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Reflection imaging of deep reservoir structure based on three-dimensional hodogram analysis of multicomponent microseismic waveforms

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We have applied a type of reflection technique, in which natural or induced microseismicity is used as a wave source and is processed in the time-frequency domain, to data from the Soultz Hot Dry Rock site to image deep structures within the reservoir that may be associated with flow paths among three wells. We selected and analyzed 119 good waveforms observed in 2003. The distribution of these 119 events was not strongly inclined, and although the selected number of events was much smaller than the total microseismic events located, we still succeeded in applying the reflection analysis. The reflection images obtained using waveforms from 2003 are in agreement with those obtained using data from 1993 and 2000. Finally, detailed three-dimensional reflection images were created by an integration of estimates using all available waveform data. Reflectors appear around the deep geothermal wells, and one of the dipping reflectors seems to be an obstacle to flow connection that shows a clear correlation with a microseismic cloud created during hydraulic stimulation in 2004. From these analyses we succeeded in interpreting the roles of preexisting structures in the development of flow connection among wells, which is an essential factor in a hot dry rock system.


1. Introduction

Measurement of microseismicity (which is often called acoustic emission: AE) is a commonly used tool for monitoring hydraulic stimulation of deep underground reservoirs used in hot dry rock (HDR) and hydrothermal power production, and in oil and natural gas production. Microseismic monitoring is also applied to environmental concerns such as those associated with underground storage of CO₂ and high-level radioactive waste repositories. The location and distribution of fractures, which often correlate with a flow path or flow zone, are mainly obtained as positions of microseismicity. In addition, advanced analytical techniques can obtain more detailed and important information on subsurface structures. For example, it has been shown that multiplet analysis using similar waveforms yields a detailed description of the structure [Poupinet et al., 1982; Moriya et al., 1994], the collapsing technique significantly enhances the structure description by using a stochastic approach [Jones and Stewart, 1997], and focal mechanism analysis can be used as a tool to characterize each source (i.e., fault plane and slip direction) and is useful for regional stress estimation.

The use of microseismic measurement is not limited to the investigation of the region surrounding a fractured reservoir. We have developed a technique for imaging areas surrounding microseismicity distributions by analyzing their reflected waves. The technique, called the “AE reflection method” [Soma and Niitsuma, 1997; Soma et al., 1997, 2002a, 2002b], is a reflection technique that uses natural or induced microseismicity as a wave source and records that data using downhole multicomponent seismic detectors. The original technique has been updated to include analysis in the time-frequency domain; it can delineate subsurface structures at great depth around the stimulated reservoir into which fluid is injected. The AE reflection method has an advantage over conventional methods: since both sources and detectors are located deep in the underground, it is possible to acquire high-energy wideband signals. The layout of the source and detector is such that it allows the delineation of subvertical structures. Furthermore, the additional cost of carrying out this measurement is relatively low because only a passive measurement is needed from an existing microseismic network, and the distribution of the source is over some kilometers without any obstacles from the surface conditions.

We have already successfully applied the AE reflection method to the Soultz HDR site and other geothermal fields to delineate deep reflectors [Soma et al., 1997, 2002a, 2002b]. At the Soultz HDR site, new wave data were...
obtained from large-scale stimulation tests carried out in 2003 [Baria et al., 2004]. In this study, we analyze the new data observed in 2003 by the AE reflection method, and compare and integrate them with the results of the data from 1993 and 2000. In addition to imaging the surrounding area of the reservoir, three-dimensional detailed reflectors are also visible in a smaller area around three geothermal wells at the Soultz HDR site. We also identified some dipping reflectors between the injection and production wells, and examined their influence on flow by consulting recent results from hydraulic stimulation in 2004.

2. AE Reflection Method

[5] The method has already been reported in detail in our previous paper [Soma et al., 2002a]. To achieve the use of natural or induced microseismicity as a wave source, the development of a novel signal processing technique was essential. An overview of the method that was developed is described below.

[6] In our method, we normally use sparse downhole multicomponent seismic observation stations because the target depth is usually deep or within basement rock, and surface conditions do not allow setting up many surface stations. Additionally, the attenuation of wave energy is significant at the frequency of interest. We use precise multicomponent downhole detectors to obtain accurate information from three-dimensional (3-D) particle motion for a 3-D hodogram as well as to ensure high sensitivity. The arrangement of the sources and detectors is similar to that in the reverse vertical seismic profile (VSP) method. The subsurface microseismic events (wave sources) occur at a depth where the reservoir is stimulated during fluid injection and the resultant shearing along natural joints. The shearing along natural joints generates a powerful and widely distributed passive seismic source without the need to use explosives.

[7] It is important to be able to discriminate the reflected wave from the coda train since the reflected energy can be very small [Soma and Niitsuma, 1997]. In the AE reflection method, we mainly focus on simple S to S wave reflections because we expect P wave reflections to be covered by S wave arrivals and coda, and the energy of the conversion waves is generally very small. Instead of using a simple conventional amplitude analysis, we proposed a technique using the linearity of the 3-D hodogram to detect the reflected wave apart from the coda [Soma et al., 1997]. We can use information about wave polarization from a 3-D hodogram if precise multicomponent seismic tools are used in the study. It is well known that the shape of the 3-D hodogram is spherical when an incoherent signal such as random noise is detected, but it becomes relatively linear upon the arrival of plane waves such as direct P or S waves, which are regarded as coherent signals (Figure 1). We expect a change to a relatively linear hodogram shape upon the arrival of the reflected wave if we assume that the reflected wave is a plane wave when the reflector is large enough compared to the wavelength, because coda are normally regarded as incoherent signals [Aki and Chouet, 1975]. Therefore, in the AE reflection method, we can discriminate reflected waves from coda by investigating the shapes of the 3-D hodograms.

[8] For the precise quantitative evaluation of the 3-D hodogram shape, we developed a covariance matrix method in the time-frequency domain by using a wavelet transform [Soma et al., 2002a], which is usually called a spectral matrix [Samson, 1977]. Generally, a time-frequency signal representation allows us to compose a spectral matrix in the time-frequency domain [Moriya et al., 1994]. Hence we can define a spectral matrix \( S_{WT}(b, a) \) of the 3-D hodogram by using the wavelet transform as follows:

\[
S_{WT}(b, a) = \begin{pmatrix}
W_x(b, a) & W_y(b, a) & W_z(b, a) \\
W_x^*(b, a) & W_y^*(b, a) & W_z^*(b, a) \\
W_x(b, a) & W_y(b, a) & W_z(b, a)
\end{pmatrix},
\]

(1)

where \( W_j(b, a) = W_j(b, a)W^*(b, a) \), \( a \) is the frequency (scale), \( b \) is the time (shift), \( W_j(b, a) \) is the wavelet coefficient of the \( j \)th component, and the asterisk indicates the complex conjugate.

[9] Then we can redefine Samson’s global polarization coefficient [Samson, 1977] as \( C_{p}(b, a) \), to be in the time-frequency domain as follows:

\[
C_{p}(b, a) = \frac{(\lambda_1 - \lambda_2)^2 + (\lambda_2 - \lambda_3)^2 + (\lambda_3 - \lambda_1)^2}{2(\lambda_1 + \lambda_2 + \lambda_3)^2},
\]

(2)

where \( \lambda_i = \lambda_i(b, a) \) \( (i = 1, 2, 3; \lambda_1 > \lambda_2 > \lambda_3) \) are the eigenvalues of the matrix (equation (1)) for each time (shift) and frequency (scale) \( a \).

[10] Using this parameter, we can evaluate the linearity of the 3-D hodogram quantitatively in the time-frequency domain: \( C_{p}(b, a) = 1 \) for an exactly linear hodogram and \( C_{p}(b, a) = 0 \) for a spherical hodogram. The arrival of coherent waves, such as reflected waves, should result in a high \( C_{p}(b, a) \) value. Hence we can use a high value of the parameter \( C_{p}(b, a) \) as an indicator of the arrival of reflected waves [Soma et al., 2002a]. In the study, we treat \( C_{p}(b, a) \) as a linearity waveform of a 3-D hodogram defined in the time-frequency domain.

[11] To image the subsurface structure, we established a 3-D inversion of the linearity waveform of the 3-D hodogram. The inversion concept is shown in Figure 2. This inversion, which can generate a 3-D estimate from a small number of detectors, is based on a type of diffraction stack migration technique that uses the linearity waveform \( C_{p}(b, a) \) of equation (2) for a 3-D hodogram in the time-frequency domain (Figure 2, step 2 left). In addition, we enhance the spatial resolution of estimates by restricting the virtual reflection points by examining the orthogonality between the propagation direction and the S wave polarization (Figure 2, step 2 right). The first eigenvector of the matrix \( S_{WT}(b, a) \) (equation (1)) is regarded as the S wave polarization direction in the time-frequency domain, and the propagation direction is represented by the vector geometrically defined by both observation and virtual reflection point. Therefore the orthogonality can be examined by analyzing inner products of the two vectors for each virtual reflection point. In the conventional processing, the strength of the linearity waveforms of the 3-D hodogram is projected on all the isodelay ellipsoid which is the distribution of virtual reflection points. Here, the distribution of the virtual reflection point can be reduced if we select the area only
where S wave polarization is perpendicular to the propagation direction from the reflected point because we focus on S wave reflections. Furthermore, to reduce the effect of a heterogeneous source distribution we normalize the event density for a number of nearby events for each source as a weight in the inversion process. This normalization helps to avoid having a dominant ellipsoidal artifact in the final image [Soma and Niitsuma, 1997]. These two operations are effective in obtaining more reliable images when a natural distribution of induced seismicity is used in the inversion.

3. Site Description and Waveforms

3.1. European Hot Dry Rock Site at Soultz-sous-Forêts

The European deep geothermal energy program uses HDR technology and has been operating at Soultz-sous-Forêts, Alsace, northeastern France (Figure 3), since 1987, supported predominantly by the EU, France, and Germany [Baria et al., 1995]. There are three deep geothermal wells (GPK2, 3, and 4), a previous shallow geothermal well (GPK1), an exploration well (EPS1), and four observation wells (4550, 4601, 4616, and OPS4) in which downhole microseismic stations can be placed (Figure 4). An initial large-scale hydraulic stimulation was made in well GPK1 in 1993 and an artificial geothermal reservoir was created at a depth of around 3500 m. During 1994–1995, a second geothermal well (GPK2) was drilled into the reservoir to a depth of around 3800 m and subsequently stimulated, which resulted in the successful development of a circulating system. In summer and autumn 1997, a 4-month long-term circulation test was performed that successfully demonstrated that this technology had a strong future potential as an energy source [Baumgartner et al., 1998].

The reservoir created before 1996 was at 3600 m depth and was regarded as a scientific pilot plant that laid the basis for a nearby commercial system to be developed at around 5000 m depth. By 2004, an underground heat exchanger with three deep geothermal wells had been developed. The GPK2 well was deepened to a depth of 5 km in 1999, and a temperature of 200°C was obtained [Baria et al., 2000]. Then a stimulation of GPK2 was carried out in June–July 2000. Drilling of the central well GPK3 was conducted in 2002, and GPK3 was stimulated in 2003. A third deep well GPK4 was drilled in 2004, and GPK4 was stimulated in autumn 2004 [Baria et al., 2005]. The program is now in its second stage and concentrates on characterization of the long-term behavior of the underground heat exchanger and on installation of a power plant on the surface [Baumgärtner et al., 2005].
[1] 3C AE waveform

\[
\text{Calculate the matrix from eq. (1)}
\]

\[
S_{WT}(b,a) =
\begin{pmatrix}
W_{xx}(b,a) & W_{xy}(b,a) & W_{xz}(b,a) \\
W_{yx}(b,a) & W_{yy}(b,a) & W_{yz}(b,a) \\
W_{zx}(b,a) & W_{zy}(b,a) & W_{zz}(b,a)
\end{pmatrix}
\]

Obtain 3 eigenvalues and 3 eigenvectors

[2] Linearity of 3D hodogram

[3] Strength of hodogram linearity within the window for scale \( a_j \) and \( \Delta T \)

\[
\text{sum}Cp(a_j, \Delta T) = \sum_{m=-\text{win}/2}^{\text{win}/2} Cp(b_j + \Delta T + m, a_j)
\]

Iso-delay band for \( \Delta T \) restricted by S-wave polarization

Detector

S

Source

N

E

W

Depth

\( L_r(\Delta T) \): path length for delay "\( \Delta T "\"

A virtual reflection point

S-wave polarization for \( (b_j + \Delta T, a_j) \)

Information about wave polarization direction


Then, spatially integrate all iso-delay bands weighted by the strength of hodogram linearity for all source AE events.

Delineate positions of gathering higher linearity of 3D hodogram as reflectors.

**Figure 2.** Schematic of the concept of 3-D inversion using the linearity of 3-D hodograms.
3.2. Wave Data for the AE Reflection Method

[14] Passive microseismic monitoring has been conducted during almost all the hydraulic stimulations at the Soultz HDR site since the beginning of its development. In this study, we focus only on data sets of the large-scale hydraulic stimulations conducted in 1993, 2000, and 2003 since these provide more than enough good multicomponent waveforms for the analyses. To apply the AE reflection method, we need precise three-dimensional particle motion data that are not corrupted by such things as electrical or tool resonance problems. We generally need more than 100 events recorded at least one location with good quality waveforms. Therefore it is advantageous to use multicomponent downhole seismic detectors (four-component detector with tetrahedral sensor configuration [Jones and Asanuma, 2004]), overall sensitivity of 1000 V/g, and noise level around 10 μg (R. Jones and H. Asanuma, personal communications, 2006) because of their high signal-to-noise ratio and low attenuation. The four-component waveforms are converted to three-component before analysis to obtain the three-dimensional particle motion for a 3-D hodogram (see Figure 1).

[15] To carry out the AE reflection method on the data from 2003, we selected 119 waveforms observed at well 4550 from among about 10000 located events. The detector was set at a depth of 1482.0 m in 2003. For the other detectors in wells 4601, 4616, and OPS4, the stability of the three-dimensional particle motion was not as good as that at well 4550 due to electrical problems, and data from other wells were not used. The distribution of the 119 selected events was not strongly inclined in comparison to the whole observed microseismicity (see Figures 5 and 6), and was reasonably usable for the AE reflection method.

[16] At the Soultz HDR site, the dominant waveform frequency is usually around 100–200 Hz. Since the geological setting is relatively simple (inside a granite basement), the arrival of P or S waves is not difficult to identify on the waveform (see Figure 1). Positions of the microseismic sources and the uniform velocity model were taken from past studies of the Soultz HDR site [Dyer et al., 1994; Asanuma et al., 2001, 2004].

4. Imaging

4.1. Analysis of Waveforms From 2003 and Comparison With Past Results

[17] We applied the AE reflection method to 119 waveforms from well 4550 that were observed in 2003. The results from data taken in 1993 and 2000 have been already reported [Soma et al., 1997, 2002a, 2002b].

[18] Examples of imaging of data from 2003 are shown in Figure 7. The images are of south–north and west–east cross sections though the wellhead of GPK1, and a west–east cross section 1000 m south of GPK1. The thick lines show the well trajectory projections and the small dots show the microseismicity distributions in 1993, 2000, and 2003. We delineate high values of stacked hodogram linearity (\(C(p, a)\) in equation (2)), which are regarded as having higher reflectivity in the analysis, by darker colored structures. The shapes and positions of reflectors are normally strongly affected by the mutual relationship between the distribution of source events and the locations of the detectors. Attenuation compensation about depth (i.e., propagation distance) was not applied when determining the reflectivity: this overemphasized the shallow reflectors and underemphasized the deep reflectors. However, it is clear that relative changes and reflectivity anomalies correspond to structural discontinuities as we confirmed in our past study by comparison between reflectivity anomalies along the well trajectories and borehole images using FMI (Formation Micro Imager; based on resistivity) and BHTV (Bore Hole Tele Viewer; based on resistivity)
Soma et al., 2002a]. In the west–east cross section (Figure 7b), relatively high reflectivities produce dipping reflectors between the depths of 3900 and 6300 m. In the north–south cross section (Figure 7a), reflectivity appears from shallow depths to a depth of 5500 m. The steeply dipping reflectors around a depth of 3500 m may correspond to a part of the regional subvertical fault system reported in an earlier study using data from 2000 [Soma et al., 2002b].

We compared the results shown in Figure 7 with previous results from data recorded in 1993 (Figure 8) [Soma et al., 2002a] and 2000 (Figure 9) [Soma et al., 2002b]. In Figure 8, 101 waveforms from well 4550 (detector depth 1483.5 m) in 1993 that occurred in the shallow artificial geothermal system (3.5 km depth range) were used for the analysis. In Figure 9, 257 deep events observed in wells 4601 (detector depth 1539.0 m) and OPS4 (detector depth 1484.7 m) in 2000 were analyzed. As shown in Figures 7 and 8, reflectors with low dip angles are detected from analyses of waveforms in both 2003 and 1993. Although the positions of the cross sections and the shapes of the reflectors are not exactly the same because of the different observation conditions, some of the reflectors, particularly below 4500 m, do resemble each other, for example at the depth of 4800 m and 5300 m in the west–east cross sections.

On the other hand, dipping reflectors at the depths of 3900–4100 m and 4200–4600 m in the west–east cross sections in Figures 7b and 7c (using waveforms from 2003) can be associated with dipping reflectors using waveforms obtained in 2000 from wells 4601 and OPS4 (Figure 9). For the south–north cross sections, similarities of the results using data from 2003 to previous results are also revealed in the comparison. Therefore we concluded that the data from 2003 roughly follow trends similar to previous analyses in 1993 and 2000 at the Soultz site, although the precise positions and the reflector shapes depend on the individual observation conditions. Hence we can reasonably integrate over all the results for a more detailed interpretation of the reflection images.

4.2. Integrated 3-D Reflected Image of the Deep Artificial Reservoir

We attempted to image 3-D structures around and inside the artificial geothermal reservoir at the Soultz HDR
Figure 6. Distribution of source events used in the AE reflection method. Light gray diamonds, dark gray squares, and white circles are selected source events in 1993, 2000, and 2003, respectively. Triangles are positions of seismic detectors in 2003, and thick lines indicate the trajectories of geothermal wells.

Figure 7. Examples of reflection images by the AE reflection method using waveforms at station 4550 (depth of detector: 1484 m) in 2003. (a) North–south cross section through GPK1, (b) west–east cross sections though GPK1, and (c) cross section at 1000 m south from GPK1. The black contours indicate higher reflectivity. The light gray, gray, and black dots are the projections of the source locations in 1993, 2000, and 2003, respectively, within ±100 m from each cross section. Dipping reflectors bounded by dotted lines are examples of similarities with the past analysis shown in Figure 9.
site by reanalyzing the waveforms from 1993, 2000, and 2003, and integrating all the results of the AE reflection method. For the integration process, we first independently analyzed the waveforms for each year and adjusted both indicated data range and classification for creating each contour map. This was done because each observation condition produced a different reflectivity magnitude (stacked hodogram linearity) for each reflection image. Then each estimate was normalized and all were integrated by a logical summation (“OR”). By this operation, maximum value among normalized reflectivities calculated from different data sets is adopted as a reflectivity for each 3-D spatial point during the integration process. To obtain a high resolution around the reservoir, the calculation, normalization, and integration of the AE reflection method were done within each cubic block with sides 500 m long. This was done because of the limitation of our CPU power. The grid size of the calculation in each cubic block was set as $10 \times 10 \times 10$ m, which is the theoretical spatial resolution of the analysis, although the dominant quarter wavelength for the frequency range of the wavelet transforms in this study was defined to be about 8.5 m (for an S wave velocity of 3400 m/s assuming an apparent peak frequency of 100 Hz in the wavelet transform).

The 3-D image created by the AE reflection method of the region around the deep artificial reservoir is shown in Figure 10. In Figure 10, small dots indicate all the microseismic events from 2000 and 2003, while the thick lines are the trajectories of the geothermal wells, GPK2, 3, and 4. The gray clouds are isosurfaces of high reflectivity, which systematically compose 3-D reflectors. The result of the processing (Figure 10) shows that the estimated reflectors are located mainly around the microseismic cloud, which corresponds to the artificially created reservoir, since the method is not so effective for delineating structures in the area of the source itself. The images around the sides of

![Figure 8](image1.png)

**Figure 8.** Reflection images using waveforms from 1993 at station 4550 (depth of detector: 1483.5 m) by the AE reflection method (modified from Soma et al. [2002a]). (a) West–east cross section at 400 m north from GPK1 and (b) North–south cross section at 300 m east. Triangles and circles show the projections of the detectors and sources, respectively.

![Figure 9](image2.png)

**Figure 9.** Reflection images using waveforms from 2000 at both stations 4601 (depth of detector 1539 m) and OPS4 (depth of detector 1484.7 m) by the AE reflection method [Soma et al., 2002b]. West–east cross sections at distances of (a) 600 m north, (b) 400 m north, (c) 200 m north, and (d) 0 m from GPK1. Gray and black dots indicate projections of the AE source distributions from 1993 to 2000 within a distance of $\pm 200$ m from each cross section. Dipping reflectors bounded by dotted lines are examples of similarities with the recent analysis shown in Figure 7.
the overall microseismic cloud show a highly inclined cluster of reflectors. A subhorizontal cluster of reflectors is also seen at the bottom of the cloud. These relate to the outer boundary of the artificial reservoir because of the contrast among the densities of the inflated fractures, although the estimated extent of the reflectors is influenced by the observation conditions. Reflectors can also be seen between the wells, which can assist in the characterization of the heat exchanger.

[23] Figures 11 and 12 are magnifications and readjustments of the reflection images between geothermal wells GPK2 and 3, and between wells GPK3 and 4, respectively. In Figure 11, there are reflectors around the depths of 4000–4500, 4500–4700, 4800–4900, 5000–5100, and 5300–5500 m. These reflectors indicate possible structures inside the stimulated reservoir that we infer to be related to flows in the HDR system, although the extent of the reflectors is limited by the illumination conditions (position of source and detectors) for the reflector. The deep subhorizontal reflectors at a depth of 5300 m below the microseismic cloud might be related to the bottom boundary of the artificial reservoir. In Figure 12, there are also significant reflectors between and below GPK3 and 4 mainly at depths of 4000–5000 and 4900–5500 m. The shallow reflector indicates a higher dip angle, and the deeper one

Figure 10. Three-dimensional integrated reflection images by the AE reflection method around the deep reservoir. (a) View from N225E and (b) from N115E. Isosurfaces shaded from gray to black indicate reflection images, thick lines are trajectories of geothermal wells, and small dots show all location of microseismicity from 1993 to 2003.

Figure 11. Three-dimensional integrated reflection images around the wells GPK2 and GPK3 by the AE reflection method. (a) View from N225E and (b) from N115E. Isosurfaces shaded from gray to black indicate reflection images, thick lines are trajectories of GPK2 and GPK3, and small dots show all locations of microseismicity from 1993 to 2003.
a medium to low dip angle, consistent with the bottom of a stimulated reservoir.

5. Discussion

[24] We have shown that the AE reflection method can identify deep structures as 3-D reflectors between wells inside basement rock. It is important to investigate the characteristics of the reflectors because deep structures could relate to flow communication between wells, which is essential for understanding the performance of underground heat energy extraction. Here, we discuss the meaning of the reflectors, influences on flow communication, and the contribution of the AE reflection method.

[25] In our previous study, detected high reflectivity showed a correlation with fractures or fractured zones as identified by well logging data [Soma et al., 2002a]. It is reasonable for the AE reflection method to detect such thin structures because the analyses are based on S wave reflection. Therefore we can suppose that the detected spatial reflectors also show structural discontinuities such as fault systems, fracture zones, and contrasting densities of fracture in the rock mass. The shallow fault system around the Soultz HDR site is shown to be highly dipping or subvertical in an interpretation of a past reflection survey (Figure 13) [Cautru, 1986], although the conventional method cannot see inside the basement rock. Furthermore, from geological observations of rock core samples, it has also been reported that principal joint sets have high dip angles (65–70°) in the deep basement [Genter and Dezayes, 1993]. One possible interpretation of the detected subvertical reflectors is as indications of preexisting fractures, fracture zones, or joint sets.

[26] A recent geological study reported that there are variations in the type of granite in the basement rock, although a rather uniform rock matrix has been imaged in studies since work commenced at the Soultz HDR site. The following geological units have been confirmed in the basement: standard porphyritic K-feldspar-megacryst granite (to a depth around 2700 m), the same granite with intense vein alteration (to a depth around 3900 m), biotite- and amphibole-rich granite becoming gradually porphyritic granite (to a depth around 4700 m), and fine-grained two-mica granite below a depth of about 4700 m [Dezayes et al., 2005]. These boundaries can cause physical property changes due to rock types and fracture densities by alteration. These are not only subvertical but also have medium dip angles. We see a relationship between the upper bound-

Figure 12. Three-dimensional integrated reflection images around the wells GPK3 and GPK4 by the AE reflection method. (a) View from N225E and (b) from N115E. Isosurfaces shaded from gray to black indicate reflection images, thick lines are trajectories of GPK3 and GPK4, and small dots show all locations of microseismicity from 1993 to 2003.

Figure 13. An example of interpretation of past reflection surveys at the Soultz HDR site [Cautru, 1986].
Figure 14. Time-space microseismic distribution around the deep reservoir. Three different data sets are included: events during injection from GPK2 in 2000 (white circles), from GPK3 in 2003 (black inverted deltas), and from GPK4 in 2004 (white triangles). Percentage indicates the ratio of microseismicity for each data set.
ary of the fine-grained two-mica granite (around a depth of 4700 m) and medium to low dipping reflectors shown in Figures 8 and 11. Furthermore, the existence of an intruded dike at great depth has also been reported. A large fracture zone is observed in GPK3 at a depth of about 4756 m (which is measured well depth, not true vertical depth), and it is hypothesized to represent the weak boundary between the standard granite and the two-mica granite. As mentioned above, another possible interpretation of the detected reflector is that it represents a change in the composition of the granite basement rock.

In hydraulic fracturing, it is well known that preexisting structure strongly influences the development of hydraulic connections between wells. Good hydraulic flow communication between wells GPK2 and 3 was reported after the GPK3 stimulation [Baumgartner et al., 2005]. High reflectivity as determined by the AE reflection method in GPK2 corresponded to fracture zones around the depths of 4590 m and 4775 m in our previous study [Soma et al., 2002a], and the depth of total water losses observed during drilling GPK3 (which are 4757 and 5091 m in measured depth [Hettkamp et al., 2004]) were close to reflectors around depths of about 4700 and 5000 m along GPK3. These high reflectivities can be inferred to be inlets of flow zones, although it is difficult to delineate the exact flow path between the two wells. The top and bottom reflectors can be associated with the boundaries of the fractured reservoir area. The distribution of detected reflectors is in agreement with flow zones between GPK2 and 3 although we cannot see the major flow path in the fractured reservoir. In the time-space microseismic distribution (Figure 14), the initial microseismic events occurred around a depth of about 4700 m in both GPK2 and 3, near the depths of small reflectors such as in Figure 11, which could be the main inlets for injections. The expansion of microseismicity is nearly limited to between the depths of the top and bottom reflectors, which indicates the flow zone.

On the other hand, it has been reported that between wells GPK3 and 4 no acceptable hydraulic connection between the two wells was achieved even after the stimulation of GPK4 in September 2004 [Baria et al., 2005], although our estimates produced a relatively large reflector between GPK3 and 4 (Figure 12). We speculate that the significant structure detected as the dipping reflector does not function like a flow path from GPK4 because it may be an intruded dike and is therefore an obstacle to flow between GPK3 and 4. Microseismicity from GPK4 associated with the stimulation in 2004 has been reported to have occurred mainly around the deep section of GPK4 below a depth of about 4500 m [Baria et al., 2005]. A flow log carried out in GPK4 during the stimulation showed a major flow exiting near the bottom of the well. The enhanced flow zone was created at the bottom of GPK4 although the stimulation had to be stopped after 3 d of pumping because of accessibility problems with the production-logging tool in the well. Figure 15 shows a comparison of the detected reflectors between wells GPK3 and 4 with the microseismic cloud generated during the stimulation of GPK4. The deep reflector detected at a depth of 5000–5500 m (see Figure 12) corresponds with the center of the microseismic cloud generated during stimulation of GPK4. The deep reflector detected at a depth of 5000–5500 m (see Figure 12) corresponds with the center of the microseismic cloud generated during stimulation of GPK4 (Figure 15). The initial microseismic events from GPK4 around a depth of 5000 m (see Figure 14) are at a depth similar to that of the deep reflector. If the reflector relates to a preexisting structure, such as a fractured zone, the structure possibly helped hydraulic fluid to enter into the rock mass. On the other hand, the shallower dipping reflector at a depth of 4000–5000 m between GPK3 and 4 has a similar shape and is in agreement with the upper boundary of the microseismic cloud and seems to be like a cover (Figure 15).

Considering that the flow connection between GPK3 and 4 is poor, we propose the following scenario to explain the poor connection. Fluid enters initially along the deep preexisting structure, and then it expands in the rock matrix, which may have small-scale fractures. Eventually, the fluid
encounters some large-scale preexisting structure that inhibits the flow across to the second well. Geological investigations have also shown that there are intruded dikes [Dezayes et al., 2005] that may inhibit the formation of a hydraulically conductive zone. An alternative scenario would be that a highly stressed barrier was created between the two wells by the injection process in each well and that this may act like a barrier to flow. It is also possible that the unfortunately short 3 d injection from GPK4 in 2004 was not enough to break or cross the barrier. [30] Up until now, it has been very difficult to obtain spatial information, other than microseismic locations, from inside deep basement rock distant from wells. The AE reflection method offers 3-D spatial information that does not depend on well positions, and it can contribute to the interpretation of flow circulation development as described in this paper.

6. Conclusion

[31] We have applied a reflection technique using natural or induced microseismicity as a wave source, which we call the AE reflection method, to waveforms from hydraulic stimulation tests at the Soultz HDR site. The waveforms observed in 2003 were analyzed, and the results were compared and integrated with those from 1993 and 2000 data. Detailed 3-D reflectors were extracted around three geothermal wells, and we compared them with the most recent microseismicity distributions obtained during the stimulation of well GPK4 in 2004.

[32] The AE reflection method is based on our analysis of the linearity of 3-D hodograms, rather than the standard reflection technique that uses the reflected wave amplitude. The linearity of a 3-D hodogram is evaluated in the time-frequency domain, which is regarded as reflectivity in the analysis. Then the linearity waveform in the time-frequency domain is projected into 3-D space and all the wave data are integrated to obtain a 3-D subsurface image, which is a process similar to diffraction stack migration.

[33] Analysis of data from 2003 was recently conducted using 119 good waveforms from well 4550, which had good 3-D particle motion. The distribution was not strongly biased, and even though 119 events was a much smaller number than the total number of waveforms detected in 2003, we still succeeded in applying the AE reflection method. The reflectors around and below the artificial geothermal reservoir were detected in a depth range of 3500 to 6300 m, and the shape of reflectors indicate highly dipping to subhorizontal structures with similar trend to past analyses.

[34] We also created 3-D integrated reflection images with higher spatial resolution using all estimates from 1993, 2000, and 2003, taking into consideration the peripheral area around the three geothermal wells, GPK2, 3, and 4. Reflectors were detected not only surrounding the artificial reservoir but also inside the reservoir between geothermal wells. The dipping reflectors were mainly located between the geothermal wells, and subhorizontal reflectors were related to the bottom of the artificial reservoir. Some deep reflectors near wells seem to relate to enhancing or creating a flow zone considering expansion process of microseismic clouds. On the other hand, the highly dipping reflector between GPK3 and 4 has a similar shape to the upper boundary of the microseismic clouds observed during GPK4 stimulation in 2004, and the reflector possibly shows a barrier to flow which inhibited the flow between the wells during the stimulation.

[35] We conclude that the AE reflection method can provide useful information for understanding, characterizing, and evaluating a hydraulic stimulation by detecting preexisting structures from areas where the application of conventional methods is often not just difficult, but is impossible, such as within deep basement rocks.

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