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PAPER

Examining the assumptions of the single-porosity archetype for transport in bedrock aquifers

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Abstract
Bedrock aquifers are often characterized using porous medium concepts, but it is unclear to what extent these aquifers comply with the additional assumptions inherent in porous medium models. The core assumption is that aquifers can be treated as continuous single-porosity porous media, with Darcy's law describing the flow. The auxiliary assumptions include that the aquifer framework is insoluble, and that permeability varies randomly in space. The combination of these assumptions is referred to here as the single-porosity archetype. The applicability of the major assumptions to transport in bedrock aquifers was examined by considering substantial data sets for each assumption. It is shown that weathering often substantially increases the permeability in both carbonate and silicate rocks. Nonrandom spatial organization of permeability frequently occurs during deposition and diagenesis of rocks, as bedding planes, fractures zones, and interflow zones. Subsequently, self-organization occurs due to feedbacks between flow and weathering. Bedrock aquifers thus deviate substantially from the assumptions of the single-porosity archetype. The common presence of continuous preferential flow paths shows that bedrock aquifers often behave as dual-porosity aquifers when considering transport.

Keywords Conceptual models · Groundwater hydraulics · Groundwater geology · Heterogeneity · Preferential flow

Introduction

A major problem in bedrock hydrogeology is the lack of a clear understanding of the nature of the permeability and of the processes that created it—for instance, Anderson (2008, p. 1) noted that "groundwater processes [in fractured rock and karst] are still largely an open research question". An important consideration is that order of magnitude differences exist between the different lithologies (Freeze and Cherry 1979; Gleeson et al. 2011). These differences imply that one or more physical or chemical properties are important for determining permeability.

Analysis of flow and transport in bedrock aquifers commonly assumes that these aquifers behave as single-porosity porous media. The conceptual framework of Hubbert (1940) has provided the basis for the mathematical analysis of such flow. This paper has been described as being "of monumental significance to groundwater theory" (Anderson 2008, p. 72). The core assumption of Hubbert (1940) was that aquifers can be treated as continuous porous media for flow, with Darcy's law describing that flow. In addition, several auxiliary assumptions were made, including that porosity and permeability are homogeneous and isotropic (p. 788), and that the aquifer framework is "insoluble and chemically inert" (p. 788). A subsequent development was to assume that permeability varies randomly in space (Freeze 1975). The combination of the core assumption plus the auxiliary assumptions is referred to here as the single-porosity archetype.

The high permeability and often fairly homogeneous characteristics of sand aquifers has made them popular for studying groundwater processes (Anderson and McCray 2011). Furthermore, permeability in unconsolidated sediments is proportional to grain size, and thus to pore size, providing an easily understood link in terms of aquifer hydraulics. The framework developed by Hubbert (1940) of treating aquifers as single-porosity porous media has become dominant in hydrogeology textbooks (e.g. Freeze and Cherry 1979; Anderson et al. 2015; Dassargues 2019; Hiscock and Bense 2021; Fetter and Kreamer 2022).

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There are two major aspects to analysing the hydraulic response of aquifers. Flow analysis considers specific discharge, which is the volume of water moving through a unit cross-section of the aquifer in a unit of time, and is a function of hydraulic conductivity and hydraulic gradient. Transport analysis considers the movement of water and often associated contaminants through aquifers. In addition to hydraulic conductivity and hydraulic gradient, it also requires a value for effective porosity. If there is no preferential flow in an aquifer then it will behave as a single-porosity porous-medium continuum. However, in bedrock aquifers, almost all the flow may be through networks of interconnected fractures, with the matrix providing most of the storage but negligible flow. Such aquifers have dual porosity and are more challenging to analyse. For instance, Theis (1967) stated "the type of aquifer study in which our homogeneous model of ground water flow is most grossly inadequate is that dealing with transport phenomena".

One possible way to improve understanding of groundwater processes in bedrock aquifers is to examine the fundamental assumptions made when treating aquifers as single-porosity porous media, to identify the circumstances where flow or transport may give poor predictions. Accordingly, the main question asked in this paper is to what extent the assumptions of the single-porosity archetype of Hubbert (1940) are valid. Three major assumptions will be subsequently addressed. The focus is on testing the assumptions in the most challenging situations, where the assumptions are most likely to be falsified. The most challenging situations are for transport rather than flow, and for bedrock aquifers rather than unconsolidated sediments.

Can bedrock aquifers be treated as insoluble and chemically inert?

Dissolution processes in bedrock aquifers

The assumption by Hubbert (1940, p. 788) that the aquifer framework can be treated as insoluble and chemically inert usually works well for unconsolidated sediments, which are dominated by quartz and clay minerals. These minerals are ultimately derived from the weathering of igneous rocks, and have very low solubilities and dissolution rates. Rocks, however, have a wide range in solubilities, and many important rock-forming minerals have much higher dissolution rates than quartz and clay minerals (Fig. 1). This figure also shows how the more permeable rocks are composed of minerals with higher dissolution rates.

The high correlation between permeability and dissolution rate is most easily understood from the many studies on limestone rocks and aquifers. Theis (1936, p. 43–44) gave an early description of the process in limestone:

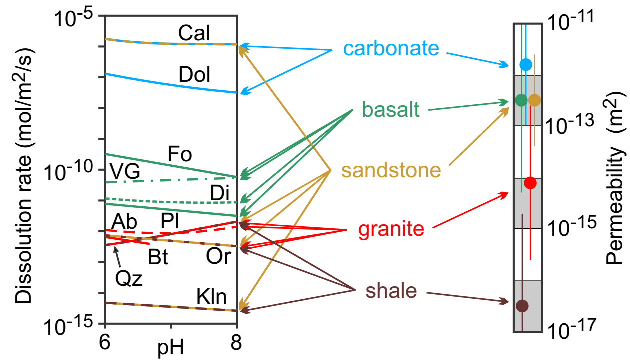

Fig. 1 Correlation between dissolution rates of major rock-forming minerals with permeability (k) of major lithologies. Lines link rocks to the minerals that comprise >5% of each rock. Dissolution rates are from Morse and Arvidson (2002) and Brantley et al. (2008). Permeability values are from Gleeson et al. (2011), with one standard deviation bars of log k shown. The minerals are calcite (Cal), dolomite (Dol), forsterite (Fo), volcanic glass (VG), diopside (Di), Ca–Na plagioclase (Pl), albite (Ab), orthoclase (Or), biotite (Bt), quartz (Qz), and kaolinite (Kln). The dissolution rate of forsterite is assumed to be representative for olivine, diopside for pyroxene, and kaolinite for clay minerals. Adapted from Worthington et al. (2016)

> Solution .... tends to increase the size of the openings through which the water moves. The openings that were originally somewhat larger tend to become much larger in proportion than the smaller ones, because of the more rapid circulation of water through them. As they increase in size the passage of water through them becomes more easy, so that they tend by diversion to reduce the flow of water through the smaller openings. By the process of piracy thus set up, the underground drainage is integrated into large trunk streams with branches, which in turn have smaller tributaries, as in a surface drainage system. Larger vertical channels as well as horizontal channels develop and give rise to sink holes at the surface.

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In modern terminology, Theis (1936) was describing feedbacks between flow and solution that results in the creation of self-organized channel networks. Extensive caves represent an end-member where these processes have operated most efficiently. A detailed understanding of this feedback process has been developed in the last 50 years. The discovery of the nonlinear kinetics of calcite (Berner and Morse 1974) was an important first step. The dissolution rate F can be expressed as

$$F = k\left(1 - c/c_{\mathrm{eq}}\right)^{n} \quad (1)$$

where k is the rate constant, c is the calcium concentration, c_{eq} is the equilibrium concentration of calcium, and n is the rate order, which near equilibrium has been found to vary from 4 to 11 (Eisenlohr et al. 1999). The nonlinear kinetics (Fig. 2) result in most dissolution taking place in the uppermost bedrock, due to low solute concentrations and low pH in precipitation, creating a highly weathered zone. Below this zone, the asymptotic approach to chemical equilibrium means that there is still finite dissolution taking place at greater depths. The nonlinear kinetics in Eq. (1) has been incorporated into reactive transport models that track both changes in solute concentrations and changes in fracture aperture. These models show how the feedback between flow and dissolution creates channel networks and, in some limited circumstances, caves (Romanov et al. 2003; Dreybrodt et al. 2005; Worthington and Ford 2009; Kaufmann 2016).

The aforementioned discussion has focussed on limestone, and its congruent weathering can be expressed (Berner and Berner 2012) as

$$CO_2 + H_2O + CaCO_3 \rightarrow Ca^{2+} + 2HCO_3^- \tag{2}$$

This reaction is an acid/base neutralization reaction, rather than just simple dissolution in water. However, it is usually described in the hydrogeology literature as dissolution (or solution), reflecting the broad definition that dissolution is a reaction that converts solid minerals to solutes (i.e. ions and molecules in solution).

The dissolution of some silicate minerals is similar, being a function of CO_2 concentrations and also producing bicarbonate ions in solution. For instance, the dissolution of albite ($NaAlSi_3O_8$), a common plagioclase feldspar mineral, can be described (Berner and Berner 2012) as

$$\begin{aligned} 2CO_2 + 11H_2O + 2NaAlSi_3O_8 &\rightarrow 2Na^+ \\ &+ 2HCO_3^- + 4H_4SiO_4 + Al_2Si_2O_5(OH)_4 \end{aligned} \tag{3}$$

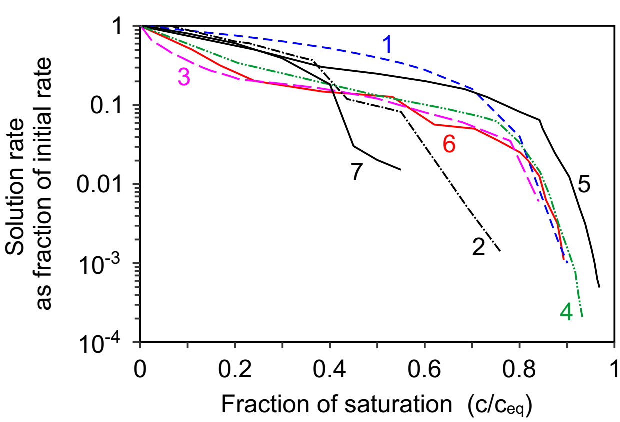

Fig. 2 Dissolution rates of carbonates as a function of saturation: **a** Iceland spar (*1* Plummer and Wigley 1976), limestone (*2* Berner and Morse 1974; *3* Plummer et al. 1978; *4* Eisenlohr et al. 1997; *5* and *6* Eisenlohr et al. 1999), and dolomite (*7* Herman and White 1985). Modified from Worthington (2015)

The dissolution of albite is partially congruent. The sodium, bicarbonate and silica wind up in solution, but the aluminium weathers to the solid clay mineral kaolinite, $Al_2Si_2O_5(OH)_4$. The clay may be deposited in situ, thus reducing permeability, and this appears to preferentially occur in the narrower fractures where velocities are lower. In the larger fractures, it appears to be more likely that the clay is carried away as a colloid and consequently that the overall permeability of the aquifer increases (Mayo et al. 2014). Furthermore, several common rock-forming silicate minerals do dissolve congruently, in particular those that lack the low-mobility elements iron and aluminium—examples include quartz and forsterite.

Lab and field data of the dissolution rates of silicates suggest that these may have a similar asymptotic approach to thermodynamic equilibrium to that shown by calcite (Fig. 2), though the absolute rates are substantially lower (Fig. 1; White and Brantley 2003). Well-documented examples of weathering, increasing the permeability in silicate rocks, include Lachassagne et al. (2011), Mayo et al. (2014), and Medici et al. (2018). Furthermore, iron oxide precipitates may be found where channels discharge into mine tunnels, also indicating weathering of the rock (Neretnieks 2006). Such iron precipitates reflect an oxidation/reduction reaction, where soluble Fe^{2+} in anoxic groundwater is oxidised to insoluble Fe^{3+}. Black et al. (2017) suggested that such channels may cross many fracture intersections without bifurcating, supporting the concept that weathering has increased channel apertures along continuous fracture pathways. The effect of weathering in enhancing permeability in silicate aquifers is documented much less than in carbonate aquifers, but this may be largely because its role is less frequently considered.

Dissolution is enhanced by high CO_2 concentrations (Eqs. 2 and 3). These are largely a function of the decomposition of organic matter, with higher concentrations being associated with higher rainfall and higher temperatures, which explains why the thickest saprolites are found in humid tropical areas (Strakhov et al. 1967). Similarly, such areas host the largest caves, both in carbonates (e.g. Waltham 2019) and in sandstones (Wray and Sauro 2017). It is also likely that the permeability of most rocks is generally higher in areas with warmer and wetter climates.

Correlation of chemical and physical factors with permeability

It has long been known that bedrock permeability varies with lithology, with most textbooks giving a figure or table showing the ranges (e.g. Freeze and Cherry 1979, p. 29; Hiscock and Bense 2021, p. 53)—for instance, shale and mudstone are generally regarded as aquitards and carbonate rocks as aquifers. A compilation of data from regional (>5 km) groundwater models found the former to be on average 50,000 times less permeable than the latter (Gleeson et al. 2011). In addition to differences in permeability due to lithology, there are also differences due to scale. The permeability of the matrix of bedrock aquifers is often several orders of magnitude less than that of fracture and channel networks (Price et al. 1993; Heppner et al. 2007; Worthington and Ford 2009; Ren et al. 2018). Consequently, local-scale measurements such as packer tests in wells usually have wide ranges of values, with intervals intersecting flowing fractures having much higher values than intervals with no open fractures. Schulze-Makuch et al. (1999) found that permeability increases with the scale of measurement up to aquifer volumes of 10^3–10^6 m^3, beyond which there is minimal change with scale. Consequently, the minimum length scale of 5 km in permeability values of Gleeson et al. (2011) suggests that the values are likely to be representative of the respective aquifers.

There has been little discussion in the literature of why there are large lithology-related differences in permeability. To understand the dominant processes, the correlation between fundamental physical and chemical properties and the permeability of different lithologies was investigated, using the data compiled by Gleeson et al. (2011). They divided bedrock into carbonate and silicate rocks, with the latter being subdivided into crystalline, volcanic, and coarse-grained and fine-grained siliciclastic sedimentary rocks. For simplicity, the silicate rocks are referred to here by the representative rocks granite, basalt, sandstone and shale, respectively. From the analysis of Dürr et al. (2005), these five classes provide the surficial lithology of over 70% of the land surface of the Earth that is not covered by ice (carbonate 10%, crystalline 26%, volcanic 7%, sandstone 8%, shale 19%), whereas unconsolidated sediments occupy the remaining 30% of the land surface.

The water-quality-data sets comprised 3,590 samples from wells or springs in the UK and USA used as potable water sources (Worthington et al. 2016). To calculate the total dissolved solids (TDS) derived from weathering, corrections were made for the bicarbonate derived from atmospheric CO_2 (Eqs. 2 and 3) and for the average 10 mg/L for TDS in precipitation. Pollution has been calculated to account for 9.5% of TDS concentrations globally in rivers (Berner and Berner 2012); however, no correction was made for it in this analysis because it was not possible to calculate it from the data sets used, and it is relatively minor compared to the large differences in TDS between the different lithologies. Worthington et al. (2016) showed that the permeability of the five major classes of bedrock lithology has positive correlations with both solute concentrations and dissolution rates, but did not present correlations for the physical attributes that might influence permeability.

Results for both physical and chemical processes show that tensile strength, Poisson's ratio, dissolution rate, and TDS have the highest correlations with permeability (Fig. 3). However, the very low permeability of shale disproportionately affects the results. If shale is omitted, then the correlations of the four remaining lithologies give r^2 values of 0.012, 0.26, 0.60, and 0.91 for tensile strength, Poisson's ratio, dissolution rate, and TDS, respectively. This suggests that the two chemical factors are more important than any of the physical factors in determining the permeability of bedrock aquifers. These two factors have a cause and effect link: the contrasting dissolution rates in different lithologies (Figs. 1 and 3f) provide the cause, and the effect is the contrasting solute concentrations (Fig. 3g, h). The high correlations between permeability and both dissolution rate and TDS suggest that dissolution may be the most important factor in determining permeability in bedrock aquifers. However, in rocks with low dissolution rates such as shale, it seems likely that physical factors may prevail over chemical factors in determining permeability. Overall, fracturing provides the pathways through which most flow takes place, and then dissolution preferentially enlarges the apertures along the flowpaths with the greatest flow, with more dissolution in the rocks with higher dissolution rates. The data used for the regression analysis are given in Table S1 in the electronic supplementary material (ESM).

Limitations of treating bedrock aquifers as insoluble and chemically inert

The question asked at the beginning of this section is whether bedrock aquifers can be treated as insoluble and chemically inert. Figures 1 and 3 show that none of the common rock-forming minerals are insoluble or chemically inert; however, it has been found that bedrock aquifers can be treated as porous media in flow models but not in transport models (Scanlon et al. 2003). This complex issue will be addressed in more detail in the section 'Discussion and conclusions'.

Are permeability variations adequately explained by the terms heterogeneity and anisotropy?

Hubbert (1940, p. 788) began by considering an aquifer that was homogeneous and isotropic with respect to both permeability and porosity, and later considered heterogeneity and anisotropy. Aquifers with little spatial variation in permeability are described as homogeneous, while those with substantial spatial variations are described as heterogeneous. For instance, the Macrodispersion Experiment (MADE) site in Mississippi (USA) is composed of clay, silt, sand,

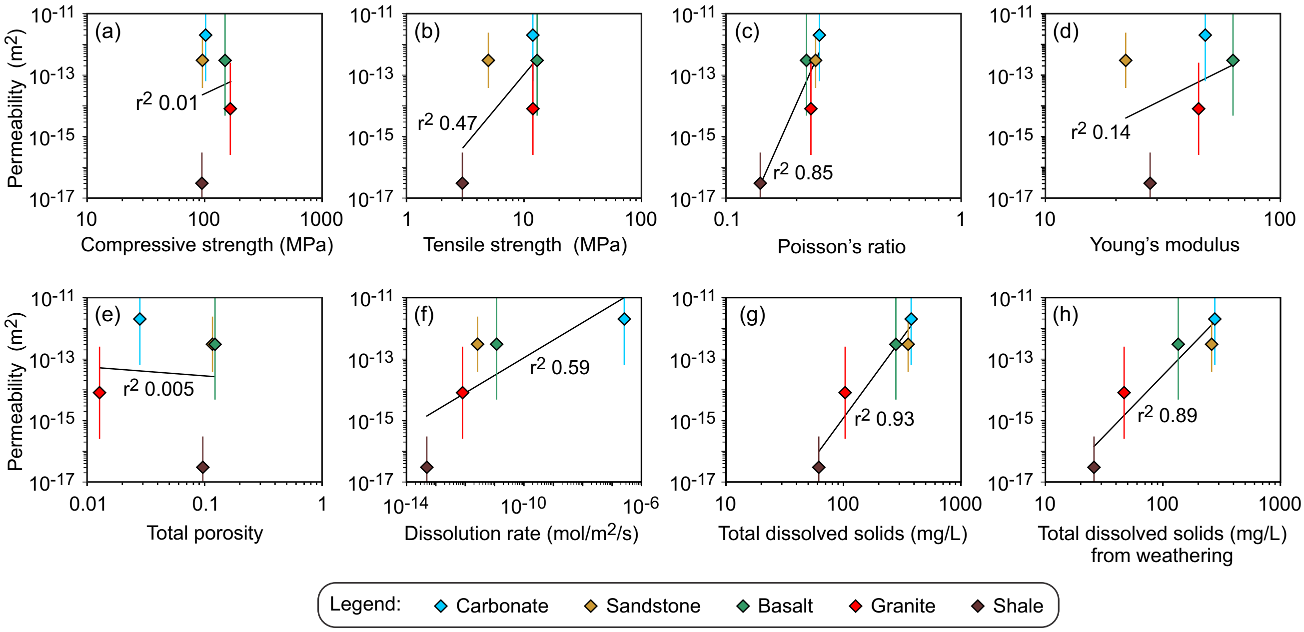

Fig. 3 Correlation of permeability in the five major bedrock lithologies with five physical and three chemical attributes of bedrock (permeability data from Gleeson et al. 2011, with one standard deviation bars of log *k* shown; compressive strength, tensile strength, Poisson's ratio and Young's modulus from Pollard and Fletcher 2005; total porosity from Wolff 1982; dissolution rate and total dissolved solids from Worthington et al. 2016)

and gravel, and has been described as highly heterogeneous (Zheng et al. 2011).

A common assumption in modelling studies is that there is random spatial variability of permeability (Freeze 1975; Rajaram 2016). However, the heterogeneity in bedrock aquifers is often very different from that in unconsolidated sediments, and a fundamental question is how to describe this heterogeneity—for example, the examples shown in Fig. 4 all display heterogeneity and anisotropy, and the aquifers might be described as highly heterogeneous and anisotropic. However, the heterogeneity and anisotropy are not random, but instead, there is preferential flow on distinct structures in every case. This preferential flow is referred to here as organized. In cases where there are feedbacks between flow and weathering that enhance the permeability, then the permeability is referred to as self-organized, as noted earlier in connection with the description of the process by Theis (1936).

The arrowed locations and the dark water-saturated areas on the cliffs in Fig. 4a–d all show how flow is often from a limited number of bedding planes, and also from limited parts of these bedding planes, indicating channelling of flow. This has long been described in the hydrogeology literature (Worthington 2013); however, the extent and interconnectivity of the preferential flow paths shown in Fig. 4a–d are unclear. One possibility is that the lateral extent of individual bedding planes with flow is short, and that such high-permeability fractures are not connected; however, such preferential flow may persist for distances of many kilometres (Muldoon et al. 2001; Schürch and Buckley 2002). Similarly, interflow zones in lavas (Fig. 4e, f) may also extend for many kilometres (Kiernan et al. 2003; Morin et al. 1993). Thousand Springs (Fig. 4e, f) discharges 19 m^3/s (Meinzer 1927) from the Eastern Snake Plain Aquifer (Idaho, USA), where 13 tracer tests from wells to three major springs over a mean distance of 1,890 m gave a mean velocity of 305 m/day (Farmer et al. 2014). These rapid velocities suggest that preferential flow paths in that aquifer are persistent over at least such distances.

The preferential flow paths in crystalline rocks (Fig. 4g) are tectonic fractures rather than depositional discontinuities, and the lateral persistence is less clear. However, fractures zones and faults in crystalline rocks may extend for substantial distances (e.g. Gascoyne 2004); furthermore, faults may be barriers, but may also act as flow paths for water vertically through hundreds of metres of lower-permeability strata (Burdon and Safadi 1963; Bense et al. 2013). The final two images in Fig. 4 show large channels, in caves in quartzite (Fig. 4h) and in limestone (Fig. 4i). Substantial dissolution is needed to create such large channels, and the aperture sizes are a function of the dissolution rate (Fig. 1), the chemical undersaturation (Fig. 2), discharge, and time. In the quartzite cave (Fig. 4h), the low dissolution rate is offset by the high rainfall in the area (~3,000 mm/year) and

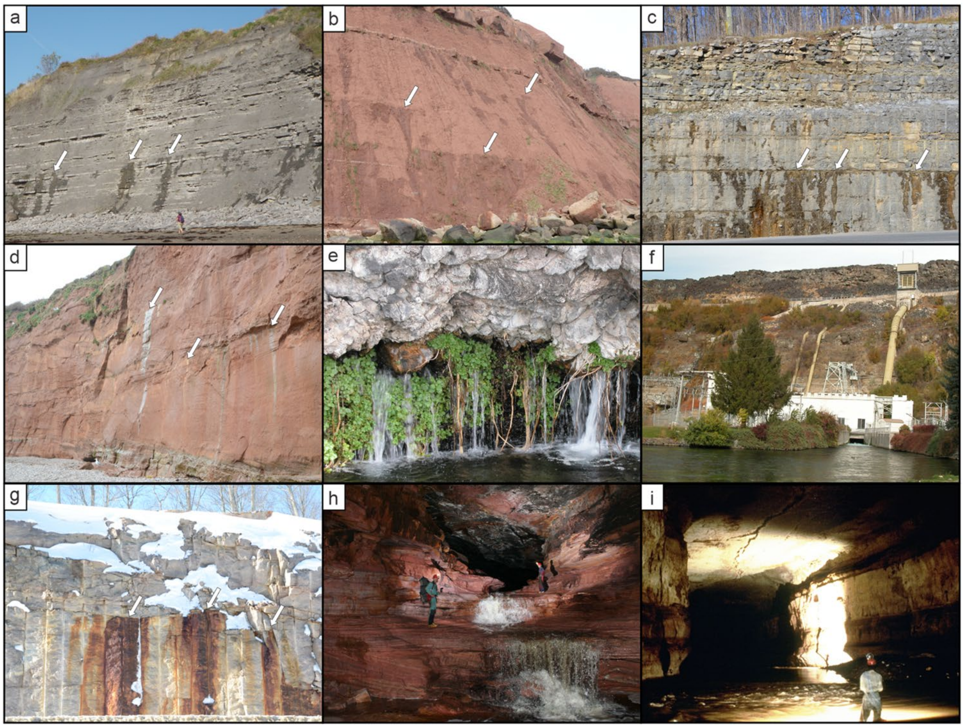

Fig. 4 Examples of preferential flow in bedrock aquifers: **a** Interbedded shale and limestone of Jurassic Blue Lias at Lyme Regis, UK; **b** Mudstones with minor sandstones of Permian Aylesbeare Group at Exmouth, UK; **c** dolostones of Silurian Lockport Formation at Hamilton, ON, Canada; **d** Triassic Otter Sandstone at Ladram Bay, UK; **e** Discharge from interflow zone in Pleistocene basalt lava at Thousand Springs, ID, USA; **f** Thousand Springs hydroelectric power station, showing 600-m-long collector system from discharge at interflow zone near the top of the cliff; **g** iron oxide staining from discharge from Proterozoic gneiss, Sundridge, ON, Canada; **h** waterfalls in Cueva Charles Brewer, Proterozoic orthoquartzite, Chimantá massif, Venezuela (courtesy Mladen Kuhta); **i** main river passage in Sof Omar Cave, Jurassic Antalo limestone, Ethiopia. Arrows (**a**–**d** and **g**) indicate channelling of water flow

the long time (~110 Ma) since the cave probably started forming in the Cretaceous (Mecchia et al. 2014; Wray and Sauro 2017). In the limestone cave (Fig. 4i), the main factor involved in creating the 20-m-wide river passage has been the high discharge of the river (several m^3/s), which drains a catchment area of 3,800 km^2 (Worthington 2004).

In many cases, the connectivity of preferential flow paths is unclear. If these paths are poorly connected, then the effective porosity of the aquifer could be close to total porosity and exceed 0.1. However, if tectonic processes have resulted in a network of well-connected open fractures, or if weathering has enhanced permeability, then effective porosity is like to be <0.01. Tracer testing usually shows effective porosity values in bedrock are <0.01, indicating that preferential flow is common in bedrock aquifers (Worthington 2022). Consequently, connectivity, organization and self-organization of preferential flow paths are important aspects of spatial variation of permeability, and useful terms to supplement heterogeneity and anisotropy in describing the spatial variation in permeability in bedrock aquifers.

Can bedrock aquifers be treated as continua for transport?

Hubbert (1940, p. 827) suggested that "The variation of most physical properties in space is continuous with distance". That is the case in homogeneous sand but is not the case in bedrock aquifers such as those shown in Fig. 4, which all have preferential flow paths. Nevertheless, it has been found that treating bedrock aquifers as single-porosity porous media usually works well for steady-state flow, as explained earlier. However, the situation is substantially different for transient flow and for transport (Theis 1967). The major practical difference between these cases is that calculations require a value for effective porosity.

There is often great uncertainty over the value of effective porosity in bedrock aquifers because it is rarely measured (using tracer tests), and it is unclear whether it should be represented by a value for total porosity (often >0.1) or by a lower value such as one for fracture porosity (10^{-5}–10^{-2}, Freeze and Cherry 1979, p. 408). Heterogeneous preferential flow is particularly important if it forms connected networks because such aquifers will exhibit dual-porosity rather than single-porosity behaviour for transport (Małoszewski and Zuber 1985). Preferential flow is common in all aquifers, with facies variation being important in sediments (Anderson 1989); however, it is more common in bedrock aquifers due not only to facies variation but also to fracturing and weathering.

Flowmeter data provide a useful way to determine the frequency of preferential flow. A literature search yielded data from 96 bedrock wells (Worthington et al. 2016). Results show that all measurable flow in most wells is from fractures, with only a small fraction having some or all flow from the matrix (Fig. 5a). And even in the wells where some of the flow was from the matrix, there was still preferential flow because facies variation resulted in only part of the saturated column yielding water (Leaf et al. 2012; Wilson et al. 2001; Paillet 2004).

Tracer tests are considered the best way to determine effective porosity (Zheng and Bennett 2002). A literature search yielded 121 such values (Worthington 2022). Results show a wide range, but most values are <0.01, suggesting that fracture flow dominates in most cases (Fig. 5b). These values are much lower than values often recommended for transport models—for instance, Haitjema and Anderson (2016) suggested that effective porosity is usually between 0.1 and 0.4, and Woessner and Poeter (2020, p. 14) suggested that total porosity can be used to represent effective porosity in most rocks with well-connected pores and fractures. The large difference in values between these two perspectives suggests that tracer tests should be used much more frequently in bedrock aquifers in cases where travel times are of interest. There have been several studies where post-audits of source protection zones have shown that measured effective porosities using injected tracers are much lower than earlier studies had assumed—examples include a Jurassic limestone aquifer in the UK (Foley et al. 2012), a limestone and dolostone aquifer in Canada (Worthington et al. 2012), and a Cretaceous chalk aquifer in the UK (Maurice et al. 2021). Environmental tracers provide a continuous input signal to aquifers, and ages of these tracers reflect not only fast flow along fractures but also diffusion into the matrix, and often have ages of years to decades. The ratio of the age of an environmental tracer to the age of an injected tracer is defined as the diffusion retardation factor (Małoszewski and Zuber 1985; Zuber et al. 2011). This is useful for understanding how contaminants can follow preferential flow paths and arrive quickly at receptors, while matrix diffusion results in very long persistence of the contaminant in bedrock aquifers. Few studies have measured ages from both environmental and injected tracers, and only 15 values could be found in the literature where groundwater ages for both tracers had been measured at the same monitoring location. In most cases, these monitoring locations were large springs and the tracer tests were over distances >5 km (Table S2 of the ESM). If an aquifer behaves as a single-porosity porous medium for transport then the diffusion retardation factor would be one; however, all results had much higher values, demonstrating the dual-porosity structure of all of these aquifers (Fig. 5c).

The presence of bacteria in routine well tests provides a fourth indicator of preferential flow. Real estate transactions in New Jersey (USA) require testing of domestic wells, and Atherholt et al. (2013) compiled data from such tests in 24,906 bedrock wells and 25,382 wells in unconsolidated sediments. Results show much higher rates of bacteria detection in the former, both for Escherichia coli (or faecal coliform) and for total coliforms (Fig. 5d). Similar findings of much more frequent contamination of bedrock wells were also made in a study by Embrey and Runkle (2006) for a much wider area in the USA. All four types of measurement used in Fig. 5 show that preferential flow in bedrock aquifers is very common; thus, they cannot be treated as continuous single-porosity porous media for transport calculations.

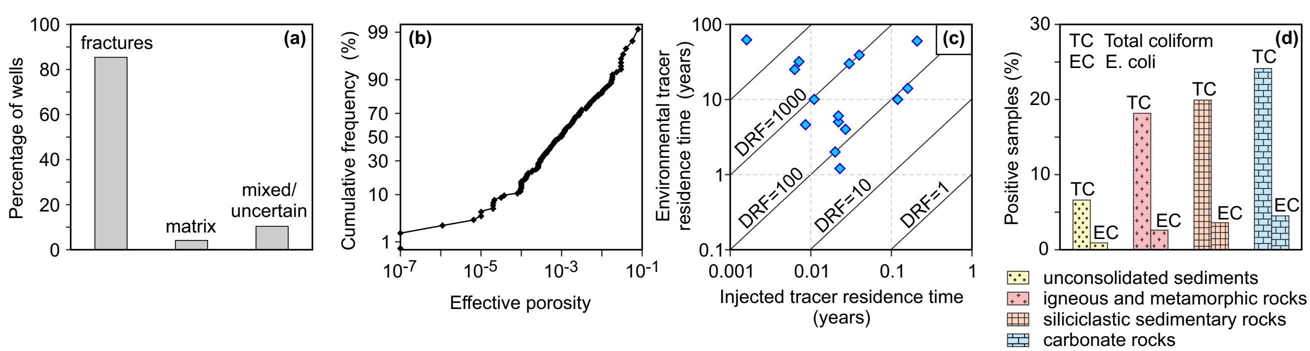

Fig. 5 Preferential flow data from bedrock wells: **a** flowmeter data from 96 wells, showing source of flow; **b** effective porosity in bedrock, derived from 121 calculations using tracer tests; **c** residence times from environmental and injected tracer tests in 15 aquifers, showing diffusion retardation factors (DRF); **d** percentage of 50,288 domestic wells in New Jersey positive for bacteria. Based on data in Worthington et al. 2016 (**a**); Worthington 2022 (**b**); Table S2 of the ESM (**c**); Atherholt et al. (2013) (**d**)

Table 1 Question and answer summary for three major assumptions of Hubbert (1940)

Question	Answer for generation of permeability	Answer for flow	Answer for transport
Can bedrock aquifers be treated as insoluble and chemically inert?	No - chemical weathering often substantially enhances permeability	Yes - however a conceptual understanding of spatial variation in permeability is invaluable	No - weathering often creates self-organized channel networks
Are permeability variations in bedrock aquifers adequately explained by the terms heterogeneity and anisotropy?	No - organization and self-organization result in preferential flow	No - organization and self-organization are common	No - organization and self-organization are common
Can bedrock aquifers be treated as single-porosity continua?	No - permeability is enhanced along preferential flow paths	Yes, at a large enough scale (typically > 100 m)	No - preferential flow results in dual-porosity behaviour

Discussion and conclusions

Hubbert (1940) offered a template for studying the hydraulics of groundwater flow as continuous single-porosity porous media, which has provided the foundation for modern hydrogeology. For instance, Deming (2002, p. 319) stated that "it established the physical principles upon which hydrogeology was to develop"; however, Hubbert (1940) only considered flow and did not consider transport. The preceding analysis has shown that three of the assumptions that Hubbert made are of limited applicability when considering transport in bedrock aquifers. The results are summarized in Table 1.

A major problem in understanding bedrock aquifers is that weathering often enlarges many fractures, creating a fracture and channel network where most wells intersect enlarged fractures or channels. The presence of such dense networks of solutionally enlarged fractures results in well-defined water tables. Treating such aquifers as single-porosity porous media usually works well for steady-state flow, even where there are extensive caves (Scanlon et al. 2003; Worthington 2009). Because many bedrock aquifers can be adequately characterized as porous media for steady-state flow, it may be mistakenly thought that these aquifers also behave as porous media for transport. However, the presence of preferential-flow networks of fractures and channels means that treating such aquifers as dual-porosity media will improve characterization and understanding of transport.

Theis (1967) was concerned that standard porous medium models do not adequately deal with transport. He stated "I consider it certain that we need a new conceptual model, containing the known heterogeneities of the natural aquifer, to explain the phenomena of transport in ground water". Theis did not elaborate on what constitutes "known heterogeneities", but the text quoted earlier from Theis (1936) indicates that it would include recognition of at least some aquifers as having self-organized preferential-flow networks.

The assumptions listed in Table 1 have often been accepted as the norm, and contribute to the standard assumptions in hydrogeology textbooks (e.g. Freeze and Cherry 1979; Anderson et al. 2015; Dassargues 2019; Hiscock and Bense 2021; Fetter and Kreamer 2022). Consequently, theoretical studies have often focussed on the groundwater hydraulics of idealized aquifers that lack preferential flow. Site characterization of bedrock aquifers that make this assumption may then fail to include tests such as tracer tests and flowmeter profiling that will indicate whether aquifers have slow seepage flow through the matrix or much more rapid preferential flow in fractures and channels. Thus, dual-porosity aquifers with rapid preferential flow may be mischaracterized as single-porosity aquifers with only slow seepage flow. This demonstrates how hydrogeology is more than just groundwater hydraulics, which was the focus of Hubbert (1940), and that groundwater geology is an important aspect when considering transport in bedrock aquifers.

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Declarations

Conflict of interest The author states that there is no conflict of interest.

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