Hydraulic properties of the crystalline basement

Ingrid Stober · Kurt Bucher

Abstract Hydraulic tests in boreholes, up to 4.5 km deep, drilled into continental crystalline basement revealed hydraulic conductivity (*K*) values that range over nine log-units from 10^{-13}–10^{-4} m s^{-1}. However, *K* values for fractured basement to about 1 km depth are typically restricted to the range from 10^{-8} to 10^{-6} m s^{-1}. New data from an extended injection test at the KTB research site (part of the Continental Deep Drilling Program in Germany) at 4 km depth provide K=5 10^{-8} m s^{-1}. The summarized *K*-data show a very strong dependence on lithology and on the local deformation history of a particular area. In highly fractured regions, granite tends to be more pervious than gneiss. The fracture porosity is generally saturated with Na–Cl or Ca–Na–Cl type waters with salinities ranging from <1 to >100 g L^{-1}. The basement permeability is well within the conditions for advective fluid and heat transport. Consequently, fluid pressure is hydrostatic and a Darcy flow mechanism is possible to a great depth. Topography-related hydraulic gradients in moderately conductive basement may result in characteristic advective flow rates of up to 100 L a^{-1} m^{-2} and lead to significant advective heat and solute transfer in the upper brittle crust.

Résumé Les tests hydrauliques en forages de plus de 4.5 km de profondeur, forés dans le socle continental cristallin révèle des valeurs de conductivités hydrauliques (*K*) qui s'étendent sur neuf unités log, de 10^{-13} à 10^{-4} m s^{-1}. Néanmoins les valeurs de *K* du socle fracturé jusqu'à 1 km de profondeur sont typiquement restreintes de 10^{-8} à 10^{-6} m s^{-1}. De nouvelles données provenant d'un test d'injection au site de recherche du KTB (faisant parti du Programme de Forages Profonds en domaine Continental en Allemagne) à 4 km de profondeur, apportent un *K* de 5 10^{-8} m s^{-1}. Les données de *K* possèdent une dépendance forte avec la lithologie et l'histoire des déformations locales d'une région particulière. Dans les régions très fracturées, le granite tend à être plus perméable que le gneiss. La porosité de fracture est généralement saturée avec des faciès hydrochimiques de type Na–Cl ou Ca–Na–Cl et des salinités comprises entre <1 et >100 g L^{-1}. La perméabilité de socle est bien comprise dans les conditions de transport du fluide advectif et de la chaleur. Par conséquence, la pression de fluide est hydrostatique et le mécanisme d'écoulement de Darcy est possible à de plus grandes profondeurs. Le relief des gradients hydrauliques dans le socle modérément conductif pourrait résulté dans des taux d'écoulement advectifs caractéristiques de plus de 100 L a^{-1} m^{-2} et conduit à des transferts significatifs de chaleur et de solutés dans la croûte supérieure cassante.

Resumen Ensayos de bombeo realizados en sondeos de hasta 4.5 km de profundidad, excavados en basamento cristalino, arrojan valores de conductividad hidráulica (*K*) que varían en más de nueve órdenes de magnitud, desde 10^{-13}–10^{-4} m s^{-1}. Nuevos datos procedentes de un test de inyección extendido en el punto de investigación KTB (parte del Programa "Continental Deep Drilling" en Alemania) a una profundidad de 4 km arroja una K=5 10^{-8} m s^{-1}. Los datos de *K* resumidos muestran una dependencia muy fuerte de la litología y de la historia de las deformaciones locales en un área particular. En regiones fuertemente fracturadas, los granitos tienden a ser más permeables que los gneiss. La porosidad por fracturación está generalmente saturada con aguas de tipo Na–Cl o Ca–Na–Cl cuyas salinidades oscilan entre <1 a >100 g L^{-1}. La permeabilidad del basamento es acorde para las condiciones para transporte advectivo de fluidos y calor. Consecuentemente, la presión de fluidos es hidrostática y el mecanismo de flujo de Darcy es posible a grandes profundidades. Gradientes hidráulicos relacionados con la topografía en basamentos moderadamente conductivos pueden ser consecuencia de rangos de flujo advectivo característicos de hasta 100 L a^{-1} m^{-2} y apuntan hacia la existencia de calor advectivo significativo y transferencia de solutos en la parte superior de la corteza frágil.

Keywords Hydraulic properties · Hydraulic testing · Fractured rocks · Crystalline rocks · Deep water

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I. Stober (✉)
Geological Survey B.-W.,
Albertstr. 5, 79104 Freiburg, Germany
e-mail: stober@lgrb.uni-freiburg.de

K. Bucher
Institute of Mineralogy and Geochemistry, University of Freiburg,
Albertstr. 23b, 79104 Freiburg, Germany
e-mail: bucher@uni-freiburg.de

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Introduction

The upper crust of the Earth's continents is, in contrast to earlier belief, not "dry." A distinctive mobile aqueous fluid phase, water, has been found in the fracture porosity of the crystalline basement to a great depth. Data, evidence and observations from deep mines and from scientific drilling down to several thousand meters are supplemented with rapidly accumulating new data from the exploration of geothermal energy sources (Fritz and Frape 1983; May et al. 1996; Nordstrom et al. 1985; Pauwels et al. 1993; Schmassmann et al. 1984; Kozlovsky 1984; Norton and Knapp 1977; Banks et al. 1996; Emmermann et al. 1995; Stober and Bucher 2000, 2005a).

The crystalline basement, mostly granite and gneiss, normally contains interconnected fracture porosity with a relatively high hydraulic conductivity. The brittle upper crust can be regarded as an aquifer in hydraulic terms. Highly mineralized salty water or brine is typically present in the fracture pore system (Möller et al. 1997; Stober 1995; Edmunds and Savage 1991; Behr and Gerler 1987; Edmunds et al. 1985; Kanz 1987; Pearson et al. 1989; Carden et al. 1985; Benderitter and Elsass 1995). The salty fluids in the upper brittle part of the continental crust have thus characterized the crystalline basement to the depth of the deepest boreholes drilled to date: 12.5 km at Kola Peninsula, Russia (Kozlovsky 1984) and 9.1 km at the Continental Deep Drilling (KTB) site, Germany (Emmermann et al. 1995).

In the brittle continental upper crust, interconnected water-conducting features, e.g., zones with enhanced hydraulic conductivity, result from brittle deformation and determine the hydraulic heterogeneity of the rocks. Water conducting structures generated by brittle deformation mechanisms such as faults, joints or veins (example given in Fig. 1) represent the dominant flow porosity (e.g., Mazurek 2000; CFCCF 1996; Caine et al. 1996).

Since the fracured crystalline basement contains water (fluid if supercritical) to a great depth, the hydraulic properties of basement rocks determine fluid flow patterns in the brittle crust. Fluid pressure conditions are near-hydrostatic in this regime and conditions for Darcy flow are generally given. Advective fluid transport is controlled by the permeability and porosity structure of the crust. Furthermore, the hydraulic properties of the basement are intimately coupled to the chemical composition of the fluids. Clauser (1992) and Manning and Ingebritsen (1999) presented comprehensive data compilations of hydraulic properties of the crust. Recently, important new data from hydraulic tests in deep boreholes have become available (Stober and Bucher 2004, 2005b). The new hydraulic data help to improve the understanding of deep-water conditions in the continental crust. In addition to the previously reported hydraulic data from deep drill holes, this paper presents a number of previously unpublished data from older hydraulic tests in deep drill holes and new data from an extraordinary injection test in the 4-km deep pilot hole at the KTB test site. Together with a data review and compilation from hydraulic tests of deep boreholes into the crystalline basement worldwide, the new data provide a refined picture of the hydraulic properties of the continental crust to about 5 km depth.

Fig. 1 Fractured gneiss outcrop near Lom, Norway, with water flows from water-conducting fractures. Interconnected fracture network with fracture density of 4 m^{-1}. Scale: *hammer*

Hydraulic properties of the crust, assessment, and parameters

The prime parameters controlling advective fluid, solute, and heat transport in fractured crystalline rocks of the crust are permeability κ and porosity ϕ. Depending on the scale of interest, different methods are used to determine the two parameters. These include laboratory measurements (e.g., Berckhemer et al. 1997), fracture analysis (CFCCF 1996; Mazurek et al. 2003; Jakob et al. 2003; Caine et al. 1996; Konzuk and Kueper 2004), geophysical modeling of heat flow data (e.g., Hayba and Ingebritsen 1994) and in-situ testing of boreholes (e.g., Stober 1995).

For this communication, reported data are summarized and new porosity and hydraulic conductivity data of fractured continental upper crust that have been derived from hydraulic testing of deep boreholes are presented. The test sections in the deep boreholes differ from 1 to 1,000 m. Most tested open holes or packer sections were from 100 m to several hundred m and the presented data are thus characteristic for the vertical scale >100 m–2 km. There are, to the authors' knowledge, only two boreholes in basement rocks with tested depths of 4,000 m and more. These are the Urach 3 geothermal well, also in Germany, (Stober and Bucher 2004) and the KTB well (Stober and Bucher 2005b). Several boreholes have been tested in the depth range of 2–4 km. At shallower depth, the database is rapidly increasing with decreasing depth. The total available information permits a relatively confident statement on the porosity and hydraulic conductivity of the basement to 1 km below surface with limited data at greater depth.

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Much of the insight on the hydraulic properties of the basement has been provided by the research on radioactive waste repositories. Particularly, the Swiss nuclear waste cooperative (NAGRA), the Stripa Project of Sweden, the projects at the Äspö Test Site in Sweden and the POSIVA Project of Finland have contributed a wealth of information on basement hydraulics.

Hydraulic properties of fractured crystalline basement that are required for fluid flow modeling in fractured crust include the following parameters. Hydraulic conductivity K (m s^{-1}) is the proportionality constant in the Darcy flow law. It can be retrieved from transmissivity data T (m^2 s^{-1}) derived from hydraulic tests in boreholes. T relates the hydraulic conductivity to the tested aquifer thickness b (m) and K is then given by T/b. The K values characterize the hydraulic conductivity of a combined system of rock matrix and interconnected fracture network. The permeability κ (m^2) is an intrinsic fluid-independent system property and can be computed from the hydraulic conductivity: κ=K μ/g ρ with g = acceleration due to gravity, ρ = density and μ = the dynamic viscosity of the fluid (water). The ratio ξ=μ/g ρ is about 1 10^{-7} at 20°C and 0.2 10^{-7} at 150°C, hence it varies for pure water with less than a factor of 5 for the pressure and temperature range of interest. For the conditions at 4 km depth, which are of particular interest here, the numerical value of the permeability κ is about K×0.3 10^{-7} (e.g., K=10^{-8} m s^{-1} ; κ~3 10^{-16} m^2). The ratio ξ also depends on the chemical composition of the fluid; however, this variation is small compared to the temperature effect. Nevertheless, it is important to note, that the factor converting permeability to hydraulic conductivity ω=κ/K is not a constant and the given value ω=0.3 10^{-7} m s permits the conversion on an order of magnitude basis only. The porosity ϕ (%) is defined as the volume of voids per total volume expressed in % (or m^3 m^{-3}). The effective fracture porosity ϕ_f of a rock unit is determined by the frequency, shape, and average aperture of the fractures. ϕ_f determines how much mobile water or fluid is stored in the fractures and this amount enters model calculations of water–rock interaction processes and fluid flow. These processes have the ability to deposit or dissolve minerals on the fracture walls, thus gradually changing porosity and permeability of the rock. The rock matrix is characterized by the matrix porosity ϕ_m. It also may contribute to the effective advective flow porosity ϕ_a of the total system, matrix, and fractures. The flow effective portion of the rock matrix porosity is generally low compared to the fracture porosity ϕ_f so that ϕ_f~ϕ_a.

The aperture of fractures in basement rocks can be measured, estimated, and modeled by a wide variety of approaches and techniques (CFCCF 1996). Typical derived aperture values for rough fractures are about 0.3 mm (Konzuk and Kueper 2004) leading to very low porosities of fractured crystalline rock aquifers. Alternatively, small fracture apertures require implausibly high fracture densities to match porosities measured in well tests (>30 m^{-1}). A range of values of 1–3 mm for apertures of flow relevant fractures would be consistent with observed fracture densities and measured in-situ porosities (e.g., Mazurek et al. 2003). Representative fracture density of basement ranges from one to ten fractures m^{-1} with the most prominent value of two to three fractures m^{-1} (Fig. 1), thus an average fracture porosity ϕ_f~0.6% follows from, e.g., three fractures m^{-1} and an aperture of 2 mm.

The flow effective porosity of the Black Forest basement varies from 0.1 to 2.1% (Stober 1995). These data have been exclusively derived from measured storativities S in well tests via the equation:

$$\phi_a = S/(\rho\, g\, c_w\, b) \tag{1}$$

where c_w represents the compressibility of water at the conditions of interest. These ϕ_a are very close to ϕ_f. Reported porosity data for granite 0.3–1.0% (Guillot et al. 2000; Mazurek et al. 2003) are within this range. Aquilina et al. (2004) measured a range of 0.9–2.3% for the flow porosity of granitic crust at 3.5–3.7 km depth from the deep geothermal well at Soultz-sous-Fôrets, France. In-situ porosity data for gneiss and amphibolite are scarce; it is likely that porosity values for these are generally lower. If the fluid in the pore space is not in chemical equilibrium with mineral assemblage of the rock matrix, water–rock interaction may increase the porosity with time by dissolution reactions, or conversely it may decrease with time due to precipitation reactions.

Hydraulic conductivity and porosity, in-situ data

Summary of case histories

Äspö hard rock laboratory in Sweden

The Äspö Hard Rock Laboratory is located on a coastal island south of Stockholm, Sweden (Stanfors et al. 1999). The laboratory is located in an excavation 450 m below the island and deep boreholes have been drilled from there to 1,700 m (bottom hole 2,150 m below surface). The basement mainly consists of granitoids of the Trans-Scandinavian Igneous Belt (age 1.85 Ga) with minor components of mafic sheets, dikes and xenoliths. The site also contains shear and fracture zones and fine-grained granites with distinctive fracture porosity. Injection tests with 3-m packer intervals resulted in derived hydraulic conductivities K=3 10^{-10} to 1 10^{-8} m s^{-1}. No clear decrease with depth could be observed at Äspö in contrast to other sites in Sweden (Walker et al. 1997). This is probably because the uppermost, generally highly fractured or altered part of the basement has not been included in the analysis in contrast to other sites. Four sets with different hydraulically conductive fractures can be distinguished. Fault gouge-filled fractures were found to be the most conductive ones (Mazurek et al. 2003). The hydraulic properties are anisotropic. The most conductive lithological unit is fine-grained granite due to the numerous joints. On a small scale, effective hydraulic conductivity is very heterogeneous due to sparse connec-

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tion of fractures. Hydraulic conductivity is strongly scale dependent.

Fractures and fracture zones with different hydraulic conductivity and thus flow rates, characterize the aquifer in the Äspö area. These create a network of structures carrying water of different ages and chemical composition that will interact (mix) with each other in a complex way. Conductive fractures contain unlithified gouge material containing clay minerals. This implies that the reactive surface area along fractures is large (not just two planar surfaces). Perhaps ion exchange and sorption may be important for water composition.

At depth >800 m below laboratory floor (blf) saline water occupies the porosity; at depth 1,680 m (blf) a Ca–Na–Cl brine with 72 g L^{-1} total dissolved solids (TDS) is present. The fluid is remarkably similar to the waters from the KTB site (see below). At a shallower depth, a complex pattern involving fluids from different periods has been observed (Laaksoharju et al. 1999). Cl generally increases with depth. The origin of the brine is unknown and the authors consider a number of possibilities including metamorphic fluids, fluid inclusions, Permian evaporites, water–rock interaction and localized freezing.

Olkiluoto research site in Finland

A very large data set on the hydraulic properties of the basement and the chemical composition of water in the basement down to 1,000 m bgl (below ground level) has been produced by the research at the Olkiluoto site in Finland (Pitkänen et al. 1992; Palmén et al. 2004). The basement is dominated by mica gneiss and smaller amounts of granite. The hydraulic conductivity varied from 10^{-9}–10^{-8} m s^{-1} with no significant dependence on depth and lithology. However, some fracture zones were highly conductive (up to 10^{-5} m s^{-1}). Also, the typical fracture density of 3–5 m^{-1} was depth independent. However, the TDS of the water systematically and uniformly increased with depth and reached about 75 g L^{-1} at 1,000 m.

Stripa research site in Sweden

The International Stripa Project (1980–1990) produced hydrogeochemical data from ~1,000 water samples to a depth of 1,232 m bgl in Precambrian granite of the Baltic Shield (Nordstrom et al. 1989a,b). Water below 700 m is a Na–Ca–Cl fluid with a generally high pH (8–10) and a TDS maximum of 1,250 mg L^{-1}. Shallow water (<300 m) is Ca–HCO_3 recharge water, and the transition zone contains mixed waters. In-situ packer test data revealed 10^{-9} m s^{-1} for the typical conductivity of the granite (Gale et al. 1982, 1987). The transition of the chemical signature of the waters corresponds to a significant change in the hydraulic conductivity from 10^{-8} m s^{-1} to extremely low values of 10^{-11} m s^{-1} or lower at depth (Nordstrom et al. 1989b). The Stripa research site differs from other crystalline basement areas in that the fluids are less salty, have a higher pH and the hydraulic conductivity is lower by two to three orders of magnitude.

Data from deep boreholes in northern Switzerland

In the 1980s, the Swiss National Cooperative for the Disposal of Radioactive Waste (NAGRA) drilled a series of exploration and research drill holes through the sedimentary cover and into the crystalline basement in northern Switzerland. Hydraulic testing of the wells (up to 2,500 m deep) provided a wealth of data on the crystalline basement between about 1,000 and 2,500 m depth. Hydraulic conductivity ranges from 10^{-13} to 4 10^{-4} m s^{-1} or nine orders of magnitude! The variation shows little depth dependence but relates to local variations in fracture density, fracture geometry and other local parameters. Excluding high-permeability fracture zones and sections with unusually low fracture densities, the characteristic hydraulic conductivity is about 10^{-9} m s^{-1} (Leech et al. 1984; Butler et al. 1989; Moe et al. 1989; Belanger et al. 1989; Ostrowski and Kloska 1989; Pearson et al. 1989). The specific data reported in the cited literature are given in Table 1.

Urach research site in Germany

At the Urach test site, a 4,444-m-deep borehole has been drilled into the predominantly gneissic basement of southern Germany. The hydraulic conductivity at this site shows a pronounced anisotropy with 10^{-9} m s^{-1} in a horizontal and 10^{-11} m s^{-1} in a vertical direction (Stober and Bucher 2000). These values have been derived from long-term pumping and injection tests. Also, the fracture porosity at this site, 0.1–0.5%, is low compared with fracture porosities of about 1% in the Black Forest basement to the west. The deep fluid is a NaCl-brine with TDS=67 g L^{-1}.

Black Forest basement in Germany

The hydraulic properties of the crystalline basement of the Black Forest area has been extensively explored (Stober 1995) and a large number of well tests have been analyzed in order to retrieve hydraulic data. The test results from 169 experiments show log mean hydraulic conductivity of K=7 10^{-6} m s^{-1} (Fig. 2). The 16th percentile is at 8 10^{-7} m s^{-1} and the 84th percentile at 7 10^{-5} m s^{-1}. The wells range in depth from a few meters to 1,000 m bgl and tested intervals range from 5 to 800 m (test intervals: 98 tests at 5–100 m; 55 tests at 100–300 m; 16 tests at 300–822 m). The summarized data show that hydraulic conductivities in fractured granite are higher than in fractured gneiss by a factor of 100 or 2 orders of

Table 1 Water from NAGRA boreholes in northern Switzerland (Pearson et al. 1989)

Location	Depth (m bgl)	TDS (g L^{-1})	Water type
Böttstein	800	1.3	Na–SO_4–HCO_3–Cl
Kaisten	1,270	1.4	Na–SO_4–HCO_3
Leuggern	1,640	1.1	Na–SO_4–HCO_3
Schafisheim	1,870	8.6	Na–Cl
Weiach	2,210	6.5	Na–Cl

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magnitude (Fig. 2). Most drill holes penetrated the mixed basement with both types of rocks present in the tested sections, which results in K values near 10^{-7} m s^{-1}.

The flow porosity derived from the well tests varies between 0.1 and 2.08% (Stober 1995). None of the tested wells showed drawdown vs. log time patterns characteristic of a double porosity medium (Stober 1995). The hydraulic conductivity shows very large variance near the surface (Fig. 3). The variance decreases rapidly with depth and also the hydraulic conductivity decreases with depth according to the relationship $\log K = -1.38 \log z - 7.88$ (z in km, K in m s^{-1}) corresponding to $\log \kappa \sim -1.38 \log z - 16.4$ (z in km, κ in m^2). The data analysis suggests, however, that within the depth range of the data, the decrease in K with depth is pronounced in gneiss but is not visible in the granite basement (Stober 1996). Several test data indicated the influence of faults or fault zones acting as distant hydraulic boundaries. The hydraulic conductivities of such boundaries were lower than the K values retrieved from the radial flow period.

Cajon Pass research well in California, USA

At Cajon Pass, a 2,000-m-deep research well was drilled into the crystalline basement consisting chiefly of gneissic rocks. Coyle and Zoback (1988) reported hydraulic properties from three tested sections between 1,829 and 2,077 m. The K values are very low but interestingly increased from the shallowest tested section (1,829–1,905 m), at K=3.8 10^{-12} m s^{-1} to the deepest test interval (2,050–2,077 m) at K=1.39 10^{-9} m s^{-1}. For the entire section (1,829–2,115 m) K=1.29 10^{-11} m s^{-1}. The Cajon Pass data confirm that gneissic basement has relatively low conductivity and, even more important, that conductivity may increase with depth. There are three different weakly mineralized water types present in different fracture systems and the maximum TDS was ~2 g L^{-1} (Kharaka et al. 1988). The differences in water composition indicate different evolutionary paths and isolation of water within relatively proximal fracture systems.

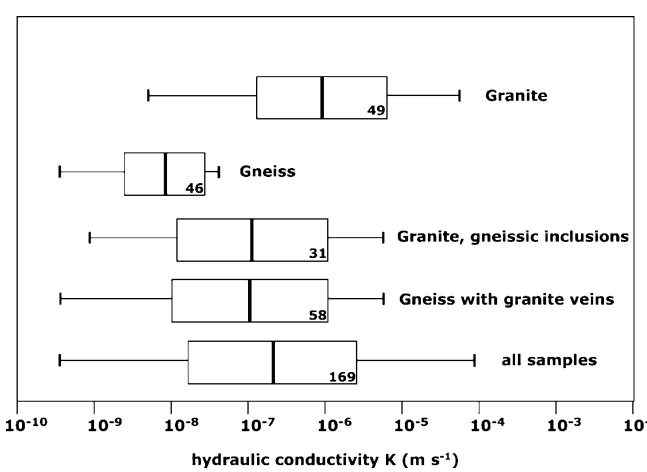

Fig. 2 Hydraulic conductivity data from hydraulic tests in 169 boreholes in the Black Forest crystalline basement to the depth of 1,000 m. Granite is clearly more conductive than gneiss, and mixed lithologies have intermediate K values. Shown are the logarithmic mean (*thick line*), the 16th and 84th percentiles and the total range of data (min and max)

Carnmenellis Granite in Cornwall, UK

The Carnmenellis granite in Cornwall has been extensively investigated in the context of a Hot Dry Rock (HDR) geothermal energy project to depths of 2,000 m (Pine and Ledingham 1984). The fracture density was found to be low at 0.1–0.3 m^{-1} but interconnected. Consequently, the hydraulic conductivity of the sparsely fractured granite at 1,500–2,000 m depth is unusually low (K~1.5 10^{-10} m s^{-1}). A series of deep boreholes into the same granite at another site, were drilled in the context of research related to underground nuclear waste repositories. Packer tests of 1-m sections from 50–650 m below surface revealed K values ranging from 10^{-9} to 10^{-5} m s^{-1}, with 10^{-8} to 10^{-7} m s^{-1} being more typical. Some of the highest flows were encountered at the deepest levels (Watkins 2003).

Canadian Shield

An interesting set of data is available from the Diavik Diamond Project in the Lac de Gras area in the Northwest Territories, Canada (Kuchling et al. 2000). The Canadian Shield at the site consists of granitic basement with diabase dike swarms and vertical cylindrical kimberlite pipes. Extensive hydraulic testing, including packer tests to 550 m depth, provided a mean hydraulic conductivity of 2 10^{-7} m s^{-1} in the country rock granite with kimberlite (4 10^{-7} m s^{-1}) being more and diabase being less (5 10^{-8} m s^{-1}) conductive. The data suggest a decrease

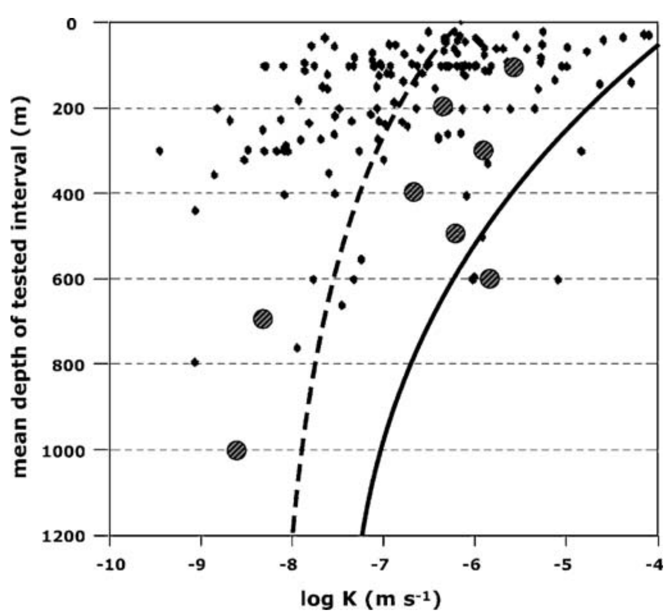

Fig. 3 Depth dependence of hydraulic conductivity (K) in the Black Forest area. *Solid curve*: depth dependence of crustal permeability proposed by Ingebritsen and Manning (1999). *Dashed curve*: fit curve through Black Forest hydraulic data. *Large dots* represent log mean of 200-m intervals

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in hydraulic conductivity with depth, with K=8 10^{-9} m s^{-1} at 550 m the deepest tested location.

The Lac du Bonnet granite batholith in southeastern Manitoba, Canada has been investigated as a part of the geosciences studies performed for the Canadian Nuclear Fuel Waste Management Program over the period 1980–1995 by Atomic Energy of Canada Limited (AECL). Compared to other areas on the Shield, the batholith is relatively undisturbed by major tectonic events. Core log data, pumping tests and hydraulic head measurements show that sub-horizontal fracture zones control the hydrogeology at AECL's Underground Research Laboratory. The fracture network is partially interconnected by subvertical fractures. The hydraulic conductivity of the fracture zones is as high as K=10^{-4} m s^{-1}, but in the unaltered, unfractured rock, values range from K=10^{-13} to 10^{-9} m s^{-1} (Stevenson et al. 1996). The hydraulic conductivity of the fractured granite on the vertical scale of some hundred meters is probably on the order of K=10^{-7} m s^{-1}, controlled by relatively few high-flow features. An extensive hydrochemical study has been undertaken at the site (Gascoyne 2004). Groundwater from 85 individual fractures have been sampled and analyzed from permeable zones in 53 boreholes, drilled to depths of up to 1,000 m in the batholith. The waters show a clear vertical stratification, with TDS gradually increasing to 90 g L^{-1}, and evolve first to Na–Ca–Cl–SO_4 and finally to Ca–Na–Cl waters at depth. Cl/Br ratios are low for the deep waters and range from 80–150 (ppm basis).

Soultz-sous-Forêts geothermal research site in France

The mainland European geothermal test site (HDR system) is located north of Strasbourg in the Rhine rift valley. The boreholes reach to >5,000 m in the crystalline basement which consists predominantly of granite. Several hydraulic tests were performed at various stages of the drilling operations. Circulation and tracer test data from two boreholes at 3,600–3,800 m were used to derive a flow porosity of 0.9–2.3% and a hydraulic conductivity of K=6.6 10^{-8} to 2.3 10^{-6} m s^{-1} (Aquilina et al. 2004). It is likely, however, that the hydraulic conductivity increased by the testing activities and that the pre-testing conductivities were smaller. At earlier stages of the drilling programs hydraulic conductivities of K=1.2 10^{-8} to 1.2 10^{-6} m s^{-1} for the depth interval 1,377–2,000 m have been reported (Jung 1990).

New data from tested boreholes in the Black Forest area

Near Waldshut-Tiengen in the southern Black Forest area, a deep drill hole penetrated several hundred meters of crystalline basement gneisses and granites. The well was drilled for geothermal energy exploration and a pumping test was performed in 2002 to assess the hydraulic properties of the fractured aquifer. During the pumping period of 42 d, 2.3 L s^{-1} (Q) were pumped from the open hole between 226 and 600 m below surface (b=374 m). The drawdown vs. elapsed time data (Fig. 4) show a final drawdown of 135 m. The early data describe nonlinear behavior, followed by a first linear segment after about 15 h. This linear segment ends after about 5 d. This segment is interpreted as being the pressure response typical of a two-dimensional radial convergent flow in a homogeneous aquifer. Consequently, the slope of this straight line corresponds to a transmissivity T=3.95 10^{-5} m^2 s^{-1} and, with b=374 m, to a hydraulic conductivity of ~1 10^{-5} m s^{-1}. This high value illustrates that the crystalline basement locally may display well-developed fracture porosities on a vertical scale of several hundred meters. From the hydraulic test data, additional hydraulic properties can be computed including the storage coefficient S=10^{-5} (Eq. 4 in Stober and Bucher 2005b) the skin factor s_F~4.03 and the drawdown caused by skin Δs_{skin}=37.31 m.

A second linear segment (Fig. 4) developed after about 5 d of pumping. This is the hydraulic response to a planar geological structure with a sealing effect, probably a watertight fault or fault zone. Fine-grained fault gouge or secondary clay seals may cause the low conductivity of

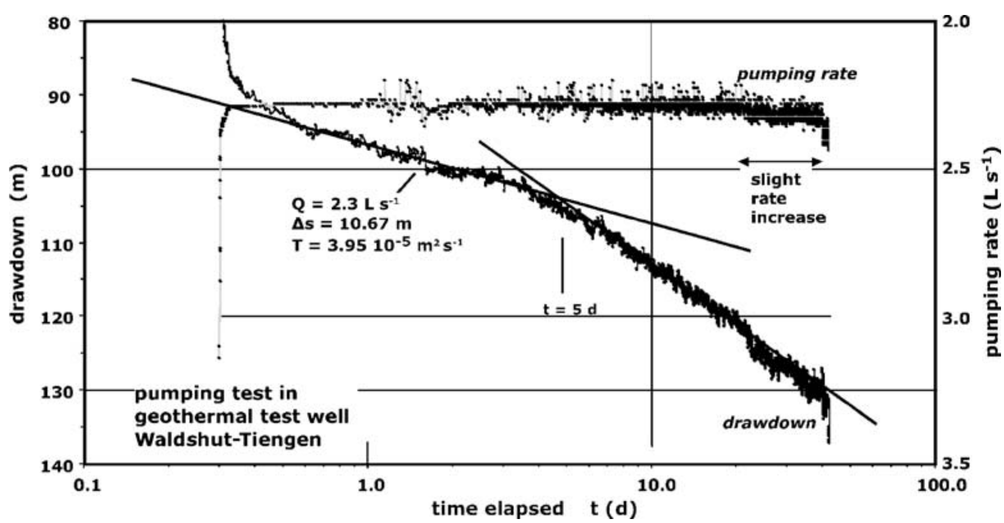

Fig. 4 Drawdown vs. log pumping time from the pumping test at the geothermal test site Waldshut-Tiengen. The period from 0.5–5 d is interpreted as a radial flow period in a homogeneous isotropic aquifer. At t >5 d the linear data array represents the influence of a distant hydraulic boundary with a lower hydraulic conductivity. Q=pumping rate, Δs=slope of *solid straight line* in meters and T=transmissivity

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the fault. The calculated horizontal distance from the well is about 2 km. After switching off the pump, the groundwater level recovered very quickly and reached the pre-pumping level after a short period of time, confirming the fast regeneration of the system.

Bad Teinach in the central Black Forest is the site of many deep drill holes into the crystalline basement. The predominant rock type is biotite-gneiss. Pumping tests in eight wells (Table 2) revealed the hydraulic properties of this complexly fractured and segmented basement. Similar to the test data from Waldshut-Tiengen, the drawdown (pressure of the corresponding water column) vs. time data from the well TEH1a show, after an initial period characterized by the wellbore storage, a linear segment that can be interpreted as radial flow period through an isotropic homogeneous aquifer (Fig. 5). The aquifer properties can be computed from the slope of the linear segment of the pressure vs. log t data. The resulting transmissivity T=2.07 10^{-4} m^2 s^{-1} converts to a computed hydraulic conductivity K=1.45 10^{-6} m s^{-1}. After the radial flow period, the data indicate the influence of a distant hydraulic boundary, with a hydraulic conductivity clearly lower than that of the fractured matrix with its relatively low but homogeneous fracture porosity.

The nearby well TEB3 (50–973 m; diameter 311 mm) was tested on 7 January 1991 with a pumping rate Q=1 L s^{-1} within 100 h (4.17 d); the water inflow points occurred between depths 30–794 m (b=764 m). The derived transmissivity T=6.61 10^{-7} m^2 s^{-1} yields a computed hydraulic conductivity K=8.65 10^{-10} m s^{-1}. The hydraulic conductivity between the two tested wells differs by four orders of magnitude. These data and similar observations from other locations and test sites in the Black Forest suggest that the basement is fragmented into blocks on the scale of kilometers. The blocks are separated by steep low-conductivity faults. The discrete blocks may be described by relatively homogeneous hydraulic properties. All block boundaries (faults) that appear as signals in the pumping tests (e.g., Figs. 4 and 5) have a significantly lower permeability than the rock matrix. However, the faults are probably extremely anisotropic and may conduct water along but not across them (Stober et al. 1999). Exceptions to the rule also exist, however, because linear high-permeability zones mapped in some valleys of the Black Forest area can be correlated with faults.

Table 2 Well test data from Bad Teinach, Black Forest

Well No.	OH (m bgl)	BH (m bgl)	Transmissivity (m^2s^{-1})	K (m s^{-1})
TEO1	178	1001	2.11E-06	2.56E-09
TEB1	234	600	6.40E-06	1.75E-08
TEB2a	80	154	1.85E-06	2.50E-08
TEB2b	80	234	2.42E-06	1.57E-08
TEH1a	186	329	2.07E-04	1.45E-06
TEH1b	191	500	3.82E-04	1.26E-06
TEH3	98	320	2.30E-05	1.04E-07
TEB3	30	794	6.61E-07	8.65E-10

OH top of open hole, *BH* bottom hole, *K* hydaulic conductivity

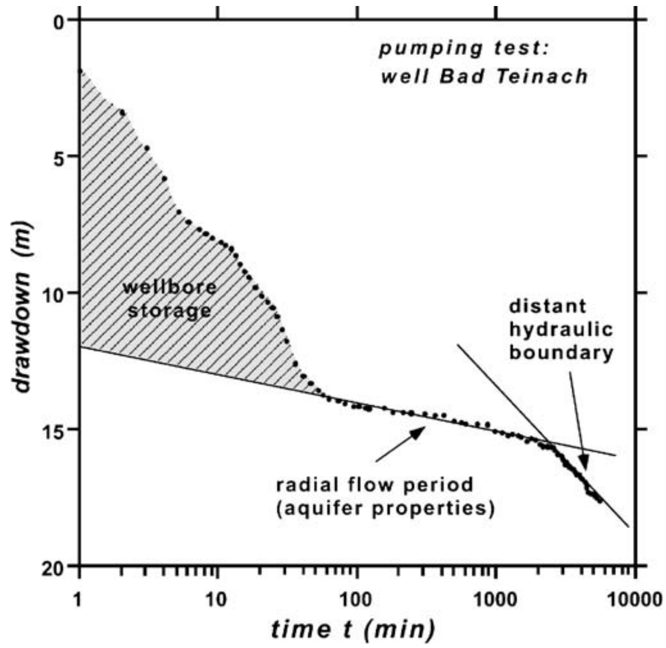

Fig. 5 Drawdown vs. log pumping time from the pumping test at Bad Teinach, Black Forest. The test is from the basement well TEH1a (Table 2), December 1973, (screened casing: 186–329 m bgl). Discharge (Q)=2 L s^{-1}, total time 192 h=8 d, with three distinct flow periods influenced by: (1) wellbore storage, (2) properties of homogeneous isotropic aquifer (transmissivity T=2.07 10^{-4} m^2 s^{-1}), and (3) distant hydraulic boundary (most likely to be a less permeable fault)

New data from the continental deep drilling site (KTB), Germany

The 4,000 m deep pilot hole at the KTB site has been extensively tested during a pumping test of 1 year duration from 2002 to 2003 (Stober and Bucher 2005b). The basement rocks consist of various para- and ortho-gneisses and amphibolites. The tested section of 150 m was between 3,850 m bgl and bottom hole at 4,000 m. The tests produced an average hydraulic conductivity K=4.07 10^{-8} m s^{-1}. The hydraulic test data also permitted the computation of the effective flow porosity ϕ_e~0.68% from the measured storativity S and Eq. (1) given above. This value for ϕ_e is the average flow porosity of the fractured basement aquifer. The test data are consistent with a hydraulic behavior of the basement that can be characterized as homogeneous isotropic aquifer. However, the influence of a hydraulic boundary becomes visible in the test data with increasing pumping time. The boundary is characterized by lower hydraulic conductivity and is about 1.5 km away from the pilot hole at 4,000 m depth. One may interpret the hydraulic boundary as a steeply dipping low-permeability fault zone acting as a fault-block boundary (Stober and Bucher 2005b). A remarkable amount of saline water (2.3 10^4 m^3) was pumped from the aquifer during the test period, showing that the fractured basement at 4 km depth can be pumped at a constant rate for a long period of time, just like a near surface aquifer. The fluid is a Ca–Na–Cl water-type with a

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total mineralization of 62 g L^{-1} and a Cl/Br ratio of 75 (ppm basis).

A large-scale fluid injection test in the KTB drill hole was carried out in 2004. This new extended unique hydraulic test lasted for 10 months. The total amount of >9 10^4 m^3 of water was injected at an average rate of 3.4 L s^{-1}. The results of one experiment within the test period are shown in Fig. 6. The pressure vs. log time data resulted in the computed hydraulic conductivity K=5.42 10^{-8} m s^{-1}. This new value represents the hydraulic conductivity of the fractured basement on the scale of hundreds of meters at 4 km depth. It is in excellent agreement with the K value derived from the pumping test (see above). This is probably the most trustworthy value for the hydraulic conductivity of continental crystalline crust at 4,000 m. The K value corresponds to a permeability κ~1.6 10^{-15} m^2.

Fluid flow in the crystalline basement: a discussion and some conclusions

Fluid flow in fractured aquifers is driven by hydraulic gradients resulting from a number of different feasible causes and imbalances including topography, and thermal and chemical disequilibrium. The typical hydraulic conductivity of the crystalline basement down to several km depths (log K= −8±1 m s^{-1}) as summarized above is well within the conditions for advective heat and solute transport (Manning and Ingebritsen 1999).

Topography induces differences in hydraulic head (hydraulic potential) leading to hydraulic gradients. Topography related hydraulic gradients cause, for instance, thermal springs or the upwelling of deep-seated fluids (Toth 1980; Stober et al. 1999). The hydraulic potential difference usually induces groundwater circulation, reaching a depth of several km. Rainwater and other precipitation migrate downward and its initial low mineralization gradually increases along the flow paths. At the early stages of groundwater evolution, the Ca–HCO_3-type water has a low total dissolved solids (TDS) content. The water evolves to a highly mineralized Na–Ca–Cl-type water or brine, with a sulphate-rich intermediate stage at many locations. This stratification is well documented at Stripa, Olkiluoto, Urach and other sites (see references above). In response to topography induced heads, deep water may rise along hydraulic conductivity boundaries such as faults and may reach the surface as a salty hot spring after a long passage through deeper parts of the crystalline basement at the point of lowest hydraulic potential, e.g., in valleys (Stober et al. 1999).

Topography-related hydraulic gradients (h) for the Black Forest area are typically 0.05 m m^{-1} with a range 0.002–0.2 m m^{-1}, corresponding to pressure gradients from 0.02 to 2 MPa km^{-1} with a typical value of 0.5 MPa km^{-1}. The background gradient at the Äspö research site is 0.066 m m^{-1} (Mazurek et al. 2003). These gradients characterize many other moderately hilly or mountainous areas with moderate slope steepness and maximum elevation differences ranging from 500 to 1,000 m. Darcy flow rates for the basement with these typical hydraulic gradients h for the upper 4 km of continental crust range from less than 1 mm a^{-1} to 4 m a^{-1} (ie. 1 mm per year). Using the hydraulic conductivity of the KTB site at 4 km depth, the velocity range is 1–20 cm a^{-1}. This corresponds to advective flow rates of 10–200 L a^{-1} m^{-2} (Fig. 7). Figure 7 is a graphical representation of Darcy's Law $q= -K\,\nabla h$ with log q on the y-axis and a one-dimensional ∇h (hydraulic gradient) on the x-axis. It shows the resulting flow q for a range of feasible gradients in the basement and derived K values from hydraulic tests

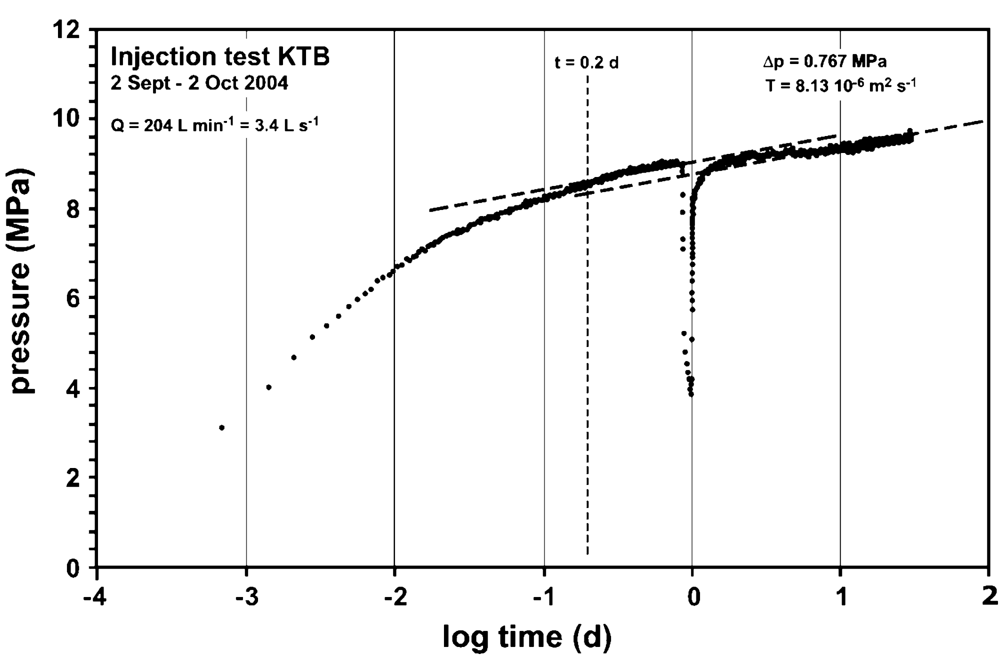

Fig. 6 Pressure (mega pascals: MPa) vs. log time (days), from an injection test of 30 d duration in the 4,000 m deep research borehole at the KTB site. The pressure excursion after 1 d (time 10^0) was caused by a short pump failure. Δp=slope of *stippled line* in MPa. The properties of the homogeneous isotropic aquifer (*stippled lines*) dominate after 0.2 d (*marker*). The *stippled lines*are displaced by the effects of the interruption in pumping

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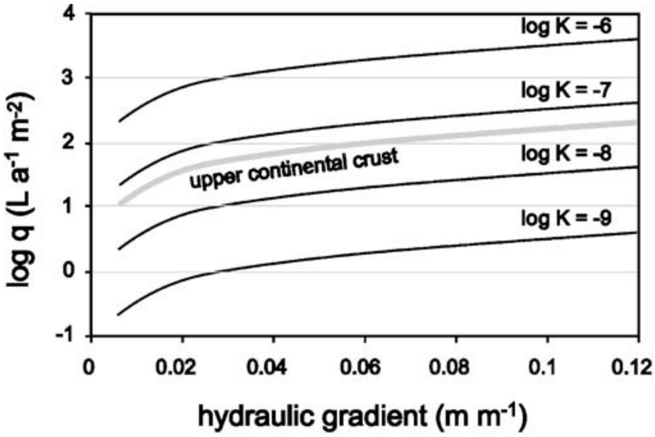

Fig. 7 Flow rate (q) vs. hydraulic gradient, for characteristic values of hydraulic conductivities (K) in the crystalline basements. The K value for the basement at 4 km depth at the KTB site is highlighted in *gray*

summarized in this paper. Within the range of reported hydraulic conductivities of log $K = -9$ to -6 typical flow rates range from 1.5–1,500 L a^{-1} m^{-2} for a gradient of 0.05 m m^{-1}. The gradients used above are present in the uppermost km of the crust and are damped with increasing depth due to frictional effects. It can be expected that at 4 km depth flow rates will be considerably reduced but still significant.

The brittle continental crust generally contains salt-water-saturated fracture porosity. Hydraulic conductivity values derived from in-situ well tests at various scales range from 10^{-4} to 10^{-11} m s^{-1}. The most typical values center around log $K = -8 \pm 1$ m s^{-1}at up to 4 km depth (Table 3). The hydraulic conductivity of the crystalline basement thus is high enough to allow for advective water/fluid and heat transport. Because of the high permeability, pressure conditions are hydrostatic throughout the brittle crust, except for strata of low-permeability sediments. In an intensely fractured basement, granite is generally more pervious than gneiss. Mean hydraulic conductivity of gneiss and granite of the Black Forest basement (SW Germany) ranges from $K = 5\ 10^{-8}$ m s^{-1} (gneiss) to $K = 1\ 10^{-6}$ m s^{-1} (granite), with an average $K = 2.14\ 10^{-7}$ m s^{-1}. The hydraulic conductivity of granite and gneiss can be extremely low in basement units with low fracture densities. Hydraulic conductivity is expected to systematically decrease with depth reaching finally $K \sim 5\ 10^{-11}$ m s^{-1} in the ductile crust (Manning and Ingebritsen 1999). The available hydraulic conductivity data from well tests in deep (but not deep enough) boreholes show large local variations with either K decreasing or K increasing with depth. At some sites, a decrease of K with depth is evident in the gneissic but not in the granitic basement (Stober 1996). The counterintuitive fact that the hydraulic conductivity may locally increase rather than decrease with depth could be explained by the active ongoing deformation and fracture formation in the crust (e.g., Cajon Pass).

Manning and Ingebritsen (1999) proposed a depth dependence of the permeability log $\kappa = -3.2$ log $z - 14$ (with depth z in km and κ in m^2) from thermal modeling and metamorphic fluid flow estimates. As a result of the

Table 3 Summary of site characterization (data compiled from literature cited in the text)

Investigation area	Depth (m)	Geology	Hydraulic conductivity K (m s^{-1})	TDS in g L^{-1} (depth) water type
Äspö, Sweden	<450 +1,680	Granitic rocks	$3\ 10^{-10}$–$1\ 10^{-8}$ ndp	72 (450+1,680 m) Ca–Na–Cl
Olkiluoto, Finland	<1,000	Mica-rich gneisses, minor granite	$1\ 10^{-9}$–$1\ 10^{-8}$ ndp up to $1\ 10^{-5}$	75 (800 m) Na–Cl
Stripa, Sweden	<1,232	Granite	$1\ 10^{-11}$ – $1\ 10^{-8}$ rep $1\ 10^{-9}$ K decreases with depth	1.25 (1,200 m) Na–Ca–Cl Ca–HCO_3 (<300 m)
Carnmenellis, Cornwall, UK	<2,200	Granite	$1\ 10^{-9}$–$1\ 10^{-5}$ ndp to 700 m rep $1\ 10^{-8}$–$1\ 10^{-7}$ K lower values at depth >700 m	19.3 (1,700 m) Na–Cl
Canadian Shield Lac de Gras (Diavik), Northwest Territories	<550	Granitic rocks	$8\ 10^{-9}$–$4\ 10^{-7}$ rep $2\ 10^{-7}$ K possibly decreases with depth	6.4 (1,000 m) 0.44 (350 m) Na–Cl[a]
Canadian Shield, Lac du Bonnet, Manitoba	<1,000	Granitic rocks	$1\ 10^{-13}$–$1\ 10^{-4}$ rep $1\ 10^{-7}$	90 (1,000 m) Ca–Na–Cl
NAGRA deep wells, N Switzerland	<2,500	Granite, gneiss	$1\ 10^{-13}$–$4\ 10^{-4}$ ndp rep $1\ 10^{-9}$	6.5 (2,210 m) Na–Cl
Urach3, S Germany	<4,444	Gneisses	$1\ 10^{-9}$	67 (4,000 m) Na–Cl
KTB, Germany	<4,000	Amphibolite	$4\ 10^{-8}$	62 (4,000 m) Ca–Na–Cl
Soultz-sous-Forêts, France	<5,000	Granites	3,600–3,800 m: $7\ 10^{-8}$–$2\ 10^{-6}$	100 (5,000 m) Na–Cl
Cajon Pass, California, USA	<2,077	Gneisses	$3.8\ 10^{-12}$–$1.4\ 10^{-9}$ rep $1\ 10^{-11}$ K increases with depth	2.1 (1,829–1,900 m) 1.2 (1,829–2,115 m) Ca–Cl; Na–HCO_3–SO_4; Na–SO_4

rep representative value for the site at depth (characteristic value); *ndp* no depth dependence

[a] measured 0.44 g L^{-1} at 350 m, expected 1 g L^{-1} at 600 m (Kuchling et al. 2000)

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type of data used for deriving the equation, the relationship describes permeability averaged over large time and length scales (Manning and Ingebritsen 2001). The equation is an excellent model for the large-scale hydraulic properties of the crust and it is consistent with this study's compilation of well test data in a very broad view (Fig. 8). However, it under-estimates the hydraulic conductivity of the granitic basement and over-estimates Kvalues for gneissic crust (Fig. 8). The summarized K data show a very strong dependence on lithology and on the local deformation history of a particular area. The presented data clearly do not suggest that a single general equation is capable of adequately describing the hydraulic conductivity-depth relationships in the crystalline basement. In this sense, the equation is perhaps somewhat misleading. Because the permeability shows high variance among localities, the relationship should not be used as a predictive tool for hydraulic properties at a specific site for the depth of interest, no matter how tempting this may appear in prospecting for geothermal energy. On the other hand, the equation correctly predicts crustal permeabilities that are much higher than those often advocated in flow models where the basement often is assumed to be impervious and fluid flow is restricted to the sedimentary cover on top of it.

The generally high basement permeability is fully supported by this study's compiled well test data. The best-fit equation through the log K data from 169 well tests in the Black Forest basement has been given above as:

$$\log \kappa = -1.38 \log z - 16.4 \qquad (2)$$

with depth z in km and κ in m^2. The equation suggests that the near surface permeability of the basement is lower than that given by the equation of Manning and Ingebritsen (1999,) and that the permeability decreases more slowly with depth. The permeability at the brittle-ductile transition predicted by Eq. (2) is 10^{-18} m^2; however, the extrapolation is probably insecure.

On the other hand, the formulation of a generalized K vs. z equation for the continental crust is very helpful because it supports the conclusion derived from the presented well test data that groundwater in the crystalline basement is a continuum that may reach from the unconfined water table near the surface to the brittle-ductile transition zone. The depth of the brittle-ductile transition depends on temperature, material and stress but it occurs typically at about 12–14 km (Wintsch et al. 1995). Deeper in the crust, brittle flow porosity is absent and fluids can migrate only by pressures much greater than hydrostatic pressure. The hydraulic conductivity of the crystalline basement at 10 km depth near the brittle-ductile transition is normally still K~2 10^{-11} m s^{-1} or higher. Note that this is higher by several orders of magnitude than the hydraulic conductivity of many shales and marls that may occur in the cover of the basement. Jurassic Opalinus clay in the cover sequence of the basement in northern Switzerland, for example, has a reported conductivity of 2 10^{-13} m s^{-1} derived from in-situ hydro tests (Marschall et al. 2005). Other shale and marl units of the same cover sequence are characterized by K values in the range 2 10^{-14} to 4 10^{-12} (Gimmi and Waber 2004). The analysis resulting from this study shows that the hydraulic conductivity of the granitic basement underlying sedimentary basins typically has log K values of −7 to −6 or locally higher, suggesting that the basement may be drastically more conductive than shale-rich cover sequences, e.g., by a factor of a million or more. Thus, large-scale fluid flow may be forced to the basement. In numerical models of crustal-scale fluid flow, the no-flow

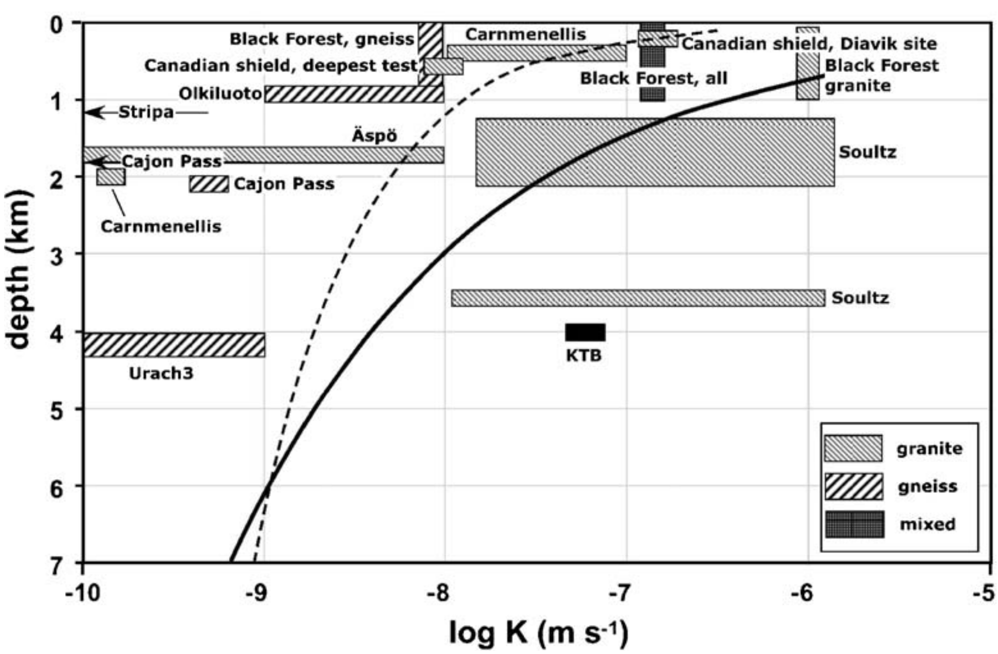

Fig. 8 Hydraulic conductivity (K) of the crystalline basement. Data from well tests summarized and cited in the text. *Solid curve*: log K depth dependence derived by Ingebritsen and Manning (1999). *Dashed curve*: extrapolation of log κ= f(log$z$) dependence derived from Black Forest hydraulic data

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boundary should be chosen at the brittle-ductile transition zone (10–14 km).

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References

Aquilina L, de Dreuzy JR, Bour O, Davy P (2004) Porosity and fluid velocities in the upper continental crust (2 to 4 km) inferred from injection tests at the Soultz-sous-Forêts geothermal site. Geochim Cosmochim Acta 68:2405–2415

Banks D, Odling NE, Skarphagen H, Rohr-Trop E (1996) Hydraulic conductivity and stress in crystalline rocks. Terra Nova 8:223–235

Behr HJ, Gerler (1987) Inclusions of sedimentary brines in postvariscan mineralisations in the Federal Republic of Germany: a study by neutron activation analysis. Chemi Geol 61:65–77

Belanger DW, Freeze GA, Lolcama JL, Pickens JF (1989) Interpretation of hydraulic testing of crystalline rock at the Leuggern Borehole, Nagra technical report, NTB 87–19, National Cooperative for the Disposal of Radioactive Waste (NAGRA), Baden, Switzerland

Benderitter Y, Elsass Ph (1995) Structural control of deep fluid circulation at the Soultz HDR site France: a review. Geotherm Sci Tech 4:227–237

Berckhemer H, Rauen A, Winter H, Kern H, Kontny A, Lienert M, Nover G, Pohl J, Popp T, Schult A, Zinke J, Soffel HC (1997) Petrophysical properties of the 9-km-depth crustal section at KTB. J Geophys Res 102(B8):18337–18361

Butler GA, Cauffman TL, Lolcama JL, Longsine DE, McNeish JA (1989) Interpretation of hydraulic testing at the Weiach Borehole, Nagra technical report, NTB 87–20, National Cooperative for the Disposal of Radioactive Waste (NAGRA), Baden, Switzerland

Carden RS, Nicholson RW, Pettitt RA, Rowley JC (1985) Unique aspects of drilling and completing hot, dry rock geothermal wells. J Petrol Technol 5:821–834

Caine JS, Evans JP, Forster CB (1996) Fault zone architecture and permeability structure. Geology 24:1025–1028

CFCCF (Committee on Fracture Characterization and Fluid Flow) (1996) Rock fractures and fluid flow. National Research Council, National Academy Press, Washington, DC, 551 pp

Clauser C (1992) Permeability of crystalline rocks. EOS Trans Am Geophys Union 73:233–237

Coyle BJ, Zoback MD (1988) In situ permeability and fluid pressure measurements at ~2 km depth in the Cajon Pass research well. Geophys Res Lett 15(9):1029–1032

Edmunds WM, Savage D (1991) Geochemical characteristics of groundwater in granites and related crystalline rocks. In: Downing RA, Wilkinson WB (eds) Applied groundwater hydrology: a British perspective. Clarendon, Oxford, pp 199–216

Edmunds WM, Kay RL, McCartney RA (1985) Origin of saline groundwaters in the Carnmenellis granite (Cornwall, England): natural processes and reaction during hot dry rock reservoir circulation. Chem Geol 49:287–301

Emmermann R, Althaus E, Giese P, Stöckhert B (1995) KTB Hauptbohrung results of geoscientific investigation in the KTB Field Laboratory, Final report: 0–9,101 m. KTB Report 95–2, E. Schweizerbart`sche Verlagsbuchhandlung, Stuttgart

Fritz P, Frape SK (1983) Saline waters in the Canadian Shield: a first overview. Chem Geol 36:179–190

Gale JE, Rouleau A, Witherspoon PA, Wilson CR (1982) Progress in the Hydrogeological Characterization of the Stripa Site, Report LBL-14878/SAC-49/UC-70, Lawrence Berkeley Laboratory, Berkeley, CA

Gale JE, MacLeod R, Welhan JE, Cole C, Vail L (1987) Hydrogeological characterization of the Stripa site. Stripa Project Technical Report 87–15, SKB, Stockholm

Gascoyne M (2004) Hydrogeochemistry, groundwater ages and sources of salts in a granitic batholith on the Canadian Shield, southeastern Manitoba. Appl Geochem 19:519–560

Gimmi T, Waber HN (2004) Modelling of tracer profiles in pore water of Argillaceous rocks in the Benken borehole: stable water isotopes, chloride, and chlorine isotopes. NAGRA Technical Report 04–05, NAGRA, Baden, Switzerland, 91 pp

Guillot L, Siitari-Kauppi M, Hellmuth K-H, Dubois C, Rossy M, Gaviglio P (2000) Porosity changes in a granite close to quarry faces: quantification and distribution by ^{14}C-MMA and Hg porosimetries. Eur Phys J AP 9:137–146

Hayba DO, Ingebritsen SE (1994) The computer model hydrotherm: a three-dimensional finite-difference model to simulate groundwater flow and heat transport in the temperature range of 0 to 1,200°C. US Geol Surv Water Resour Invest Rep 94–4045:234

Ingebritsen SE, Manning CE (1999) Geological implications of a permeability-depth curve for the continental crust. Geology 27:1107–1110

Jakob A, Mazurek M, Heer W (2003) Solute transport in crystalline rocks at Äspö–II: blind predictions, inverse modelling and lessons learnt from test STT1. J Contam Hydrol 6:175–190

Jung R (1990) Hydraulic fracturing and hydraulic testing in the granitic section of borehole GPK1, Soultz-sous-Forêts, Final Report BMFT 0326425 B/EC EN3G-0080-D(B), Bundesanstalt für Geowissenschaften und Rohstoffe (BGR), Hannover

Kanz W (1987) Grundwasserfließwege und Hydrogeochemie in tiefen Graniten und Gneisen [Groundwater flowpaths and hydrochemistry in deep granite and gneiss]. Geol Rundschau 76:265–283

Kharaka YK, Ambats G, Evans WC, White AF (1988) Geochemistry of water at Cajon Pass, California: preliminary results. Geophys Res Lett 15:1037–1040

Konzuk JS, Kueper BH (2004) Evaluation of cubic law based models describing single-phase flow through a rough-walled fracture. Water Resour Res 40(W02402):1–17

Kozlovsky YA (1984) The world's deepest well. Sci Am 251: 106–112

Kuchling K, Eng P, Chorley D, Geo P, Zawadzki W (2000) Hydrogeological modeling of mining operations at the Diavik Diamonds Project. 6th International Symposium on Environmental Issues and Waste Management in Energy and Mineral Production, Calgary, Alberta, 30 May–2 June 2000, pp 1–8

Leech RE, Kennedy KG, Gevaert D (1984) Sondierbohrung Böttstein: hydrological testing of crystalline rocks. NAGRA technical report, NTB 85-09, National Cooperative for the Disposal of Radioactive Waste (NAGRA), Baden, Switzerland

Laaksoharju M, Tullborg EL, Wikberg P, Wallin B, Smellie J (1999) Hydrogeochemical conditions and evolution at the Aspo HRL, Sweden. Appl Geochem 14:835–859

Manning CE, Ingebritsen SE (1999) Permeability of the continental crust: implications of geothermal data and metamorphic systems. Rev Geophys 37:127–150

Manning CE, Ingebritsen SE (2001) Role of the brittle-ductile transition in large-scale fluid flow in active continental crust. Abstracts, GSA Earth System Processes, Edinburgh, T11.1

Marschall P, Horseman S, Gimmi T (2005) Characterisation of gas transport properties of the opalinus clay, a potential host rock formation for radioactive waste disposal: oil and gas science and technology. Rev IFP 60:121–139

May F, Hoernes S, Neugebauer HJ (1996) Genesis and distribution of mineral waters as a consequence of recent lithospheric dynamics: the Rhenish Massif, Central Europe. Geol Rundschau 85:782–799

Mazurek M (2000) Geological and hydraulic properties of water-conducting features in crystalline rocks. In: Stober I, Bucher K (eds) Hydrogeology of crystalline rocks. Kluwer, Dordrecht, pp 3–26

Mazurek M, Jakob A, Bossart P (2003) Solute transport in crystalline rocks at Äspö–I: geological basis and model calibration. J Contam Hydrol 6:157–174

Moe H, McNeish JA, McCard JP, Andrews RW (1989) Interpretation of hydraulic testing at the Schafisheim Borehole, Nagra technical report, NTB 89–09. National Cooperative for the Disposal of Radioactive Waste (NAGRA), Baden, Switzerland

Möller P, Weise SM, Althaus E, Bach W, Behr HJ, Borchardt R, Bräuer K, Drescher F, Erzinger J, Faber E, Hansen BT, Horn EE, Huenges E, Kämpf H, Kessels N, Kirsten T, Landwehr D, Lodemann M, Machon L, Pekdeger A, Pielow H-U, Reutel C, Simon K, Walther J, Weinlich FH, Zimmer M (1997) Paleo- and recent fluids in the upper continental crust: results from the German Continental Deep Drilling Program (KTB). J Geophys Res 102:18245–18256

Nordstrom DK, Andrews JN, Carlsson L, Fontes J-C, Fritz P, Moser H, Olsson T (1985) Hydrogeological and hydrogeochemical investigations in boreholes. Final report of the phase I geochemical investigations of the Stripa groundwaters. Technical Report Stripa Project 85–06, Stockholm

Nordstrom DK, Olsson T, Carlsson L, Fritz P (1989a) Introduction to the hydrogeochemical investigations within the International Stripa Project. Geochim Cosmochim Acta 53:1717–1726

Nordstrom DK, Ball JW, Donahoe RJ, Whittemore D (1989b) Groundwater chemistry and water-rock interactions at Stripa. Geochim Cosmochim Acta 53:1727–1740

Norton D, Knapp R (1977) Transport phenomena in hydrothermal systems: the nature of porosity. Am J Sci 277:913–936

Ostrowski LP, Kloska MB (1989) Final interpretation of hydraulic testing at the Siblingen Borehole, Nagra technical report, NTB 89–10, National Cooperative for the Disposal of Radioactive Waste (NAGRA), Baden, Switzerland

Palmén J, Vaittinen T, Ahokas H, Nummela J, Heikkinen E (2004) 3D-model of salinity of bedrock groundwater at Olkiluoto. Working report 2004-53, Posiva Oy, Finland, 92 pp, http://www.posiva.fi. Cited 19 August 2006

Pauwels H, Fouillac C, Fouillac A-M (1993) Chemistry and isotopes of deep geothermal saline fluids in the Upper Rhine Graben: origin of compounds and water-rock interactions. Geochim Cosmochim Acta 57:2737–2749

Pearson FJ, Lolcama JL, Scholtis A (1989) Chemistry of waters in the Böttstein, Weiach, Riniken, Schafisheim, Kaisten and Leuggern boreholes: a hydrochemically consistent data set. NAGRA Technical Report 86–19, Baden, Switzerland, 102 pp

Pine RJ, Ledingham P (1984) In-Situ hydraulic parameters for the Carnmenellis granite hot dry rock geothermal energy research reservoir. J Pet Technol 25:1982–1990

Pitkänen P, Snellman M, Leino-Forsman H, Front K (1992) Groundwater chemistry and water-rock interaction at Olkiluoto. Report YJT-92-02, Nuclear Waste Commission of Finnish Power Companies, Helsinki, 321 pp

Stanfors R, Rhen I, Tullborg EL, Wikberg P (1999) Overview of geological and hydrogeological conditions of the Aspo hard rock laboratory site. Appl Geochem 14:819–834

Schmassmann H, Balderer W, Kanz W, Pekdeger A (1984) Beschaffenheit der Tiefengrundwässer in der zentralen Nordschweiz und angrenzenden Gebieten [Chemistry of deep groundwater in central northern Switzerland and surrounding areas]. NAGRA Technical report 84–21, NAGRA, Baden, Switzerland, 335 pp

Stevenson DR, Kozak ET, Davison CC, Gascoyne M, Broadfoot RA (1996) Hydrogeologic characteristics of domains of sparsely fractured rock in the granitic Lac du Bonnet Batholith, southeastern Manitoba, Canada. Atomic Energy of Canada Limited Report, AECL-11558, COG-96-117, Atomic Energy of Canada Limited, Ottawa, ON

Stober I (1995) Die Wasserführung des kristallinen Grundgebirges [Water in the crystalline basement]. Ferdinand Enke Verlag, Stuttgart, Germany, 191 pp

Stober I (1996) Researchers study conductivity of crystalline rock in proposed radioactive waste site. EOS Trans Am Geophys Union 77:93–94

Stober I, Bucher K (2000) Hydraulic Properties of the upper Continental Crust: data from the Urach 3 geothermal well. In: Stober I, Bucher K (eds) Hydrogeology of crystalline rocks, Kluwer, Dordrecht, pp 53–78

Stober I, Bucher K (2004) Fluid sinks within the Earth's crust. Geofluids 4:143–151

Stober I, Bucher K (2005a) Deep-fluids: Neptune meets Pluto. In: Voss C (ed) The future of hydrogeology. Hydrogeol J 13:112–115

Stober I, Bucher K (2005b) The upper continental crust, an aquifer and its fluid: hydraulic and chemical data from 4 km depth in fractured crystalline basement rocks at the KTB test site. Geofluids 5:8–19

Stober I, Richter A, Brost E, Bucher K (1999) The Ohlsbach Plume: natural release of deep saline water from the crystalline basement of the Black Forest. Hydrogeol J 7:273–283

Toth J (1980) Cross-formational gravity flow of groundwater: a mechanism of transport and accumulation of petroleum (the generalized hydraulic theory of petrol migration). In: Roberts WH, Cordell HJ (eds) Problems of petrol migration, Am. Assoc. Petrol. Geologists, Studies Geol vol 10, AAPG, Tulsa, OK, pp 121–167

Walker D, Rhén I, Gurban I (1997) Summary of hydrogeological conditions at Aberg, Beberg and Ceberg. SKB Technical Report, TR 97–23, SKB, Stockholm, Sweden

Watkins DC (2003) Determining a representative hydraulic conductivity of the Carnmenellis Granite of Cornwall, UK, based on a range of sources of information. In: Krásný J. Hrkal Z, Bruthans J (eds) Groundwater in fractured rocks, IHP International Hydrological Program, Bratislava, Slovak Rep, pp 113–114

Wintsch RP, Christoffersen R, Kronenberg AK (1995) Fluid-rock reaction weakening of fault zones. J Geophys Res 100 (B7):13021–13032

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