-resolution lab-based core-logger measurements of the first core section (Fig. 3) which then allowed the inversion of bulk densities from the downhole sonic measurements. The calculated bulk densities vary between 2.54 and 2.81 g/cm³ (mean value 2.71 g/cm³) and are in good agreement with laboratory measurements on cuttings and core samples. As the sonic velocity is fairly constant especially in the Lower Devonian sections, the vertical stress was calculated by a constant density gradient of 2.71 g/cm³ which indicates an overburden stress of 68 MPa at 2500 m depth.

An in situ measurement of pore pressure is not available, as sampling using a formation tester (RCL, Baker Atlas) was not successful. Several points of interest – formation as well as fracture

![Fig. 9. Correlation between breakout width and downhole petrophysical properties and caliper. The breakouts found in the upper part (1440–1895 m) are marked in gray, those of the lower part (1895–2520 m) in black: Correlation to (a) natural gamma radiation (GR), (b) resistivity, (c) compressive wave velocity (vp), (d) shear wave velocity (vs), (e) ratio of compressive to shear wave velocity (vp/vs) and (f) borehole caliper (CAL) indicate a clear lithology dependence of breakout occurrence.](image-url)
zones – were tested, but delivered no results because of the tight formation. In accordance with regional considerations and the lack of log anomalies pointing towards elevated pore pressures, hydrostatic pore pressures were assumed.

The frictional strength of the governing fracture system is defined by the coefficient of sliding friction. Townend and Zoback (2000) have demonstrated that coefficients of friction between 0.6 and 1.0 are consistent with measured stress states in the upper crust. Byerlee (1978) has reviewed laboratory results of frictional strength for different rocks and has shown that at high normal stress the friction is nearly independent of rock type. Friction scatters between 0.6 and 1.0. Byerlee (1978) fits the data resulting in frictional coefficients of 0.85 up to 2 kbar and 0.6 above 2 kbar normal stress. We accordingly used a coefficient of sliding friction of 0.85 for modelling.

A mud weight slightly above hydrostatic pressure was used during drilling in order to stabilize the borehole. While drilling the section below 1260 m the mud pressure was kept between 1.02 and 1.08 g/cm³. In the absence of direct downhole pressure data, the maximum mud weight of 1.08 g/cm³ measured at surface was used to calculate the mud pressure at depth.

For the following modelling of the horizontal stresses we used the software package GMI-SFIB™ (Stress and Failure of Inclined Boreholes, GMI, 2008). As introduced by Zoback et al. (2003), the range of possible horizontal stress magnitudes at depth was calculated based on the limits of frictional strength of the earth crust, which is defined by the coefficient of sliding friction. Within these limits a certain breakout width corresponds to the stress field, the rock strength and the wellbore orientation and thus can be used to constrain the horizontal stresses. Input data for each case study are the magnitude of the vertical stress, the pore pressure, the orientation of SHmax, the borehole orientation, the width of breakouts and the rock strength. If the minimum horizontal stress is known it puts an additional

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![Fig. 10. Composite log of borehole failure, rock strength and faults: tracks 1 and 2 show the calculated unconfined compressive strength (UCS) and the distribution of borehole failure (compressive borehole breakouts BO, drilling-induced tensile fractures IF), track 3 indicates faults that were interpreted to be active. The last track shows a comparison between the observed breakouts (black) and the predicted/modelled breakouts (gray). The predicted/modelled breakouts were calculated along the well path of RWTH-1 on the basis of the derived geomechanical model (stresses, pore pressure, rock properties, mud weight = 1.08 g/cm³).](image-url)
constraint on the calculated maximum horizontal stress. However, as mentioned in Section 2 no borehole tests are available to constrain the least horizontal stress.

Based on the described input parameters possible horizontal stress ranges at depth are calculated and mapped within a four-sided stress polygon (Zoback, 2007) in a SHmax–Shmin-plot (see example Fig. 12). The outlines of the polygon define the limits of allowable stress differences by the Mohr–Coulomb frictional faulting theory for the assumed frictional equilibrium of pre-existing faults. Based on the relation of the principal stresses the stress conditions consistent with a normal faulting, strike-slip faulting and reverse faulting regime according to Anderson’s faulting theory are outlined.

The red contours mark the compressive strength of the rock, which is consistent with the observed breakout width using Mohr–Coulomb failure criterion. If breakouts occur and the compressive strength of the rock is known, the plots show possible ranges of minimum and maximum horizontal stresses. Shmin and SHmax combinations that fall above the compressive failure line for a given rock strength indicate breakouts of a certain width or larger to occur, while those that fall below indicate that no breakouts or breakouts of less than the given width should be observed. As the compressive failure lines are gently sloped, errors of the minimum horizontal stress have minor impact on the magnitude of SHmax.

The blue contours mark the tensile rock strength. Stress magnitudes that plot left of the tensile 0-line indicate stress states consistent with the occurrence of drilling-induced tensile fractures assuming no tensile rock strength. The existence of compressive as well as tensile wellbore failures puts narrower bounds on possible stress states at depth (Zoback et al., 2003), because the maximum horizontal stress has to exceed compressive rock strength while the minimum horizontal stress must be small enough to allow tensile rock failure.

3.4. In situ stress regime

Drilling-induced tensile fractures occur in vertical wells only if there is a significant difference between the two horizontal stresses (Zoback, 2007). Using a coefficient of sliding friction of 0.6 Zoback (2007) has shown that the conditions for the occurrence of drilling-induced tensile fractures around a vertical wellbore in the absence of excess mud weight or wellbore cooling are essentially identical to the values of Shmin and SHmax associated with a strike-slip faulting regime in frictional equilibrium. A higher coefficient of sliding friction (0.85 in this study) opens a small window, where the generation of induced fractures is possible even at normal faulting conditions at a very small minimum horizontal stress (Fig. 12). If in addition a differential temperature between the borehole fluid and the formation...

Fig. 11. Empirical relation between UCS and shear wave velocity after Urai (1995) and Ingram and Urai (1999). The UCS data of RWTH-1 derived from triaxial single failure tests on core plugs of the RWTH-1 borehole plot within the range of well consolidated siliciclastic rocks used by Urai (1995) to develop the relation.

Fig. 12. Example for the calculation of in situ stress magnitudes at 2314 m (TVD = 2272 m, vertical stress Sv = 60.4 MPa, pore pressure = 22.3 MPa, mud pressure = 24.1 MPa). The borehole wall shows borehole breakouts (breakout width = 37°) as well as drilling-induced tensile fractures indicating an orientation of SHmax = 126°. The stress polygon is modelled after Zoback (2007) based on Mohr–Coulomb Theory of Sliding Friction for a coefficient of sliding friction = 0.85. The Mohr–Coulomb failure criterion was used to calculate the red lines of compressive strength in accordance with the observed breakout width (coefficient of internal friction = 0.6, poisons ratio = 0.25). Further model parameters are: Biot coefficient of effective stress = 1, borehole azimuth = 268°, borehole deviation = 11.5°. Blue lines = tensile strength, NF = normal faulting regime, SS = strike-slip faulting regime, RF = reverse faulting regime.
comes into play, drilling-induced fractures may develop for lower stress differences of the far field stress and thus for higher values of $S_{\text{shmin}}$. At the same time cooling will stabilize the well with respect to breakouts. However since this is a temporal effect only, it will not be considered here. The temperature effect is caused by additional thermal stresses at the wellbore wall. Our modelling software computes the thermal stresses after Stephens and Voight (1982) as a function of the temperature difference, the linear coefficient of thermal expansion, the static Young's modulus and the Poisson ratio. The approximate formula does not consider thermo-elastic coupling. When applying cooling the blue tensile failure lines in move to the right (see examples Fig. 14c and d). This demonstrates, that if temperature effects are significant, drilling-induced tensile fractures could form in a normal faulting or even a reverse faulting regime. We will show below, that the high strength of the strongly compacted Lower Devonian rocks require high maximum horizontal stress for compressive borehole failure (breakouts), which indicates a strike-slip faulting regime. In the following, we will first constrain the stress field not considering any temperature effects and then discuss the potential temperature effect in a separate section further down.

The magnitudes of horizontal stress have been estimated at 25 depth positions between 1435 m and 2337 m (Table 1). As no direct measurement of minimum horizontal stress is available, stress magnitudes were initially estimated by modelling cases with both, breakouts and drilling-induced tensile fractures. An example is shown computed for a depth of 2314 m (TVD = 2272 m, Fig. 12). The borehole wall shows both compressive breakouts of 37° width, and drilling-induced tensile fractures. The compressive failure lines are calculated using the Mohr–Coulomb failure criterion with an internal friction of 0.6. As the compressive strength is the most sensitive parameter for the derivation of stress magnitudes, the failure is determined by the log derived range of the compressive strength within the breakout interval. As tensile as well as compressive failure is observed, the horizontal stresses could be constrained quite well at the intersection of the tensile and compressive failure lines. The calculated strength of the intact rock based on Eq. (1) is 146 ± 24 MPa and constrains the possible horizontal stress magnitudes to $S_{\text{shmin}} = 44 ± 2$ MPa and $S_{\text{SHmax}} = 87 ± 10$ MPa.

The cases with drilling-induced tensile fractures and compressive borehole breakouts indicate a minimum horizontal stress gradient of 0.019 MPa/m. This trend was then used to model additional cases at depths, where only compressive borehole breakouts are observed. The results are compiled in Table 1 and plotted in Fig. 13. Solid lines mark the hydrostatic pore pressure, the vertical stress and the derived trend of minimum horizontal stress ($S_{\text{shmin}} = 0.019$ MPa/m). The $S_{\text{SHmax}}$ shows a gradient of 0.038 MPa/m on average. This is nearly twice as much as for $S_{\text{shmin}}$. This clearly indicates a strike-slip faulting regime ($S_{\text{shmin}} < S_{\text{V}} < S_{\text{SHmax}}$). Stress magnitudes and differences between the stresses increase with depth.

Tensile wall fractures are more likely to form when wellbore pressure is larger than the pore pressure and when the mud is much cooler than the formation temperature (Zoback, 2007). While drilling the RWTH-1 borehole, the mud weight was only slightly increased to fresh water, but the real mud pressure at bit was not detected. Every time the tool-string runs into the borehole due to changes of the drill bit, operational problems or borehole cleaning operations, the downhole mud pressure could be considerably increased due to a piston effect. As no correlation of drilling-induced tensile fractures and these operations was observed, it is assumed that the slightly increased mud pressure was a realistic assumption for stress modelling.

A differential temperature between the formation and the cooler drilling mud will result in thermal stresses at the borehole wall which will allow drilling-induced tensile fractures to form more easily. A formation temperature gradient of 0.03 °C/m was derived for the location from temperature logs taken two years after drilling took place (no heat production in the meantime). Temperature data for the drilling mud are only available from the flowline at surface. Flowline temperatures are between 30 °C and 55 °C for the borehole below 1263 m and maximal 40 °C below formation temperature at the time of reporting (Fig. 14a). Temperature differences below 2100 m range mostly between 20 and 30° with maxima around 40° and are clearly smaller above. As the surface temperature of the drilling mud is assumed to be lower than at the drill bit, the calculation with a differential temperature of 40 °C means a worst case scenario. Next to the temperature difference the thermal stresses depend on the linear coefficient of thermal expansion $\alpha$, the static Young's modulus E and

### Table 1

Unconfined compressive strength and modelled stress magnitudes.

<table>
<thead>
<tr>
<th>TVD (m)</th>
<th>Vertical stress (MPa)</th>
<th>Pore pressure (MPa)</th>
<th>Breakout width (°)</th>
<th>UCS (MPa)</th>
<th>$S_{\text{shmin}}$ (MPa)</th>
<th>$S_{\text{SHmax}}$ (MPa)</th>
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Fig. 13. In situ stress magnitudes observed from borehole failure analysis. Solid black lines mark the hydrostatic pore pressure ($P_p$), the vertical stress ($S_v$) and the derived trend of minimum horizontal stress ($S_{\text{shmin}} = 0.019$ *TVD) and maximum horizontal stress ($S_{\text{SHmax}} = 0.038$ *TVD).
the Poisson ratio \( \mu \). Thermal stresses and thus the temperature effect increases with increasing \( \alpha_T \), \( E \) and \( \mu \).

Zoback, 2007 shows values for \( \alpha_T \) for different rock types published by Griffith (1936). Griffith (1936) found a correlation of \( \alpha_T \) and the content of silica in the rock, which is reasoned in the significantly higher thermal expansion coefficient for quartz than for most other rock forming minerals. The quartz content of the cores of the RWTH-1 was determined by X-ray diffraction technique with quantitative Rietveld Analysis (Becker, 2008) and ranges between 33% for the mudstones and 72% for the sandstones with the sandstones at total depth having values of 66%. The correlation found in Griffith data suggests a maximum coefficient of thermal expansion \( \alpha_T \) of \( 8 \times 10^{-6} \) °C\(^{-1} \). The mudstones would rather have an \( \alpha_T \) of \( 4 \times 10^{-6} \) °C\(^{-1} \). To allow for uncertainties we use a maximum \( \alpha_T \) of \( 8.5 \times 10^{-6} \) °C\(^{-1} \) for modelling the worst case scenario (Fig. 14c).

Fig. 14b shows values of the dynamic (log derived) \( \mu \) and the static \( E \) for the depth sections where drilling-induced tensile fractures were observed. The static \( E \) was estimated from the dynamic by using a factor of 0.8, to better fit the values of \( E \) measured in the laboratory. Clearly, drilling-induced fractures were preferably observed in sandier lithologies (GR > 115), with lower Poisson ratios (\( \mu < 0.28 \)) and larger static Young’s moduli (\( E > 45 \) GPA). At the same time low Poisson ratios imply high Young’s moduli.

Fig. 14c shows the modelling results for the worst case scenario calculated at 2314 m depth (TVD = 2272 m, compare to Fig. 12). An intermediate \( \mu = 25 \) and \( E = 51 \) GPA, as found from logs for that depth,
a maximal temperature difference of 40°C and a thermal expansion coefficient $\alpha_T$ of $8.5 \times 10^{-6} \text{ °C}^{-1}$ were assumed. The worst case scenario shows that the minimum horizontal stress would increase by 14.7 MPa, if assuming the most likely rock strength and no tensile rock strength. At the same time $\sigma_{max}$ would only increase moderately by 3.1 MPa. Modelling of the temperature effect further up the well where temperature differences are smaller ($\Delta T = 20^\circ \text{C}$) and the lithology is more shaly ($\alpha_T = 6.5 \times 10^{-6} \text{ °C}^{-1}$), results into much smaller differences as compared to the original modelling results (see example 1625 m, Fig. 14d). $\sigma_{min}$ and $\sigma_{max}$ would increase by 3.5 MPa and 0.7 MPa, respectively.

In summary, all scenarios even for the maximum temperature difference of 40°C result into a strike-slip stress regime. The correlation of the occurrence of drilling-induced tensile fractures and the elastic moduli may indicate that cooling helped to form the drilling-induced fractures. This would suggest a potentially higher $\sigma_{min}$ than derived above and shown in Table 1 and a slightly higher value of $\sigma_{max}$. Considering tensile strength of the rock on the other hand would reduce the magnitude of $\sigma_{min}$.

To verify the stress model we used the geomechanical model and the mud weight at drilling as an input to predict compressive failure along the well path of RWTH-1 well. We then compare the predicted failure to the failure observed in the image logs. The last track of Fig. 10 shows the comparison of the observed and the predicted breakouts, which were calculated using the GMI software GM WellCheck™ (GMI, 2010). The predicted and the observed failure distributions are in general agreement. The width of the observed breakouts tends to be slightly larger than the predicted. This may be explained by a reduction of rock strength in highly fractured zones, which is possibly not sufficiently accounted for by calculating the UCS using the slow shear velocity. Below 2350 and commonly at fault zones breakouts are predicted but not observed. This may be attributed to stress perturbation or stress drop at recently active fault zones as for example described by Shamir and Zoback (1992) or Barton and Zoback (1994, Fig. 15). However, a detailed analysis is out of scope of this paper.

Results of the borehole failure analysis clearly indicate a strike-slip faulting regime. The magnitude of the maximum horizontal stress may be lower than modelled, if the rock strength is overestimated by the empirical Eq. 1 used. On the other hand, using a Biot coefficient smaller than 1 which may be indicated by the strongly compacted and low porosity rock (He-Expansion <0.7%) would require even higher magnitudes of the maximum horizontal stress to explain both the compressive and the tensile wellbore failure. If a thermal stress played a role in the formation of drilling-induced tensile fractures, the differential stress could be significantly lower, but stress magnitudes remain consistent with a strike-slip faulting regime.

4. European stress regime changes reflected in the Aachen borehole

Central Europe displays a complex history of structural deformation since Paleozoic times. Mountain building events during Caledonian and Variscan orogenies are being followed by structural reorganisations mainly associated with the opening of the North Atlantic and the Alpine collision. Therefore a complex structural grain is observable including the entire inventory of tectonics such as thrust faulting, normal and strike-slip faulting as well as inversion structures. A general structural model proposed by Jager (2007) is therefore one of repeated (oblique) reactivation of basement faults which continue to control the structural grain despite changes in tectonic regime and stress direction. According to Müller et al. (1997) Europe is characterized by an almost homogeneous SE–NW orientation (145°) of the maximum horizontal compression, but the tectonic regime experiences short-scale lateral variations from the dominant strike-slip (49% of the data) to normal faulting (30%) and to thrust faulting (21%). This is conceptualized by a tectonic model consisting of upper crustal fragments decoupled from the lithospheric mantle by the ductile lower crust, wherein the lateral boundaries of the fragments fail in thrust, strike-slip or normal faulting depending on their orientation with respect to the far field stress (Müller et al., 1997).

In the study area, the main structural elements have been formed during the Variscan Orogeny with NE–SW trending folds and thrust faults demarcating the front of the Variscan Mountain Belt. The northernmost thrust ("Aachen–Midi Thrust") which intersects the surface only a few hundred meters south of the RWTH-1 borehole, is grossly in line with the North Artois dextral shear zone defined in the north of France and is presumably connected to the Hainaut Shear Zone on Belgian territory (Leynaud et al., 2000). This was already proposed by Ahorner (1975) who defined the whole seismotectonic zone south of the Brabant Massiv from Aachen to the Channel coast near Ostend as Brabant shear zone or Belgian strike-slip zone.

In contrast to this Variscan structural grain, the Lower Rhine Embayment, which is part of the recently active European Cenozoic rift system of western and central Europe basically consists of several NW–SE trending graben structures, such as the tectonically active Roer Valley Graben which itself is bordered by major normal faults (e.g. Peel Boundary Fault Zone and Feldbiss Fault Zone). As the Aachen borehole is located on the SW flank of the Roer Valley Graben locally framed by the normal faults of the Laurensberger Fault system a recently active normal faulting regime was expected for the Aachen borehole.

In this study borehole failure analysis in the RWTH-1 borehole below 1263 m depth showed that the maximum horizontal stress trends SE–NW which is in accordance with the general west European stress trend and the regional orientation of the major structural trend within the Lower Rhine Embayment. An anticlockwise rotation of the minimum horizontal stress component from 50° (NE–SW) in the Roer Valley Graben to 20° (NNE–SSW) in the Rhenish Massif, 16° (NNE–SSW) in the Liege region and 354° (N–S) in the Brabant Massif has been observed by Camelbeeck and Eck (1994). Therefore, the azimuth of minimum horizontal stress observed from breakouts in the RWTH-1 borehole (45° ± 14°) furthermore suggests, that the borehole location should belong to the seismotectonic zone of the Roer Valley Graben.

The calculated magnitudes of principle stresses are consistent with a strike-slip faulting regime despite the fact that the location of the borehole is within the SW-bordering faults of the Roer Valley Graben System, for which pure normal faulting activity is suggested by earthquake focal mechanism and surface investigations. The occurrence of drilling-induced tensile fractures within the borehole reveals a significant difference between the horizontal stresses and further support the strike-slip faulting regime.

Overlying and small scale changing stress regimes have been observed in the region. Investigations of focal mechanisms for the Rhine Graben including the Rhenish Massif and the Lower Rhine Embayment by Plenefisch and Bonjer (1997) indicate that the structural regime is between strike-slip and extensional deformation. Hinzen (2003) combined all earthquake data from the Lower Rhine Embayment and displayed a mix of strike-slip and extensional events. He proposed that the prevalence for one or the other is controlled by event depth and geographic distribution. Camelbeeck and Eck (1994) argued from earthquake focal mechanisms for a complex tectonic deformation field in the region extending from the Roergraben to the city of Liège (40 km SW of Aachen). While the Roergraben is dominated by normal faulting, strike-slip earthquakes occur in the area of Liege (Leynaud et al., 2000).

The observed strike-slip deformation is in accordance with the regional trend towards a strike-slip regime proposed by Hinzen (2003). He interpreted the extensional stress regime in the shallow crust of the Lower Rhine Embayment as a local effect overlaying the
Fig. 15. Potential stress perturbation at the 1895 m-fault. Right below the fault at 1895 m breakouts are predicted from the rock mechanical model but are not observed in the image logs. The drilling-induced fractures on the image also appear to be shifted in orientation.
regional trend towards a strike-slip regime, whereas the Rhenish Massif acts as a buffer zone for the stress in the shallow crust between the two sections of the Rhine rift system.

After Ahorner (1975) the orientation of crustal weakness zones in relation to the regional stress field define the style of large-scale seismotectonic deformation, which is predominantly of strike-slip type along major fracture zones striking NNE or WNW like the Upper Rhine graben or the Belgian/Brabant zone, respectively, because these zones of crustal weakness are arranged parallel to the direction of maximum shear stress (Ahorner, 1975). On the contrary, NW-trending fracture zones such as the Lower Rhine graben, which are arranged parallel to the direction of maximum compressive stress and perpendicular to the direction of least compressive stress, show mostly dip-slip dislocations of tensional type (normal faulting, Ahorner, 1975).

The borehole failure analysis of this study reveals an in situ stress regime consistent with strike-slip deformation. This is the first time that an extension of the major shear zones south of the Brabant Massif (North Artois Shear Zone and Hainaut Shear Zone) into the Aachen city area, where it joins the major normal faulting set of Roer Valley Graben zone, can be depicted based on subsurface borehole data. This major crustal shear zone was defined by (Ahorner, 1975) as an important lateral branch, which splits up the Rhenish earthquake zone somewhere between Cologne and Aachen and goes from there westward across Belgium to the Channel coast near Ostend.

The intersection of the inherited Variscan deformation grain and the Cenozoic deformation leading to recent strike-slip and normal faulting proves there exist tectonically different deformation responses at a short distance between the long-lived Brabant Massif, the Variscan Eifel Mountains and the younger Rhine Rift System. The earthquake activity in the vicinity of Aachen and the hot thermal spring activity further attest this complex structural history.

5. Conclusions

Compressive borehole breakthroughs and drilling-induced tensile fractures are well-developed and imaged in the RHWT-1 borehole. The occurrence of breakthroughs especially within two depth intervals between 1440 and 1895 m and between 2160 and 2350 m depends on lithology as the breakout width correlates to lithology dependent log properties (e.g. gamma radiation, resistivity, and velocity). The borehole failure analysis yields a mean direction of maximum horizontal stress of \(126° \pm 15°\). This SE–NW direction is in accordance with the general West European stress trend. The occurrence of near vertical induced fractures suggests that one principal stress is vertical and nearly parallel to the borehole axis.

Laboratory single stage failure tests performed on RHWT-1 core samples reveal an UCS of 117 to 336 MPa of the brittle, strongly compacted mudstones, siltsstones and sandstones. Pre-existing fractures in silstone samples which were reactivated during sample loading reduce the UCS to 40 MPa. The Mohr–Coulomb failure criterion is used to model the relation between breakout width and stress magnitudes. Stress magnitudes modelled in accordance to the Mohr–Coulomb Theory of Sliding Friction indicate minimum and maximum horizontal stress gradients of 0.019 MPa/m and 0.038 MPa/m, respectively. The stress magnitudes are consistent with a model of strike-slip deformation and provide additional evidence for the extension of the Brabant Shear Zone defined by Ahorner (1975) into the Aachen city area, where it joins the major normal faulting set of the Roer Valley Graben zone.

Acknowledgments

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References

Ute Trautwein-Bruns graduated in Geology from the RWTH Aachen University, Germany (1999), and received her PhD from TU Berlin, Germany (2005). She worked at German Research Centre for Geosciences GFZ (2000–2004). Since 2004, she has a postdoc position at the RWTH Aachen University with focus on the acquisition and interpretation of the geoscientific data set from the deep geothermal well RWTH-1. Her research interests include reservoir geology, geomechanics and petrophysics, trautwein@geo.rwth-aachen.de.

Katja C. Schulze is Technical Advisor at GeoMechanics International (GMI) where she works since 2001 on different types of consultancy projects for the oil and gas industry. She holds a Diploma (masters, 1996) in physics and a Doctorate (PhD, 2002) in geophysics from the University of Bonn, Germany. Her masters thesis deals with electromagnetic and high-resolution geoelectric measurements on a hydraulic test field. Her PhD projects involved long term fluid level recordings in the scientific boreholes KTB (Germany) and Kola Superdeep (Russia) with the focus on induced fluid level variations (tidal, barometric) and the derivation of fluid–rock properties (poroelastic and hydraulic). kschulze@geomi.com.

Stephan Becker is currently PhD-Student at the RWTH Aachen University and focused in his MSc thesis on the Aachen fold on thrust belt integrating surface geology, reflection seismic and new subsurface data from the deep geothermal well RWTH-1. In his current PhD-studies he focuses on diagenesis and reservoir quality predictions in carbonate stringers of the South Oman Salt Basin. Further research interest is the microtectonic of the RWTH-1 well. becker@geo.rwth-aachen.de.

Peter A. Kukla graduated in geology from Wuerzburg University, Germany, and Witwatersrand University, South Africa (Ph.D.). His professional career included positions at Witwatersrand University (1986–1990), Shell International E&P (1991–2000), and at RWTH Aachen University (since 2000) as full professor of geology and head of the department and director of the Geological Institute, with research focus on applied sedimentology, reservoir geology, and quantitiative geodynamics. E-mail: kukla@geo.rwth-aachen.de.

Janos I. Urai is currently a professor of structural geology, tectonics, and geomechanics at RWTH Aachen University and Inaugural Dean, Department of Applied Geoscience, German University of Technology in Oman (GUtech) in Muscat. He is interested in basic and applied aspects of rock deformation in the presence of fluids at a wide range of scales in hydrocarbon reservoirs. E-mail: j.urai@ged.rwth-aachen.de.

Lecoq, T., Petersmann, T., Alexandre, P., Cambeek, 2008. Earthquake Reslocation in the Ardenne (Belgium): Identification of Active Structures in Intraplate Context. 43(1), 75–81. In his current PhD-studies he focuses on diagenesis and reservoir quality predictions in carbonate stringers of the South Oman Salt Basin. Further research interest is the microtectonic of the RWTH-1 well. becker@geo.rwth-aachen.de.

Stephan Becker is currently PhD-Student at the RWTH Aachen University and focused in his MSc thesis on the Aachen fold on thrust belt integrating surface geology, reflection seismic and new subsurface data from the deep geothermal well RWTH-1. In his current PhD-studies he focuses on diagenesis and reservoir quality predictions in carbonate stringers of the South Oman Salt Basin. Further research interest is the microtectonic of the RWTH-1 well. becker@geo.rwth-aachen.de.
