

Karst hydrogeology of the Chalk and implications for groundwater protection



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Abstract: The Chalk is an unusual karst aquifer with limited cave development, but extensive networks of smaller solutional conduits and fissures enabling rapid groundwater flow. Small-scale karst features (stream sinks, dolines, dissolution pipes, and springs) are common, with hundreds of stream sinks recorded. Tracer velocities from 27 connections between stream sinks and springs have median and mean velocities of 4700 and 4600 m d⁻¹. Tests to abstraction boreholes also demonstrate very rapid velocities of thousands of metres per day. Natural gradient tests from observation boreholes have rapid velocities of hundreds of metres per day. There is strong geological control on karst with dissolution focused on stratigraphical inception horizons. Surface karst features are concentrated near the Paleogene boundary, or where thin superficial cover occurs, but rapid groundwater flow is also common in other areas. The Chalk has higher storage and contaminant attenuation than classical karst, but recharge, storage and flow are influenced by karst. Point recharge through stream sinks, dolines, losing rivers, vertical solutional fissures, and soakaways enables rapid unsaturated zone flow. Saturated zone networks of solutional fissures and conduits create vulnerability to subsurface activities, and enable long distance transport of point source and diffuse pollutants, which may be derived from outside modelled catchment areas and source protection zones.

Globally around 20–25% of people rely on karst groundwater for supply (Ford and Williams 2007). The term karst is applied to soluble rocks with landscapes characterized by caves, large springs, sinking streams, dry valleys and dolines (Ford and Williams 2007). Hydrogeologically, karst results in the solutional enlargement of fractures to form larger self-organized conduit networks (Worthington and Ford 2009) enabling rapid groundwater flow, often over long distances (>1 km).

The Upper Cretaceous Chalk Group of NW Europe is a highly transmissive carbonate aquifer (Downing *et al.* 2005). It provides public water supplies to millions of people; agricultural and industrial water resources; sustains rivers and wetlands; and is an important habitat for groundwater invertebrates (Maurice *et al.* 2016a). It is also an important example of the least well-understood type of karst aquifer – those in which cave development is limited. There is solutional development of fractures and stratigraphical discontinuities, forming fissures and conduits enabling rapid groundwater flow, but few enterable caves. Surface karst features are present, but often subdued; with dry valleys the only common

large scale karst landform. The Chalk is often regarded as non-karstic because of the lack of caves and the subtle nature of the surface karst features. In this paper we use the term ‘fissure’ for solutionally enlarged fractures with a generally planar cross-sectional shape and ‘conduit’ for linear solutional voids which are usually circular or elliptical in cross-section, whilst ‘caves’ are conduits large enough for humans to enter.

Decades of research have led to a good understanding of Chalk hydrogeology (Price *et al.* 1993; Allen *et al.* 1997). The bedrock matrix has high porosity of *c.* 35% (Bloomfield *et al.* 1995), but drainable porosity in the unconfined aquifer is much lower at *c.* 0.5–5% (Lewis *et al.* 1993). Matrix permeability is very low (mean 6.3×10^{-4} m d⁻¹) due to the small pore throats (Allen *et al.* 1997). Fractures are frequent with spacings of *c.* 0.05 m in the upper weathered zone (Price *et al.* 1976); and *c.* 0.1 to 0.5 m below (Bloomfield 1996; Zaidman *et al.* 1999). This dense fracture network has a hydraulic conductivity of *c.* 0.1 m d⁻¹, and a transmissivity of *c.* 20 m² d⁻¹; but >90% of pumping tests show higher transmissivity than this, which comes from a

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small number of solutionally enlarged fractures (Price 1987; Allen *et al.* 1997; MacDonald and Allen 2001).

Studies have demonstrated the presence of surface karst and rapid groundwater flow in the Chalk (e.g. Atkinson and Smith 1974; Rodet 1985; Banks *et al.* 1995; MacDonald *et al.* 1998; Massei *et al.* 2002; Maurice *et al.* 2006; Edmonds 2008; Gombert *et al.* 2010; El Janyani *et al.* 2014), but there is limited understanding of the extent of karst and the implications for chalk hydrogeology. The aims of this paper are to (1) present evidence of karst from newly compiled data for the Chalk of southern and eastern England; (2) provide new conceptual understanding of Chalk karst hydrogeology; and (3) discuss the implications for groundwater protection. The findings can be applied to the Chalk elsewhere in northern Europe and other similar aquifers.

Evidence of karst in the Chalk in England

In this section recent data compilation and literature review are used to present evidence for karst and rapid groundwater flow in the Chalk in England. The geomorphological evidence for karst includes dolines, dissolution pipes, stream sinks, large springs, caves, and smaller conduits and fissures. Hydrogeological evidence includes rapid velocities proved by tracer tests, high transmissivity in pumping tests, and the presence of short residence time indicators in boreholes (i.e. substances which degrade within the subsurface within short timescales).

Dolines and dissolution pipes

Dolines are surface depressions caused by karstic solutional processes (Waltham *et al.* 2005). Two main groups occur; those caused by subsurface dissolution and subsequent downward gravitational movement of the overlying material (bedrock or superficial deposits); and those generated by differential corrosional lowering of the ground surface, focused on higher permeability zones in the bedrock.

Surface depressions are very common throughout the Chalk of southern and eastern England, especially where there is a thin cover of Paleogene or superficial deposits, in particular the Clay-with-Flints. These surface depressions can be identified from digital terrain models and LiDAR, topographical maps, or are recorded on geological maps. However, many are pits dug for agricultural purposes, or clay or gravel extraction; or depressions caused by collapsed mine workings (Edmonds 2020).

Natural dolines do occur in the Chalk. Sperling *et al.* (1977) recorded densities of up to 157 per square kilometre in Dorset. Most are likely to be suffusion dolines formed by gradual subsidence rather

than collapse dolines, although some dropout failures do occur (McDowell *et al.* 2008). The largest are up to 21 m deep and 86 m in diameter (Sperling *et al.* 1977, 1979; Waltham *et al.* 1997). Most Chalk dolines are small and shallow (less than 2–3 m deep), although those associated with stream sinks close to the Paleogene boundary may be deeper (Fig. 1a). Some dolines away from the Paleogene margin may have originally formed as stream sinks, which became relict as the landscape evolved.

It is often not possible to determine whether a surface depression is a karstic doline, an old chalk or clay pit, or a combination (Prince 1964; Foley 2017). Geophysical surveys (Jeffrey *et al.* 2020) and historical maps and documents may help. Irregular shapes suggest excavations, although these may have been sited on natural karst depressions. Circular or oval-shaped surface depressions appear similar to karst dolines but may be ploughed-in excavated pits. Relying on remote sensing methods without considering land-use history and geomorphological evolution may result in the over-estimation of doline numbers.

Subsurface 'dissolution pipes' are sediment-filled voids created by dissolution, which normally have no surface expression, and are sometimes viewed as a form of buried doline (Waltham *et al.* 2005). They are revealed by engineering projects or exposed in quarries and cliffs, and extend to depths of <1 to >50 m below the surface (Fig. 1b). Different morphologies have been described including basin-shaped features, vertical pipes and horizontal sediment-filled seams (Thorez *et al.* 1971). Densities can be up to several hundred per hectare (Lamont-Black and Mortimore 1999), and they can coalesce to form areas of very irregular rock-head. High densities are likely to occur wherever thin superficial deposits overly the Chalk (Gibbard *et al.* 1986; Worsley 2016; Farrant *et al.* 2017, 2021a).

Stream sinks and river losses

Stream sinks (also known as swallow holes) are places where water from surface streams enters the aquifer directly (Fig. 1c–e). Three main groups occur: where water flows off the overlying lower permeability Paleogene deposits onto the Chalk; along losing stretches of rivers crossing the Chalk outcrop; and streams draining areas of low permeability superficial deposits such as glacial till. Karstic stream sinks are especially likely to develop where adjacent strata produce acidic soils enabling runoff with lower pH and enhanced dissolution of the Chalk (MacDonald *et al.* 1998).

Hundreds of stream sinks have been identified (Fig. 2a, data from the British Geological Survey). These were identified from topographical maps, individual catchment surveys (e.g. Maurice 2009; Farrant *et al.* 2017), geological mapping, or

Chalk karst hydrogeology



Fig. 1. Karst features in the Chalk: (a) doline with stream sink; (b) dissolution pipes; (c, d, e) stream sinks; (f, g) caves (photographs f and g courtesy of Terry Reeve).

hydrogeological memoirs (e.g. Whitaker 1921). Comprehensive surveys have not been carried out in all areas, and more may be present. Most occur in southern England, where small streams draining the Paleogene Lambeth Group flow onto the Chalk (Walsh and Ockenden 1982; Banks *et al.* 1995; Maurice *et al.* 2006). Highest densities occur where the Paleogene is dominated by clay-rich facies. Where the basal Paleogene is sand-rich, significant recharge

may occur directly into the Chalk without the generation of surface streams resulting in fewer stream sinks, but more dissolution pipes and dolines developed beneath the cover.

Stream sinks are rare in northern England and East Anglia where the Paleogene cover is absent although small ephemeral, ill-defined sinks are locally common where streams flow off overlying glacial till deposits onto the Chalk.

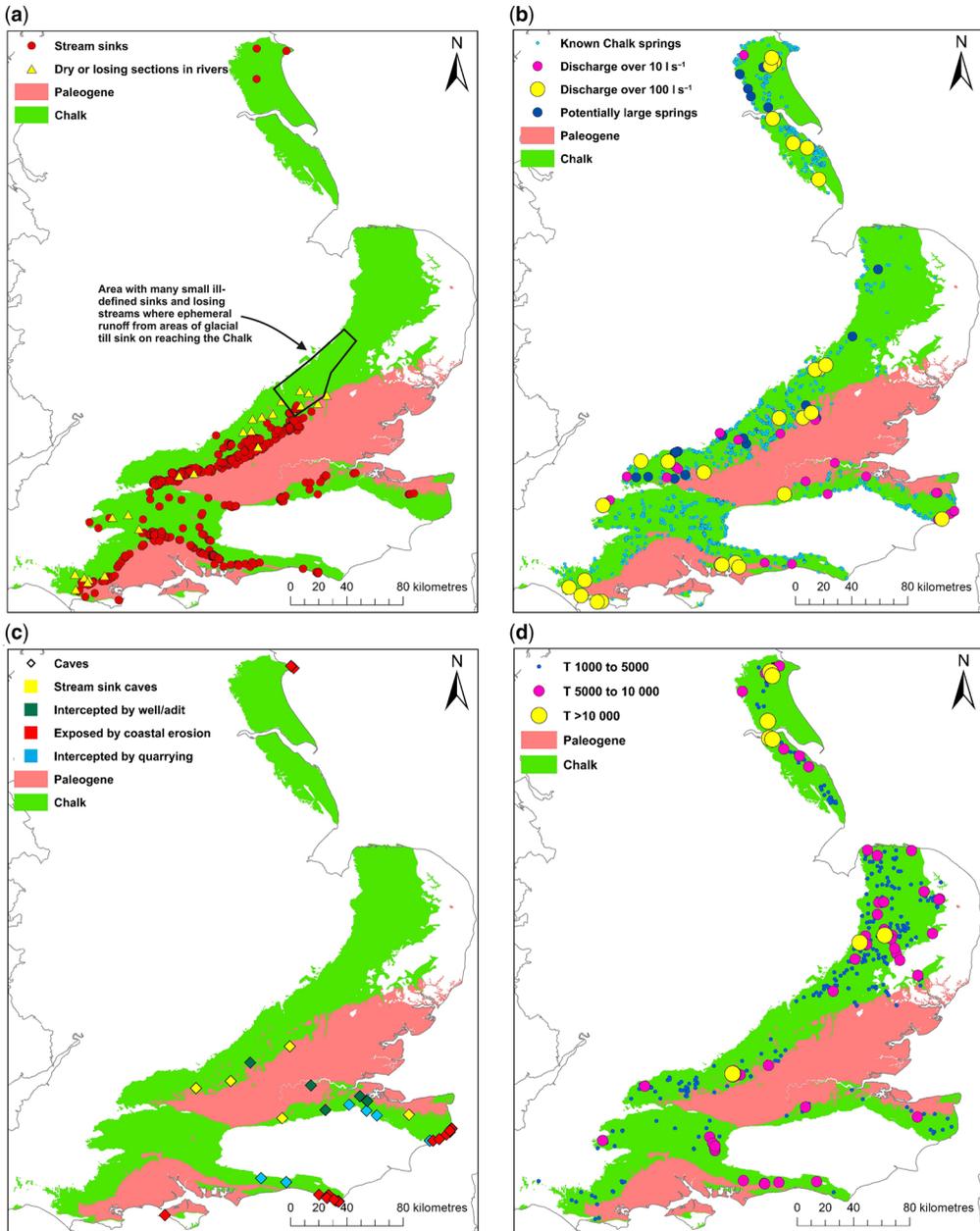


Fig. 2. Distribution of karst, see text for data sources: (a) stream sinks; (b) springs; (c) caves; (d) high transmissivity (T).

Few of the streams that sink into the Chalk have been gauged. Water End in Hertfordshire in southern England is the largest (Waltham *et al.* 1997), with an average flow of 80 l s^{-1} , and capacity for 1000 l s^{-1} before overflowing (Walsh and Ockenden 1982). Price (1979) reports measured flows of 0.5 to 85 l s^{-1} at eight stream sinks in the South Downs. Field

observations from southern England (Maurice 2009; Farrant *et al.* 2017, 2018) suggest that most stream sinks are fed by small, ephemeral streams, with flashy responses to rainfall and maximum flows likely to be *c.* 1 to *c.* 10 l s^{-1} , although some have small perennial flows from Paleogene springs.

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Chalk stream sinks often have multiple sink points, sometimes over distances of hundreds of metres. The sink point varies with water sinking progressively further upstream in lower flow conditions. In many, water sinks through sediment into the Chalk below. The sink points change over time as holes become blocked with sediment and new holes open up. Some streams completely sink into large surface depressions or blind valleys, whilst others overflow to other sinks or into other stream catchments during high flow. Many have been artificially modified, used as field drains, or bypassed.

Many larger rivers on the Chalk have substantial losses (e.g. [Grapes et al. 2005](#); [Griffiths et al. 2006](#); [Allen and Crane 2019](#); [Sefton et al. 2019](#)). Such river flow loss is typical in classically karstic aquifers ([Bonacci 1987](#)), and where river losses are not due to direct surface water abstraction, they are indicative of recharge to groundwater. There is no national dataset on Chalk rivers with dry or losing sections, but they are common and some examples are shown as yellow triangles in [Figure 2a](#). Some have dry sections, sometimes several kilometres in length, between upstream and downstream flowing sections, including 8 in the Wessex basin ([Allen and Crane 2019](#)) and 11 in the Chilterns ([Sefton et al. 2019](#)). Some develop classic sinkholes in their riverbeds, such as the River Mole in Surrey where $c. 120 \text{ l s}^{-1}$ water sinks into multiple swallow holes ([Fagg 1958](#)); and the Gypsy Race in Yorkshire ([Farrant et al. 2016](#)).

Springs

Springs are common in the English Chalk ([Fig. 2b](#), data from the British Geological Survey and

Environment Agency). Springs provide evidence of karst processes within the Chalk as concentrated outflows are likely to come from fissures and conduits; and the larger the flow, the more extensive the karstic networks are likely to be. Flow data are sparse but those with known discharge $>10 \text{ l s}^{-1}$, together with potentially large springs (observed in the field or used in tracer tests but without flow data) are shown in [Figure 2b](#). The largest Chalk springs are the Bedhampton Springs in southern England, with a discharge of up to $c. 2000 \text{ l s}^{-1}$, and other significant examples occur ([Table 1](#)). However, the Chalk aquifer has been heavily modified by abstraction and many spring discharges are substantially lower than they would naturally have been, and some springs no longer flow ([Whitaker 1921](#); [Day 1996](#); [Adams 2008](#)).

Most springs occur at the base of the Chalk escarpment or in the lower reaches of dip slope dry valley networks. Scarp springs are generally small ([Allen et al. 1997](#)) with a few exceptions of $>100 \text{ l s}^{-1}$ (e.g. [Wendover Springs](#), [Table 1](#)). Substantial perennial springs are very common in the lower reaches of the dip slope valleys. The exact locations of the active headwaters vary with discharge and may migrate several kilometres up-valley, giving rise to seasonal flows known as winterbournes fed by ephemeral springs ([Day 1996](#); [Allen et al. 1997](#)). For example in southern England, [Lynchwood Springs](#), 7 km upstream of the perennial head of the River Lambourn, vary from 0 to 690 l s^{-1} and reactivate rapidly ([Grapes et al. 2005, 2006](#)). Rapid activation of ephemeral springs occurs when the flow capacity of a karstic conduit or fissure system feeding a perennial spring is

Table 1. Examples of significant Chalk springs ($>200 \text{ l s}^{-1}$)

| Spring | Flow (l s^{-1}) | Reference |
|---|----------------------------|--|
| Bedhampton and Havant, Hampshire | 600–2000 | Allen and Crane (2019) |
| Blue Pool, Berkshire | ~200 | Maurice et al. (2006) |
| Wendover, Buckinghamshire | 17–255 | UK National River Flow Archive 2020, https://nfra.ceh.ac.uk |
| Chadwell Spring, Hertfordshire | 25–330 | Whitaker (1921) |
| Lynchwood Springs, Hertfordshire | 0 to 690 | Grapes et al. (2006) |
| Bellau springs, Lincolnshire | 210 | BGS records |
| Tetney Blow holes (artesian spring), Lincolnshire | 245 | BGS records |
| Wellbeck Spring, Lincolnshire | 340 | BGS records |
| Springs at Barrow upon Humber, Lincolnshire | 540 | BGS records |
| Melbourn Springs, East Anglia | 230 | Whitaker (1922) |
| Shepreth Springs, East Anglia | 210 | Whitaker (1922) |
| Watercress Farm spring, Surrey | 100–540 | BGS records |
| Fishbourne Springs, West Sussex | 150–400 | Jones and Robins (1999) |
| Arish Mell, Dorset | 200–690 | Houston et al. (1986) |
| Winterbourne Abbas, Dorset | 460 | Casey and Ladle (1976) |
| Sutton Poyntz, Dorset | 500 | Limbrick (2003) |
| Waterston House, Dorset | 350 | Webb and Zhang (1999) |

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exceeded and a higher pathway, previously dry, is flooded. Examples of such overflow springs are analysed by Smart (1983) and Smith (1979). Some Chalk springs become turbid following rainfall (Fig. 3), indicating connectivity with karstic stream sinks (Codrington 1864; Banks *et al.* 1995).

However, although Chalk springs share these characteristics with springs in highly karstic aquifers, Chalk springs appear to differ significantly in terms of baseflow. In fully karstic aquifers, large-scale conduits carry a large proportion of recharge directly from stream sinks to springs with a rapid response to rainfall (Ford and Williams 2007). Although there are few data on Chalk spring discharge, spring-fed Chalk rivers are comprehensively monitored and have a high baseflow index, generally of >0.9 (UK National River Flow Archive 2020, <https://nfra.ceh.ac.uk>). A comparison by Marsh *et al.* (2000) indicates less variability of flow in Chalk-fed rivers than on any other rock type. These data suggest that Chalk springs are supplied by longer-term storage than springs in classical karst, and that they are not predominantly supplied by rapid flow from surface stream sinks.

Caves and conduits

Small solutional conduits of *c.* 5 to 20 cm in diameter appear common in the Chalk, exposed in quarries and cliffs, and observed in images of borehole walls (Waters and Banks 1997; Schurch and Buckely 2002; Maurice *et al.* 2012; Farrant *et al.* 2017, 2021a). In coastal cliffs conduits are often localized on certain stratigraphical discontinuities, such as marl seams, hardgrounds or sheet flints. For example, along the East Sussex coast densities of more than 20 per kilometre stretch of cliff have been observed on favourable horizons (Farrant *et al.*

2021a). Some contain sediment infills indicating a link to the surface, sometimes >50 m above.

In addition to discrete conduits, some stratigraphical horizons carry anastomosing networks of small-scale dissolutional conduits 1–5 cm in diameter, termed ‘dissolution tubules’ (Lamont-Black and Mortimore 2000). These form a sponge-work zone up to 1 m thick above the discontinuity, sometimes extending laterally for hundreds of metres. They probably formed by mixing-corrosion where waters of different chemical composition mixed, producing under-saturation with respect to calcite (Dreybrodt *et al.* 2010; Farrant *et al.* 2021a, b).

Compared to most karst, conduits large enough to be termed caves are rare in the Chalk. Nevertheless, at least 44 caves have been exposed by quarrying or coastal cliff retreat, intersected during adit and well construction, or revealed by stream sinks (Fig. 2c, locations estimated from papers cited in this section and in Table 2; and from T. Reeve, pers. comm. 2017). The longest is Beachy Head Cave with a mapped length of 354 m (Reeve 1981, 2021; Lowe 1992; Waltham *et al.* 1997), but most are less than 20 m (Table 2).

At most stream sinks the water sinks through sediment into the underlying Chalk, and it is not usually possible to see the bedrock or cavities below. At five sites the Chalk is exposed and it is possible to enter short sections of passage. One example at Warren Row in Berkshire comprises a short crawl, a 4 m high chamber containing a cascade, and a further small passage before the water disappears into an inaccessible conduit (Chelsea Speleological Society 1990; Reeve 2021). Most stream sinks are likely to have small caves beneath them, possibly partially sediment filled.

Chalk caves revealed by quarrying are dry relict passages formed when water tables were higher. Their karstic origin is indicated by tube-like passage



Fig. 3. (a) Chalk karst spring near stream sinks; (b) with turbidity following rainfall (photographs courtesy of the Environment Agency).

Table 2. Lengths of some explored caves in the Chalk in England

| Name | Location | Probable Chalk stratigraphical formation | How the cave was exposed | Length (m) | Reference |
|---|---------------|--|--------------------------|------------|--|
| Flamborough Head (part marine) | Yorkshire | Flamborough Chalk Fm | Coastal cliffs | 67 | Howson (2000); T. Reeve (pers. comm, 2017) |
| Robin Lythe's Cave | Yorkshire | Flamborough Chalk Fm | Coastal cliffs | 20–30 | T. Reeve (pers. comm., 2017) |
| Warren Row cave | Berkshire | Upper Seaford Chalk Fm | Stream sink | 20 | T. Reeve (pers. comm., 2017) |
| Waterend cave | Hertfordshire | Seaford Chalk Fm | Stream sink | <10 | T. Reeve (pers. comm., 2017) |
| Yattendon cave | Berkshire | Seaford Chalk Fm | Stream sink | <10 | Maurice (2009) |
| Stood cave | Kent | Lewes Nodular Chalk Fm | Adit | ~60 | Bradshaw <i>et al.</i> (1991); CSS (1973) |
| Knockholt cave | Kent | Lewes Nodular Chalk Fm | Well | 10 | Bradshaw <i>et al.</i> (1991) |
| Chatham cave | Kent | Lewes Nodular Chalk Fm | Well | ~10? | Bradshaw <i>et al.</i> (1991) |
| Blackheath sewer tunnel caves (3 caves) | London | Lewes Nodular Chalk Fm | Sewer tunnel | All <10 | Bradshaw <i>et al.</i> (1991); CSS (1973) |
| Lower Ensdon caves (3 caves) | Kent | Seaford Chalk Fm | Stream sinks | up to 10 | T. Reeve, Chelsea Speleological Society (1979, 2010) |
| Hope Point Cave, Kingsdown | Kent | Lewes Nodular Chalk Fm | Coastal cliffs | 25 | T. Reeve (pers. comm., 2017) |
| Flittermouse Hole | Kent | Lewes Nodular Chalk Fm? | Quarry | 6 | T. Reeve, Chelsea Speleological Society (1973) |
| Boxley Quarry Caves (2 caves) | Kent | Lewes Nodular Chalk Fm | Quarry | ~85 | T. Reeve, Chelsea Speleological Society (1990) |
| Colley Hill cave | Surrey | Seaford Chalk Fm | Subsidence | 9 | T. Reeve, Chelsea Speleological Society (1963, 1976, 1979) |
| Canterbury Cave, St Margaret's Bay | Kent | Lewes Nodular Chalk Fm | Coastal cliff | 110 | T. Reeve, Chelsea Speleological Society (1979) |
| Langdon Bay | Kent | Lewes Nodular Chalk Fm | Coastal Cliff | 20 | T. Reeve, Chelsea Speleological Society (1979, 2010) |
| Cave at St Margaret's Bay | Kent | Lewes Nodular Chalk Fm | Coastal cliff | 8 | T. Reeve (pers. comm., 2017) |
| Beachy Head Caves | Sussex | Lewes Nodular Chalk Fm | Coastal cliff | up to 354 | Reeve (1981); Lowe (1992) |
| Houghton Quarry Cave, Amberley | West Sussex | Holywell Nodular Chalk Fm | Quarry | 10–15 | T. Reeve (pers. comm., 2017) |
| Patrick's Rift | Sussex | Seaford Chalk Fm | Coastal cliff | 28 | Reeve (1979) |
| Seaford Head caves (several) | Sussex | Lewes Nodular Chalk Fm | Coastal cliff | up to 24 | T. Reeve, Chelsea Speleological Society (1979, 1992, 1997, 2012) |
| Frenchman's Hole (part marine) | Isle of Wight | | Coastal cliff | 120 | Reeve (1982) |

morphology and features such as scallops, that indicate turbulent flow. Others have been discovered during adit and well construction, for example at Strood in Kent, where 60 m of natural stream passages up to 3 m wide and 5 m high were discovered in 1879 (Chelsea Speleological Society 1973).

Many caves are exposed in coastal Chalk cliffs (Reeve 2021; Farrant *et al.* 2021a). At least 21 contain good evidence for karstic dissolution, including passage morphologies, scalloping and sediment infills (Fig. 1f, g). Active cliff retreat frequently reveals and erodes segments of cave passage, so the overall number recorded is likely to be an underestimate. Most coastal caves in SE England are developed on thin marl seams or sheet flints in the Lewes Nodular and Seaford Chalk formations, and were probably formed by mixing dissolution along these stratigraphical inception horizons (Farrant *et al.* 2021a, b). They are commonly associated with smaller dissolution tubules on the same horizons (Lamont-Black and Mortimore 2000). Further details of caves in the Chalk are provided by Reeve (2021).

Evidence of karst from boreholes

Successful Chalk abstraction boreholes and adits are supplied by a small number of solutional fissures or conduits (Price *et al.* 1977; Ward 1989), and are therefore themselves indicative of connected networks of voids formed by karst processes. The median transmissivity of 2100 Chalk borehole pumping tests (with bias as predominantly from successful boreholes) is $540 \text{ m}^2 \text{ d}^{-1}$, ranging from <1 to $>20\,000 \text{ m}^2 \text{ d}^{-1}$ (MacDonald and Allen 2001). It is likely that the higher the transmissivity, the more developed and extensive the karstic solutional networks are (Foley and Worthington 2021). Many Chalk boreholes have high transmissivities (Fig. 2d), with more than $5000 \text{ m}^2 \text{ d}^{-1}$ at 60 sites. Based on planar fracture equations outlined in Snow (1968) and Domenico and Schwartz (1990), transmissivities of $1000 \text{ s m}^2 \text{ d}^{-1}$ are likely to come from solutional fissures and small conduits with apertures of several centimetres, which is commensurate with the solutional conduits of up to 20 cm ('Caves and conduits' section) and solutional fissures observed in borehole images (Farrant *et al.* 2017).

Other indicators of rapid flow pathways at abstraction boreholes include the presence of coliforms, turbidity, or rapidly degrading pesticides (e.g. Lawrence *et al.* 1996; Farrant *et al.* 2017). At some abstractions, there is evidence of connectivity with rivers or the sea over many kilometres, indicating connected networks of solutional voids (Monkhouse and Fleet 1975; Atkinson and Smart 1981; Howard 1982; Foley *et al.* 2001). In the South

Downs, semi-diurnal fluctuations in salinity observed at boreholes several kilometres from the coast indicate rapid lateral flow of saline water (Jones and Robins 1999). There is also evidence from borehole studies that as in more classical karst aquifers, the Chalk aquifer can be multi-layered, with flowpaths developed at different depths which may be isolated from one another (Karapanos *et al.* 2020).

Tracer tests

Tracer tests from 55 locations in the Chalk of southern and eastern England (Fig. 4a) provide groundwater velocities for 97 individual connections (based on first arrival of tracer), proving very rapid flows over distances of up to 19 km (Table 3, Fig. 4b, c).

Velocities from 27 tracer tests between stream sinks and springs have median and mean velocities of 4700 and 4600 m d^{-1} , somewhat higher than the values of 1740 and 1940 m d^{-1} for 3015 tests in karst aquifers from 24 countries (Worthington and Ford 2009). The apparently higher velocities in the Chalk could be a consequence of the data compilation method. The test type reported appears similar: Worthington and Ford (2009) state that most are natural-gradient traces from stream sinks to springs. However, the Chalk velocities are all based on first arrival times, and where multiple tests were undertaken over a connection, the fastest velocity was used. If velocities from Worthington and Ford (2009) were based on the peak tracer concentration, or mean values were used, or some tests were not stream-sink to spring connections, then this could account for the apparently lower velocities. However, it is unlikely that the groundwater velocities in the Chalk are substantially lower than those in other karst aquifers, as velocities based on first arrival and peak tracer concentration times are not substantially different. For example, in a chalk test reported by Maurice *et al.* (2006) the first arrival velocity was 5700 m d^{-1} compared to 4700 m d^{-1} for the peak tracer concentration.

Comparison of the Chalk velocities with those from stream sink to spring connections in the classically karstic Carboniferous Limestone of the UK (e.g. Atkinson *et al.* 1973; Atkinson 1977; Smart *et al.* 1991; Banks *et al.* 2009; Maurice and Guilford 2011) suggests that Chalk velocities are similar. Atkinson (1977) reports a mean groundwater velocity of 6330 m d^{-1} from 48 connections between stream sinks and springs in the Carboniferous Limestone of the Mendip Hills, UK; with a range of 520 to $21\,200 \text{ m d}^{-1}$. This suggests that velocities in the highly karstic Carboniferous Limestone may be slightly higher than those in the Chalk. However, travel times from tracer tests in the UK Derbyshire Carboniferous Limestone reported by Gunn (1991)

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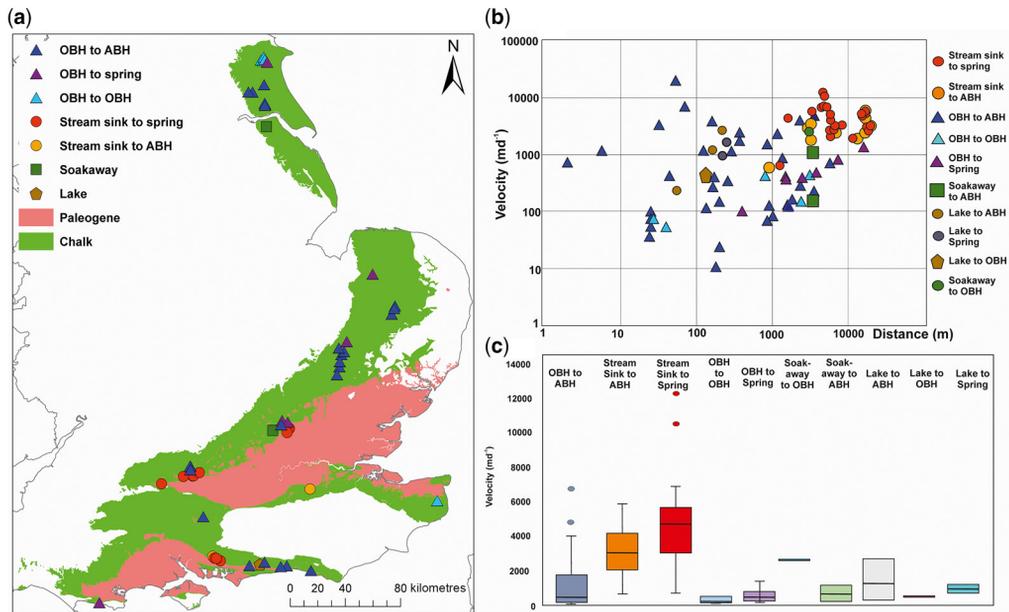


Fig. 4. Tracer tests, see Table 3 for data sources and numbers of tests. (a) Distribution of tests; (b) velocities with distance; (c) velocities in different test types (OBH, monitoring borehole; ABH, abstraction).

suggest lower groundwater velocities. Overall, it is clear that groundwater velocities from stream sinks in the Chalk are extremely rapid and are of the same magnitude (thousands of metres per day) as those in other karst aquifers.

Forced gradient tracer tests to Chalk groundwater abstractions also demonstrate very rapid velocities of thousands of metres per day, whilst natural gradient tests from observation boreholes demonstrate rapid flow of hundreds of metres per day to monitoring boreholes and springs (Table 3). High velocities occur over long distances (Fig. 4b), and are generally highest in tests from stream sinks (Fig. 4c). Velocities are comparable to those from 1893 dye tracer tests in the Chalk of northern France (Gombert *et al.* 2010; Table 4), carried out predominantly for water supply catchment delineation. Such rapid groundwater velocities are unequivocal evidence for karstification having produced open voids and channelled flow.

Tracer recoveries estimated for 42 connections in the English Chalk (Table 3) range from extremely low (<0.000007) to high (c. 100%). There are many low recoveries in the Chalk which may reflect the importance of diffusion from conduits/fissures into the matrix porosity (Barker and Foster 1981; Barker 1991, 1993; Barker *et al.* 2000; Matthias *et al.* 2009); and/or high dilution as a consequence of the complex multiple flowpaths supplying springs and boreholes. High attenuation/dilution may cause

failed tracer tests in the Chalk (e.g. Gombert *et al.* 2010; Maurice *et al.* 2012; see Table 4).

In addition, three unsaturated zone experiments in England applied tracer to the ground surface. These demonstrated rapid flow to between 15 and 38 m below the surface with travel times of 1–3 days (Lawrence *et al.* 1996; Zaidman *et al.* 1999; Allshorn *et al.* 2007).

Spatial variability of karstic groundwater flow

In England, chalk karst is often associated with the Chalk–Paleogene boundary (Banks *et al.* 1995; MacDonald *et al.* 1998; Maurice *et al.* 2006). At a catchment in southern England, Maurice *et al.* (2006) divided the Chalk into three zones based on the distance from this boundary. Karst Zone 1 is close to the boundary and has frequent stream sinks and dolines and dissolution pipes. Karst Zone 2 is an intermediate area, where Clay-with-Flints superficial deposits (Paleogene remnants) occur and there are dolines and dissolution pipes, but no stream sinks. Karst Zone 3 is furthest from the boundary where surface karst appears rare. Whilst there are clear surface geomorphological differences, sub-surface karst and rapid groundwater flow is harder to assess. Maurice *et al.* (2006) presented some evidence for rapid groundwater flow in areas of England

Table 3. *Tracer tests in the Chalk in England*

| | Number of proven connections | Number of injection sites | Distances (km) | Mean distance (km) | Range of groundwater velocities (km d ⁻¹) | Mean (Median) groundwater velocity (km d ⁻¹) | Range of tracer recoveries* (%) | References |
|--|------------------------------|---------------------------|----------------|--------------------|---|--|--|---|
| Stream sink to spring | 27 | 13 | 1.2 to 18.9 | 9.9 | 0.6 to 12.3 | 4.7 (4.6) | <0.000007 to 69.1 (for 15 connections) | Banks <i>et al.</i> (1995), Maurice <i>et al.</i> (2006), Maurice <i>et al.</i> (2010), Harold (1937), Cook (2010), Brauns <i>et al.</i> (2017), Price (1979), Barton <i>et al.</i> (undated), Atkinson and Smith (1974) |
| Stream sink to abstraction borehole | 12 | 5 | 0.9 to 19.5 | 14.3 | 0.6 to 5.8 | 3.0 (3.0) | 0.1 to 3.8 (for 3 connections) | Harold (1937), Cook (2010), Price (1979), Richards and Brincker (1908) |
| Monitoring borehole to monitoring borehole | 5 | 5 | 0.03 to 3.1 | 1.3 | 0.05 to 0.44 | 0.23 (0.15) | Not applicable | Ward <i>et al.</i> (1997), Watson (2005) |
| Monitoring borehole to spring | 8 | 6 | 0.4 to 15.9 | 4.3 | 0.1 to 1.3 | 0.49 (0.39) | 0.05 to 1.2 (for 2 connections) | Ward <i>et al.</i> (1997), Ward <i>et al.</i> (1998), Hull (1995), Sims (1988), Cook (2010) |
| Monitoring borehole to abstraction | 36 | 29 | 0.002 to 5.7 | 0.9 | 0.01 to 19.4 | 1.6 (0.4) | 0.15 to ~100 (for 22 connections) | Cook (2010), Ward <i>et al.</i> (1998), Kachi (1987), Kirk and Chadha (1989), Skilton and Wheeler (1988), Hartmann <i>et al.</i> (2007), Atkinson and Smart (1981), Howard (1982), Joseph and Brown (1976), Maurice <i>et al.</i> (2016b), Colisch (1976), Bottrell <i>et al.</i> (2010), Matthias <i>et al.</i> (2007) |
| Soakaway to monitoring borehole | 1 | 1 | 3 | – | 2.6 | – | – | Price <i>et al.</i> (1992) |
| Soakaway to abstraction | 2 | 2 | 3.0, 3.2 | – | 0.2, 1.1 | – | 0.000002 (for 1 connection) | Price <i>et al.</i> (1992); Robertson <i>et al.</i> (1966) |
| Lake to monitoring borehole | 1 | 1 | 0.15 | – | 0.4 | – | – | Atkinson and Low (2000) |
| Lake to abstraction | 3 | 1 | 0.06 to 0.22 | 0.15 | 0.2 to 2.6 | 1.4 (1.2) | – | Atkinson and Low (2000) |
| Lake to spring | 2 | 1 | 0.22, 0.25 | – | 0.9, 1.6 | – | – | Atkinson and Low (2000) |

*Note that some tests recovered tracer at more than one outlet so these are recoveries for each connection, not for each tracer test.

Chalk karst hydrogeology

Table 4. Results from 1893 dye tracer tests performed in the Haute Normandy French Chalk between 1960 and 2007 reported by *Gombert et al. (2010)*

| Test type | Success rate (%) | Mean velocity (km d ⁻¹) |
|--|------------------|-------------------------------------|
| Stream sink to spring | 62 | 3.5 |
| Stream sink to borehole/ borehole to spring | 40–55 | 2.4 to 3.4 |
| Surface to borehole | 22 | 0.29 |

in the equivalent of Karst Zones 2 and 3 from a small number of tracer tests. The England-wide karst data presented here enable new consideration of the extent of rapid groundwater flow in areas away from the Chalk–Paleogene boundary.

In *Figure 5* the evidence for karst is clipped to only show sites >5 km from the Chalk–Paleogene boundary. These datasets are incomplete and yet show a remarkably extensive body of evidence to suggest that connected networks of conduits and fissures are common and occur throughout the Chalk. Forty-nine connections with rapid groundwater flow have been demonstrated by tracer tests conducted in areas away from the Chalk–Paleogene margin (injection sites shown on *Fig. 5*). These proved rapid groundwater flow (0.01 to 6.7 km d⁻¹, mean 1.7 km d⁻¹) over distances of 0.002 to 15.9 km (mean 0.9 km). Large springs, losing rivers, and high transmissivity are frequent away from the Paleogene margin; and caves occur, including Beachy Head Cave, the longest known cave in the English Chalk. Whilst stream sinks and dolines appear rare in the Chalk of northern England where adjacent Paleogene strata are absent, large springs,

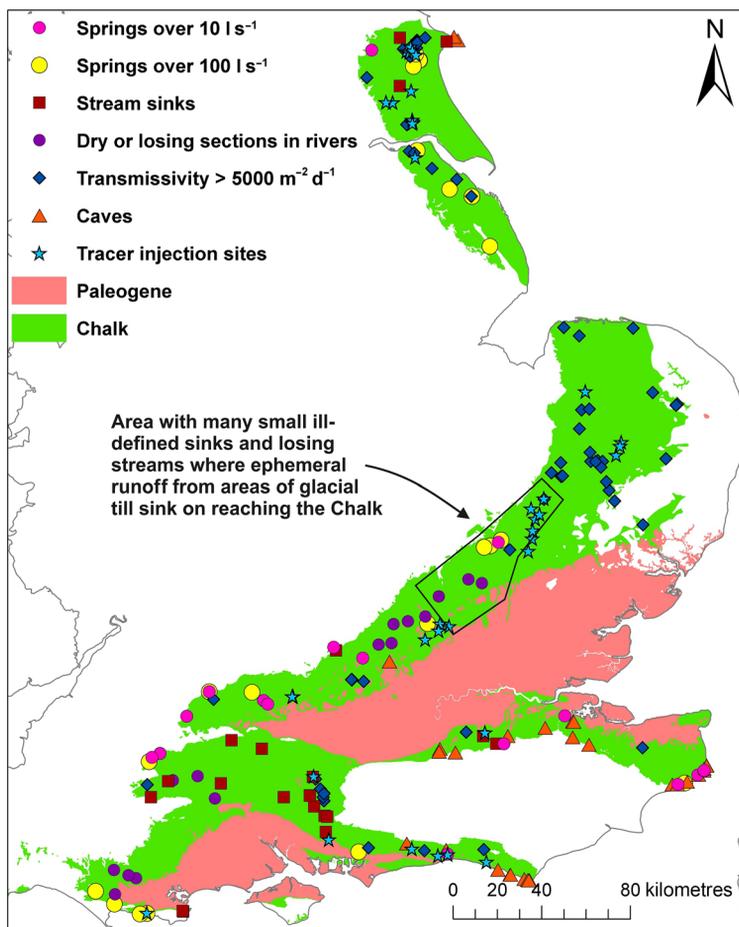


Fig. 5. Evidence for karst >5 km from the Paleogene boundary.

high transmissivity, and rapid flow indicated from tracer tests indicate subsurface karstic development still occurs.

In addition to the evidence shown on Figure 5, many single borehole dilution tests have demonstrated rapid dilution in areas away from the Paleogene in northern (Parker 2009; Agbotui *et al.* 2020), eastern (Kachi 1987) and southern (Maurice *et al.* 2012) England; implying well-connected networks of conduits and fissures. The locations of conduits have not been collated, but they are common in areas away from the Paleogene in boreholes, quarries and cliffs (e.g. Farrant *et al.* 2017, 2021a). In Berkshire conduits up to 10 cm in diameter were observed in three boreholes away from the Paleogene, and sediment was observed in a void at 74 m depth implying a connected flowpath to the surface (Maurice *et al.* 2012). Surface depressions also occur away from the Paleogene, and some of these will be karstic dolines, although many are likely to be anthropogenic.

Overall, there is evidence for karst and rapid groundwater throughout the Chalk in England. Foley and Worthington (2021) also conclude that rapid flow is widespread within the Chalk. This is possible because mixing dissolution can occur anywhere including at depth, and in confined settings (Worthington and Foley 2021; Farrant *et al.* 2021b). Moreover, landscape evolution has changed the position of the Paleogene boundary, and thus the location of stream sinks and rapid flow pathways over the past c. 1 Ma has also changed. Subsurface conduits formed when the geological boundary was in a different place may still be active in areas away from the current boundary (Maurice *et al.* 2006).

Outside England, the Normandy area of northern France has particularly strong evidence of chalk karst (Rodet 2007). Thousands of tracer tests have demonstrated rapid flow (Gombert *et al.* 2010), and accessible caves are common with over 6 km of passages surveyed in the Lower Seine valley (Rodet 1985; Ballesteros *et al.* 2020; Nehme *et al.* 2020; Farrant *et al.* 2021b). The more extensive cave development than seen in England may be due to differences in lithology, the extent of superficial cover, and geomorphological setting (Nehme *et al.* 2020; Farrant *et al.* 2021b). The Chalk in Northern Ireland forms a condensed sequence of indurated chalks and hardgrounds where karst is locally significant. Tracer tests from nine stream sinks proved velocities of 0.5 to 2.8 km d⁻¹ over 0.3 to 3.6 km, and there is a 500 m long cave (Barnes 1999).

Identifying the locations of karstic flowpaths is difficult, but as in all karst aquifers, conduit development in the Chalk is focused on three areas: stream sinks formed by allogenic runoff, convergence zones focused on springs, and deeper conduits and fissures formed by mixing dissolution (Ford and

Williams 2007; Farrant *et al.* 2021b). In karst there are often particular inception horizons where dissolution is focused. This is an important influence on karstic flows in the Chalk with strong evidence that fissures, conduits, caves and springs are often located on particular stratigraphical horizons associated with marls, hardgrounds and flints (Allen *et al.* 1997; Schurch and Buckley 2002; Gallagher *et al.* 2012; Maurice *et al.* 2012; Farrant *et al.* 2016; Ballesteros *et al.* 2020). Stratigraphical studies may enable improved understanding of local spatial variability in the amount and location of rapid groundwater flow.

Chalk karst hydrogeology

Karst aquifers have been classified based on different types of void network and the proportion of groundwater flux through each (Atkinson 1985, 1986). Conduits and fissure networks form the first two types with channelized flow, while diffuse flow occurs through a combination of primary porosity and unenlarged fractures. This classification was elaborated by Hobbs and Smart (1986) to include consideration of recharge and storage as well as flow in the saturated zone, in a conceptual framework.

The Chalk is an unusual karst aquifer because cave development is limited, which may be a result of several factors reducing the rate of conduit enlargement (Farrant *et al.* 2021b). High matrix porosity enables groundwater to tend towards saturation quickly; high fracture density disperses dissolutional development along many fractures; and landscape evolution and the erosion of Paleogene sediments may have limited cave development if stream sinks were not functional for long enough for larger caves to form. However, the evidence presented here suggests that extensive connected networks of smaller conduits and fissures are common in the Chalk. So how do they function and what are the implications for Chalk hydrogeology? The Chalk is a complex aquifer with four porosity components: the bedrock matrix, the unmodified fracture network, solutionally enlarged fractures (fissures) and conduits/caves. Figure 6 shows typical sizes and apertures of these different voids. Understanding the amount of recharge, storage and flow that occurs in these different components and how fast the groundwater moves through them is central to understanding how the aquifer functions. This is considered below, but there remain many uncertainties (Table 5), which might form the basis of future investigations.

Recharge and unsaturated zone flow

Chalk recharge has traditionally been divided into slow 'matrix' flow and rapid 'bypass' flow. Bypass

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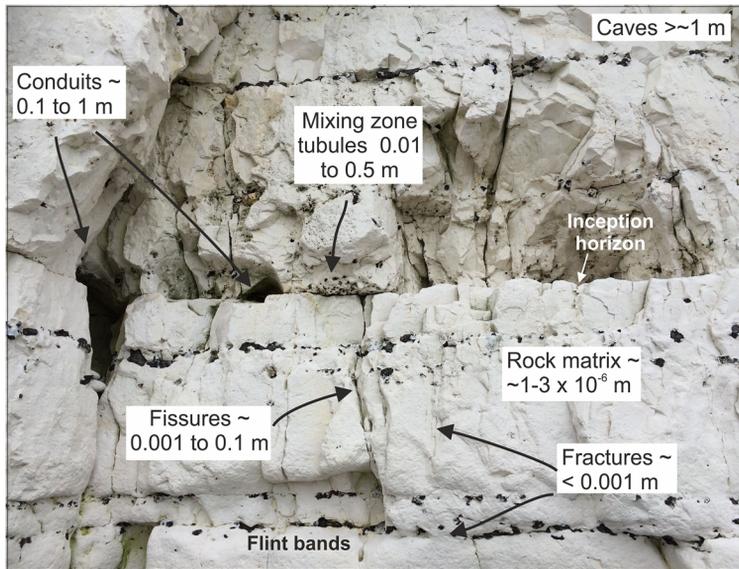


Fig. 6. Components of the Chalk aquifer.

flow has been inferred from the responses of monitoring boreholes to rainfall (e.g. Lee *et al.* 2006); and is greatly influenced by the duration and intensity of rainfall events (Ireson and Butler 2011). In this section the rates and proportions of recharge through the four chalk porosity components are considered using information from the limited number of tracer tests that are available, together with insights from the literature, with a summary in Table 5 highlighting the uncertainties.

Small-scale tracer tests and pore water profiles have demonstrated downward flow rates of about 1 m a^{-1} through the bedrock matrix in the unsaturated zone (Foster and Smith-Carrington 1980; Wellings 1984; Barraclough *et al.* 1994; Brouyère *et al.*

2004; Van den Daele *et al.* 2007). This flow rate has been consistently observed in different geographical locations and in different types of study, suggesting that it is likely to be a reliable estimate and that this downward flow rate is likely to occur throughout the Chalk. Unsaturated zone flow through the unmodified fracture network is likely to be faster than through the bedrock matrix (Price 1987).

Recharge through fissures is likely to be rapid, and a study in the Yorkshire Chalk by Allshorn *et al.* (2007), has demonstrated flows through the unsaturated zone of 9.5 to 19 m d^{-1} over vertical distances of 15 to 38 m . Tracer applied to shallow pits dug through the soil to bedrock was detected in

Table 5. Recharge, storage and flow in the different components of the Chalk aquifer with indicative levels of uncertainty: high (*italic text*), moderate (*plain text*) and lower (**bold text**)

| | Bedrock Matrix | Fractures | Fissures | Conduits/caves |
|---|--|---|---|---|
| Recharge rate/ unsaturated zone flow velocity | Slow ($\sim 1 \text{ m a}^{-1}$) | <i>Episodic flow during recharge, velocity uncertain</i> | Episodic rapid flow (tens m d^{-1}) | Very rapid (thousands m d^{-1} via stream sinks) |
| Recharge proportion (varies spatially) | | High ($\sim 70\text{--}90\%$)? | Low? ($5\text{--}20\%$)? | Low? ($0\text{--}15\%$)? |
| Contribution to drainable storage | Low | | High | <i>Low (to moderate?)</i> |
| Flow rate (saturated) | Very slow ($< \sim 0.1 \text{ m a}^{-1}$) | <i>Slow to moderate? Velocity uncertain ($\sim 1 \text{ m d}^{-1}$)?</i> | Rapid (tens to thousands m d^{-1}) | Very rapid (thousands m d^{-1}) |
| Proportion of flow (saturated) | Very low? | <i>Moderate to High?</i> | <i>High?</i> | <i>Low to moderate?</i> |

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fissure inflows to anthropogenic tunnels within the unsaturated zone, providing direct evidence of velocities in unsaturated zone fissures.

Recharge into conduits and caves is indicated by the presence of stream sinks. Many tracer tests from these have demonstrated flow velocities of thousands of metres per day (through both the unsaturated and saturated zones) indicating rapid recharge and suggesting very rapid unsaturated zone flow.

There is some uncertainty about the proportions of recharge through the different porosity components (Table 5). Smith *et al.* (1970) investigated tritium in Chalk groundwater concluding that about 15% of recharge occurs via rapid bypass flow in the unsaturated zone. The karst data presented here suggest that the proportion of rapid recharge may vary spatially and temporally within the aquifer. Rapid recharge through conduits and caves is likely where stream sinks are present (Fig. 2a). Where Chalk rivers have substantial losing sections, these may also contribute a high proportion of 'bypass' recharge into networks of solutional fissures and conduits; and rapid recharge may also occur through fast infiltration soakaways and sustainable urban drainage systems (SUDs). However, the actual proportion of recharge which is via these rapid routes is not known.

The role of dolines and dissolution pipes in recharge is less obvious than stream sinks. If they are infilled or lined with impermeable clays then recharge may be limited, and they are unlikely to provide rapid flow pathways to the subsurface. Some dolines contain pools suggesting slow recharge. However, many do not contain water and their formation processes suggest that there may be connected networks of solutional voids beneath them that could enable bypass recharge. Some dolines may be relict stream sinks, and therefore have well-developed cavities beneath them.

There is evidence from tracer tests for rapid recharge and unsaturated zone flow via solutional fissures in areas away from obvious surface karst features. In the Yorkshire Wolds, tracer injections at arbitrary points on the surface proved rapid unsaturated zone flow (Zaidman *et al.* 1999; and the study by Allshorn *et al.* (2007) discussed above). Zaidman *et al.* (1999) applied saline tracer to the surface with artificial recharge in an interfluvial area, and geophysics was used to monitor subsurface tracer migration. Tracer was detected 15 m below the surface after 1 day, indicating rapid bypass flow, induced by the artificial recharge. Substantial tracer was retained close to the surface and over 10 months this was remobilized by natural intense precipitation events resulting in detection again at 15 m depth. A second experiment demonstrated rapid bypass flow, with responses observed at 25 m 1–2 days after rainfall. A tracer test has also been conducted

in Berkshire where the top soil was removed and tracers were applied to the surface with substantial irrigation (Lawrence *et al.* 1996). Lithium bromide, bacteriophage and 1 μm microspheres were detected at the water table within 3 days, indicating rapid unsaturated zone flow to 25 m below the surface. Larger-sized tracer particles (up to 10 μm) did not reach the water table but travelled at least 10 m down through the unsaturated zone (indicated by core samples). Substantial tracer was retained in the top 10 m, and tracer recoveries were extremely low indicating very high attenuation (Lawrence *et al.* 1996).

A key question is how frequently vertical fissures enabling rapid bypass flow through the unsaturated zone occur without surface expression. The tracer studies suggest that they may be relatively frequent, at least in some areas, as in all cases tracer was detected even though it was not injected into a karst feature. If vertical fissures were infrequent, then these tracer tests from arbitrary points would not have produced positive results. There is further evidence from coastal and quarry sections where vertical fissures can penetrate over 90 m below the surface (Fig. 7; Farrant *et al.* 2021a). The vertical connectivity of the fracture network is influenced by lithology. Conjugate fractures are common in parts of the succession with marl seams; such as in the New Pit Chalk, Lewes Nodular Chalk, lower Seaford Chalk and Newhaven Chalk formations. These are relatively limited, extending down to the next marl seam, where they may intersect bedding-controlled conduits and fissures. Where marl seams are absent, such as in the middle and upper Seaford Chalk Formation, vertical fissures penetrate much greater thickness of Chalk, sometimes over 100 m.

Overall, rapid bypass flow is more likely in areas with focused point recharge via stream sinks, and via losing rivers and high infiltration rate soakaways and SUDs on outcrop Chalk. However, there may also be many vertical fissures with no surface expression enabling rapid unsaturated zone flow. Estimating the proportions of recharge through the matrix, fractures, fissures and conduits; and how this varies spatially and temporally, remains an important area of research.

Storage

The Chalk has high storage, with a median storage coefficient of 0.0023 from 1200 pumping tests (MacDonald and Allen 2001). Whilst the matrix has high porosity and therefore holds large amounts of water, the small size of the pore throats limit drainage and therefore the relative contribution to the storage coefficient is small. Allen *et al.* (1997) suggest that most of the specific yield comes from fractures and fissures, implying these may hold the greatest

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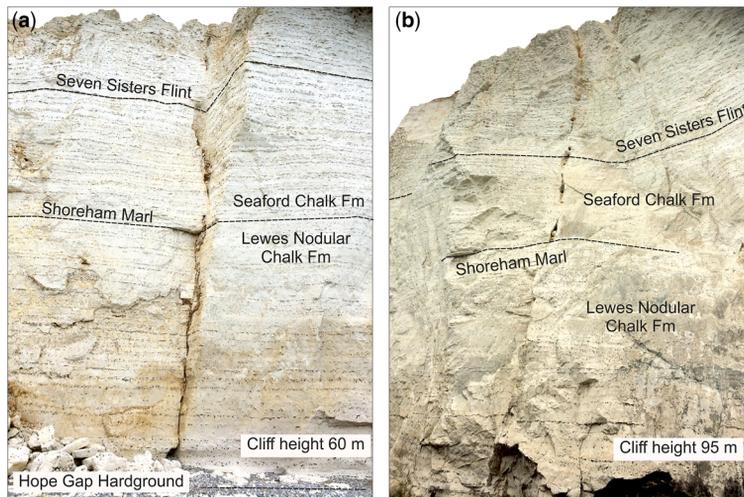


Fig. 7. Solutional vertical fissures up to 0.5 m wide in southern England: (a) Seaford Head; (b) Beachy Head.

proportion of drainable storage (Table 5). Storage in conduits and caves remains unknown. In karst aquifers caves are generally considered to provide only a small contribution to storage (Ford and Williams 2007), but the nature of the Chalk with many smaller conduits and dissolution tubules rather than fewer larger caves may result in greater storage in conduits.

Saturated zone flow

Saturated zone flow rates in the four porosity components are summarized in Table 5. Tracer tests prove rapid flows of tens to thousands of metres per day in conduits and fissures in the saturated zone (Figure 4, Table 3), while flow rates in the matrix and fractures are more uncertain but likely to be low (Price 1987). Saturated zone matrix flow is likely to be slower than in the unsaturated zone due to the substantially lower hydraulic gradients. The proportions of flow in the different void types are uncertain (Table 5), and the interactions between the matrix, fractures, fissures and conduits in the saturated zone are complex.

Key questions are how frequently does rapid groundwater flow occur; and what proportion of water supplying springs and abstraction boreholes is rapid? The tracer data provide strong evidence that rapid saturated zone flow is common (Table 3; Fig. 4). However, there is also evidence that much of the water in the Chalk aquifer travels slowly. High storage values, and slow seasonal borehole water level responses (Allen *et al.* 1997), together with the high base flows observed in Chalk rivers, suggest a high proportion of slower moving water in the aquifer. This is supported by CFC and SF₆ residence times from 21 pumping stations ranging from

a few years to a few decades, although some sites were contaminated (Darling *et al.* 2005).

An explanation reconciling this apparently contradictory evidence is that springs and abstractions are supplied by complex well-connected fissure and conduit networks, sustaining the high discharges and yields, and producing the rapid flows observed in tracer tests; but these networks are in turn supplied by the smaller fractures and fissures, and perhaps the bedrock matrix. Further research is needed on rapid flow in the saturated zone.

Implications for groundwater protection

Vulnerability to pollution

Like all karst aquifers, the Chalk is vulnerable to pollution because connected networks of fissures and conduits enable rapid flow over many kilometres. Limited cave development and substantial storage in the bedrock matrix and unmodified fracture network in both the unsaturated and saturated zones provides more protection through dispersive and diffusive attenuation mechanisms than in classically karstic aquifers. However, one consequence is that pollution events may persist for longer.

Vulnerability occurs where surface karst features (stream sinks, dolines, losing rivers, and vertical fissures) provide pathways enabling rapid contaminant transport through the unsaturated zone. Saturated zone karst creates vulnerability to subsurface activities and enables long distance transport of point source and diffuse pollutants. A bromate plume in Hertfordshire (Cook *et al.* 2012) illustrates this vulnerability and provides insight into the interactions

between the different aquifer components. Bromate from a long-term surface spill has been stored in the porous bedrock matrix and fracture network for many decades. Dispersion and diffusion enable the bromate to move into the subsurface karstic fissure/conduit system which transports bromate to springs and boreholes more than 20 km away (Fig. 8). It is an important example demonstrating how saturated zone karst can result in the transport of pollutants over very long distances.

Catchment delineation

In karst aquifers catchment delineation is difficult. Water balance can help determine the likely catchment size, but identifying the contributing surface areas is complicated by groundwater flow between topographical catchments, divergent flow to multiple outlets, and the potential for different flow paths going in different directions at different depths in the aquifer (Malard *et al.* 2015). Further complexity arises from often substantial changes in catchments under high and low water levels.

It is especially difficult to define catchments in karst aquifers like the Chalk where caves are rare and large springs and abstractions are fed by many different conduit and fissure networks which may extend several kilometres. Tracer testing may identify some flow paths, and greatly improve conceptual understanding of Chalk catchments, but because of the complexity and number of flowpaths, the locations of many may remain unknown.

The Hertfordshire example illustrates this complexity and the challenges in catchment delineation in the Chalk. Tracer tests have demonstrated 23 different connections between three different sets of

swallow holes in the Water End area with divergent flow to 11 different outlets 7 to 19 km to the NE and east (Harold 1937; Cook 2010). Bromate in the outlets demonstrates that they are also connected to Sandridge, much further to the west, in a different topographical catchment away from the stream sinks (Fig. 8). Without this ‘accidental’ tracer test this part of the catchment of these outlets would probably not have been identified. It is likely that other Chalk springs and abstractions in England may obtain water from outside modelled catchment areas.

Catchment delineation in karst requires a combination of cave mapping, tracer testing, water balance, and borehole monitoring networks; to enable a good conceptual understanding of the likely locations of the conduit and fissure networks (Malard *et al.* 2015). In the Chalk where this may be particularly challenging, some measure of the uncertainty in the catchment location may be useful for management.

Nitrate pollution

High nitrate is a major issue in the Chalk and is difficult to resolve due to multiple pollution sources (El Gaouzi *et al.* 2013) and its long persistence (Wang *et al.* 2012). Nitrate at a supply may be derived from different sources with variable residence times making it difficult to understand the role of karst in nitrate transport. Nitrate generally moves slowly through the unsaturated zone over decades (Wang *et al.* 2012). However, in some places, karst may enable rapid transport of nitrate through the unsaturated zone via stream sinks, losing sections of rivers, soakaways, dolines, or vertical solution pathways with no surface expression. This has not been studied but data presented here show that

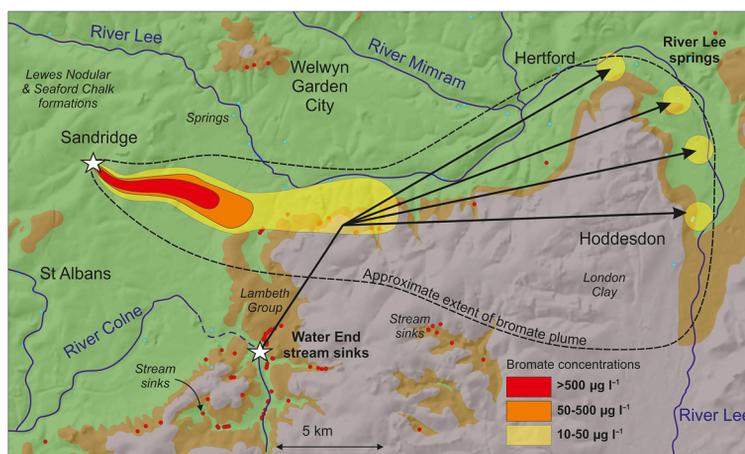


Fig. 8. Evidence of interactions between aquifer components from a bromate pollution plume, modified from figures and data in Cook (2010). Many individual tracer connections omitted for clarity.

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these features may be common and warrant future research. Nitrate concentrations in stream sinks are not generally measured but are likely to vary spatially and temporally, with some contributing nitrate and others providing dilution, depending on the land use in the stream sink catchment, and the time of year. Rapid bypass flow via vertical solution features with no surface expression may also impact on nitrate if point pollution sources (e.g. septic tanks, sewers, slurry pits and animal waste piles) happen to be located on or near them.

A major and perhaps overlooked impact of karst on nitrate pollution in the Chalk is the role of saturated zone karst in transporting nitrate over long distances. Karst and tracer data suggest that springs and abstractions are likely to be supplied by extensive networks of fissures and conduits extending over long distances. However, as they have generally not been considered karstic, catchment areas may extend outside modelled areas; and some nitrate may be from areas not currently considered in catchment management and mitigation.

Groundwater protection methods

Groundwater Protection in the UK is based on Source Protection Zones (SPZs). SPZ 1 is the area around an abstraction or spring with a 50-day groundwater travel time through the saturated zone; SPZ 2 is the 400-day saturated zone travel time area, and SPZ 3 is the total catchment (Environment Agency 2019). There is high degree of protection and restrictions on potentially polluting activities in SPZ 1 and some protection in SPZs 2 and 3 (Environment Agency 2018). In Europe SPZs in karst are usually defined using the EPIK catchment vulnerability method which considers the epikarst, protective over, infiltration rates and karst networks (Doerflinger *et al.* 1999). In Switzerland EPIK is used in most karst aquifers and the DISCO method (Pochon *et al.* 2008) in fractured or weakly karstic aquifers. DISCO involves classifying spring sources in three categories based on the vulnerability of the spring, with different methods of SPZ delineation based on this classification.

In the UK, most SPZs are delineated using MODFLOW groundwater modelling, more recently with the Flowsource adaptation of Black and Foley (2013) for defining capture zones (Environment Agency 2019). Flowsource enables varying proportions of water from individual model cells to be captured by the abstraction. Incorporating karst conduit systems in Chalk groundwater models is challenging as their locations are poorly known; and they appear to consist of many converging and diverging flowpaths creating complex networks. However, recharge via stream sinks could be fairly easily incorporated into groundwater models and,

given the strong geological controls on rapid groundwater flow, new methods of integrating geological and karst data and understanding into groundwater models could enable improved catchment delineation. Recent work (Worthington *et al.* 2019; Foley and Worthington 2021; Medici and West 2021) suggests that much lower effective porosities (c. 0.001 to 0.0001) than are currently used in groundwater models may be required to represent the rapid flow that occurs in the Chalk, to provide a more realistic estimate of the distance from which groundwater reaches abstractions in less than 50 days.

In UK karst aquifers a manual method is used for SPZ delineation based on water balance, karst and tracer test data, coupled with conceptual understanding (Environment Agency 2019). SPZs in the highly karstic Carboniferous limestones have been defined in this way, and more recently SPZs at a karstic site in Jurassic limestones (Environment Agency 2019). These have resulted in very large SPZs. Protecting such large areas may be challenging but these SPZs provide a better representation of the areas from which rapid groundwater flow may impact the abstraction than might be obtained by conventional modelling.

Karst methods have not been applied to Chalk SPZ delineation, primarily due to limited understanding of Chalk karst. There therefore remains a discrepancy between modelled SPZs with 50-day saturated zone travel times that cover very small areas, and the evidence from tracer testing (Table 3) which demonstrates groundwater flow over kilometres in just hours or days. A specific example is shown in Figure 9. Some groundwater takes just 14 hours to travel 850 m through the whole of SPZ1 to abstraction 3 from a monitoring borehole. There is no reason that this flowpath would suddenly start at the monitoring borehole, and this rapid flowpath must extend further up-gradient into the aquifer. Many tracer tests have demonstrated rapid flow to abstractions, and it seems likely that all successful abstractions are fed by connected networks of conduits and fissures, but a question that remains is how far these flowpaths extend. Given the new evidence for karst and rapid flow in the Chalk presented in this paper, and other recent work (e.g. Foley and Worthington 2021), it is likely that karst-specific methods of SPZ delineation in the Chalk would be appropriate at many abstractions.

As further karst datasets and understanding are developed, Chalk springs and abstractions could be classified with different levels of risk of karst and rapid groundwater flow, to assist with catchment management and groundwater protection strategies. Scores could be calculated based on factors indicating rapid groundwater flow such as: coliforms or turbidity at the supply; high transmissivity or pumping rate; conduits visible on abstraction

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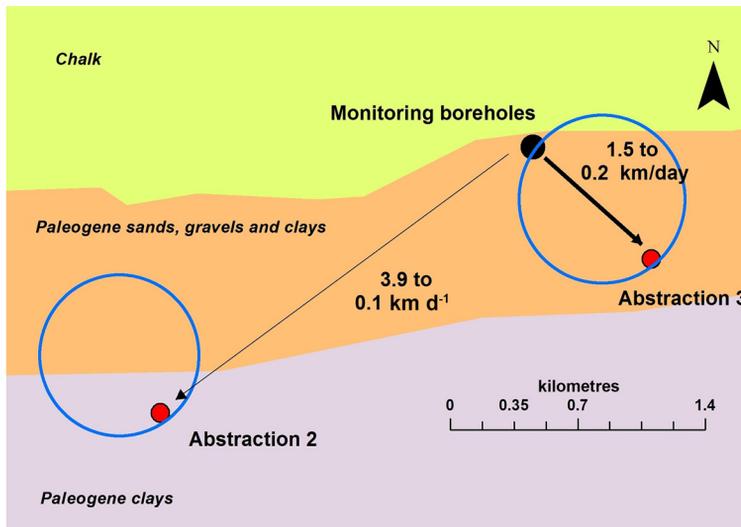


Fig. 9. Example of the discrepancy between modelled 50-day SPZ1 (blue circles represent approximate SPZ1 area) and rapid groundwater flow demonstrated by borehole tracer tests (thick arrow denotes connection; thin arrow denotes very weak connection).

borehole images; evidence of connectivity with surface water rivers or the sea; tracer tests proving rapid flow to the supply or within the catchment; and stream or river sinks, springs and caves in the catchment.

Conclusions

This study presents evidence that karst in the Chalk is more common than previously thought. Many tracer tests prove rapid groundwater flow. There are high densities of stream sinks; and karstic conduits, springs, dolines, and dissolution pipes are common. However, the Chalk is very different from classical karst aquifers because although some small caves occur, subsurface karst mainly comprises smaller conduits (*c.* 5–20 cm) and fissures. High storage and pollutant transfer between the four porosity components via dispersion and diffusion enable much higher attenuation in both the unsaturated zone and saturated zone than in classical karst aquifers. However, considering the Chalk as karstic is useful as many of the principles that apply to more classical karst aquifers also apply to the Chalk. The unique nature of Chalk karst may also help improve understanding of classical karst aquifers where major caves are the main focus of research, and fissures/smaller conduits are not so well understood.

Geology is central to the development of Chalk karst with stream sinks concentrated on the Chalk–Paleogene boundary, dolines and dissolution pipes

associated with shallow superficial deposits, and subsurface cave, conduit and fissure development concentrated on particular lithological inception horizons. The evidence for karst and rapid flow is strongest near to the Chalk–Paleogene margin where stream sinks are present. However, the karst data presented here show extensive evidence for subsurface solutional development and rapid flow in areas away from this geological boundary. In these areas it is likely that rapid recharge through the unsaturated zone is less common, but it can still occur. This may be obvious where there are losing rivers, or soakaways or SUDs with high infiltration rates; but difficult to detect where it is via vertical solutional fissures with no surface expression. There is considerable evidence that saturated zone karst (connected networks of fissures and conduits) is common in areas away from the Paleogene margin.

Groundwater management and protection is challenging because the locations of all the karstic conduit/fissure systems will never be known, although tracer tests may greatly improve conceptual understanding of catchments. At Chalk abstractions and springs, there can be rapid flow from anywhere within the catchment, although there will not be rapid flow from everywhere, and the rapid flow component is diluted by water from longer-term storage. It is relatively easy to identify springs/abstractions which are impacted by rapid groundwater flow from the surface where water quality is impacted; and protection strategies focused on stream sinks, losing rivers, soakaways, SUDs, and dolines are

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likely to be effective. Springs and boreholes that are not connected to the surface via rapid flow are less vulnerable to surface activities, but rapid groundwater flow over long distances is still likely in the saturated zone. All springs with high discharge, or high yielding abstractions are therefore vulnerable to subsurface activities, and due to the difficulties in catchment delineation there may be pollutant sources from areas far outside modelled catchment areas.

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