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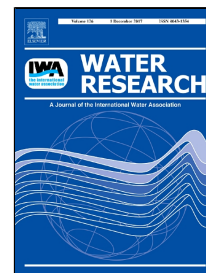
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Highlights:

- Conceptual review of hyporheic zones of rivers as natural anoxic-oxic bioreactors

- Interactions between organohalide respiration and biogeochemical cycling

- Aerobic vinyl chloride mineralisation during hyporheic mixing is conceptualised

- Field experience, challenges and characterisation technologies critically reviewed

1   **Natural attenuation of chlorinated ethenes in hyporheic zones: a review of**

2   **key biogeochemical processes and in-situ transformation potential**

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17   **Abstract**

18   Chlorinated ethenes (CEs) are legacy contaminants whose chemical footprint is expected to

19   persist in aquifers around the world for many decades to come. These organohalides have been reported

20   in river systems with concerning prevalence and are thought to be significant chemical stressors in urban

21   water ecosystems. The aquifer-river interface (known as the hyporheic zone) is a critical pathway for

22   CE discharge to surface water bodies in groundwater baseflow. This pore water system may represent a

23   natural bioreactor where anoxic and oxic biotransformation process act in synergy to reduce or even

24   eliminate contaminant fluxes to surface water. Here, we critically review current process understanding

25   of anaerobic CE respiration in the competitive framework of hyporheic zone biogeochemical cycling

26   fuelled by in-situ fermentation of natural organic matter. We conceptualise anoxic-oxic interface

27   development for metabolic and co-metabolic mineralisation by a range of aerobic bacteria with a focus

28   on vinyl chloride degradation pathways. The superimposition of microbial metabolic processes

29   occurring in sediment biofilms and bulk solute transport delivering reactants produces a scale

30   dependence in contaminant transformation rates. Process interpretation is often confounded by the

31   natural geological heterogeneity typical of most riverbed environments. We discuss insights from recent

32   field experience of CE plumes discharging to surface water and present a range of practical monitoring

33   technologies which address this inherent complexity at different spatial scales. Future research must

34   address key dynamics which link supply of limiting reactants, residence times and microbial

35   ecophysiology to better understand the natural attenuation capacity of hyporheic systems.

36   **Keywords:** Chlorinated ethenes, hyporheic zone, natural attenuation, biogeochemistry,

37   biotransformation, heterogeneity.

38   **Abbreviations:**

39   1,1,1-TCA: 1,1,1-trichloroethane; DCA: 1,2-dichloroethane; Ac: acetate, CA: chlorinated ethane;

40   cDCE: *cis*-1,2-dichloroethene; CE: chlorinated ethene; Da: Damköhler number; DIC: dissolved

41   inorganic carbon; DNAPL: dense, non-aqueous phase liquid; DOC: dissolved organic carbon; DOM:

42   dissolved organic matter; DON: dissolved organic nitrogen; HCB: hexachlorobenzene; HCE: higher

43   chlorinated ethene; HFC: hyporheic flow cell; K: saturated hydraulic conductivity; LCE: lower

44   chlorinated ethene; OHR: organohalide respiration; OHRB: organohalide-respiring bacteria; OM:

45   organic matter; PCE: tetrachloroethene; PCR: polymerase chain reaction; POM: particulate organic

46   matter; SCFA: short chain fatty acid; SOM: sedimentary organic matter; TCE: trichloroethene; TCM:

47   trichloromethane; TEA: terminal electron acceptor; TEAP: terminal electron accepting process; VC:

48   vinyl chloride.

49   **1   Introduction**

50   Legacy soil and groundwater contamination by chlorinated ethenes (CEs) remains a global

51   environmental concern (Burston et al., 1993; Jackson, 1998; Rowe et al., 2005; Rivett et al., 2006, 2012;

52   2014). The higher CEs (HCEs), tetrachloroethene (PCE) and trichloroethene (TCE) are parent

53   compounds employed historically as non-polar solvents for industrial cleaning and degreasing

54   applications which enter aquifers as dense, non-aqueous phase liquids (DNAPLs) (Pankow and Cherry,

55   1996; Rivett et al., 2014) and release dissolved-phase plumes (e.g. Benker et al., 1996; Rivett and

56   Feenstra, 2005; Koch and Nowak, 2015). Diffusive retention of HCEs in low-permeability and fractured

57   formations is increasingly recognised as an important driver of plume longevity from the resulting back-

58   diffusive flux (Parker et al., 1994; Seyedabbasi et al., 2012; Damgaard et al., 2013; Yang et al., 2014).

59   Under predominantly anoxic conditions, HCEs are reductively transformed to the lower CEs (LCEs)

60   *cis*-1,2 dichloroethene (cDCE) and vinyl chloride (VC) (Squillace and Moran, 2007; Chambon et al.,

61   2010; Huang et al., 2014; Lu et al., 2015). CEs have been detected in surface water with alarming

62   frequency around the world (e.g. Christof et al., 2002; McGuire et al., 2004; Ellis and Rivett, 2007;

63   Yamamoto et al., 2014 Wittlingerová et al., 2016). Groundwater baseflow input of CEs has been

64   implicated as a contributory factor in the global impoverishment of surface water ecological quality

65 known as urban stream syndrome (Meyer et al., 2005; Roy and Bickerton, 2011; Roy et al., 2017). The

66 aquifer-river interface, inclusive of the hyporheic zone (Fig. 1) forms a key diffuse pathway for

67 discharge of groundwater plumes hydrologically connected with local surface water bodies (Table 1)

68 (Conant et al., 2004; Chapman et al., 2007; McKnight et al., 2012; Weatherill et al., 2014; Puigserver et

69 al., 2014; Simsir et al., 2017; Sonne et al., 2017).

70   The hyporheic zone of streams and rivers has been defined by various disciplines from alternative

71 perspectives (e.g. Boulton et al., 1998; Buss et al., 2009; Krause et al., 2011; Cardenas, 2015). In this

72 review, we conceptualise the hyporheic zone as a dynamic volume of the saturated zone where the pore

73 water domain is physically influenced by exchanges across the sediment–water interface (Cardenas,

74 2009). This spatially and temporally variable volume may encompass shallow benthic sediments and

75 adjacent parts of the alluvial aquifer (Boano et al., 2014). The transition zone is frequently heterogeneous

76 with a wide range of saturated hydraulic conductivities (K) and associated scaling of solute residence

77 times (Haggerty et al., 2002; Sawyer and Cardenas, 2009; Gomez-Velez et al., 2014). Fluvial sediments

78 are sites of organic matter (OM) retention with high interstitial surface area linked to enhanced microbial

79 metabolic activity and characterised by sharp physical and biogeochemical gradients (Krause et al.,

80 2013, 2014; Boano et al., 2014; Atashgahi et al., 2015; Narango et al., 2015; Briggs et al., 2016). A

81 generic scenario is depicted in Fig. 1 where OM present in the sediment sequence induces anoxia along

82 discharging groundwater flow paths through the interface. The CE plume depicted may be composed

83 solely of parent HCEs (e.g. McKnight et al., 2010) or a mixture of LCEs (e.g. Sonne et al., 2017; Rønde

84 et al., 2017). Shallow oxic zones are created where local hyporheic exchange flows with the river

85 penetrate the underlying anoxic zone (Hester et al., 2013; 2017, Trauth et al., 2014). Heterotrophic

86 biogeochemical cycling of carbon, nitrogen, sulfur and metals will take place along anoxic sediment

87 flow paths (Baker et al., 2000; Mermillod-Blondin et al., 2005; Lautz and Fanelli, 2008; Rahimi et al.,

88 2015). Fluvial sediments can form niche environments for anaerobic biotransformation of CEs by

89 reductive dechlorination and much interest has arisen in their intrinsic capacity to assimilate

90 groundwater CE loads in baseflow (Table 1) (Conant et al., 2004; Abe et al., 2009; Hamonts et al., 2009;

91 Weatherill et al., 2014; Atashgahi et al., 2015; Freitas et al., 2015; Simsir et al., 2017). Discharging

92 groundwater flow paths through fluvial sediments may enter into mixing zones around infiltrating

93 surface water during hyporheic exchange flows (Boano et al., 2014; Gomez-Velez et al., 2014; Hester

94 et al., 2017). Dynamic pore water oxygen (O_2) gradients are induced in mixing zones as a result (Trauth

95 et al., 2014; Vieweg et al., 2015; Cardenas et al., 2016). These mixing zones provide a separate niche

96 environment for aerobic mineralisation of LCE daughter products originating in the anoxic zone or in

97 partially dechlorinated plumes (e.g. Rønde et al., 2017). Hence, hyporheic zones of groundwater-fed

98 rivers may constitute a unique natural 'biobarrier' for complementary anaerobic and aerobic biological

99 treatment of diffuse baseflow CE pollution (Hamonts et al., 2007; Abe et al., 2009; Atashgahi et al.,

100 2017a; Simsir et al., 2017).

101 The aim of this review is to provide a critical evaluation of the current understanding of

102 synergistic anoxic-oxic biotransformation processes and their environmental supporting conditions in

103 typical riverbed sediment sequences. We consider the fate of parent HCEs in anaerobic heterotrophic

104 food webs driven by OM fermentation and observed metabolic interactions with key biogeochemical

105 cycles. We conceptualise the development of anoxic-oxic interfaces in surface water mixing zones

106 which support aerobic bacteria capable of assimilating LCEs via metabolic and cometabolic pathways

107 with a particular focus on VC. We discuss challenges in integrating microbially-mediated transformation

108 processes occurring in sediment biofilms with larger scale hydrological fluxes observable at field scale.

109 We critically review available in-situ monitoring technologies and support our discussions with insights

110 from recent field experience. This process-based understanding is intended to underpin the risk-based

111 management of legacy CE contamination. The coupled biogeochemical and transport processes

112 understanding is highly relevant for river restoration and prospective ecological engineering applications

113 aimed at reducing the impact of organohalide stressors in urban streams (e.g. Lawrence et al., 2013;

114 Rasmussen et al., 2016; Roy et al., 2017).

2 Biotransformation of chlorinated ethenes in riverbed sediments

116 Biotransformation of CEs in hyporheic zones will occur when requisite bacteria, a circumneutral

117 pH, nutrients and electron donors/acceptors are present and where solute residence time equals or

118 exceeds the transformation timescale (half-life) of the contaminant (Meckenstock et al., 2015). In anoxic

119 zones, CEs may undergo anaerobic reductive dechlorination (hydrogenolysis) by heterotrophic bacteria

120 where chlorine atoms are sequentially replaced by hydrogen (Fig. 2) in organohalide respiration (OHR)

121 (Vogel et al., 1987; Hug et al., 2013; Leys et al., 2013). LCEs produced by OHR may reach oxic

122 interfaces where metabolic and co-metabolic aerobic pathways lead to complete natural attenuation.

123 This part of our review profiles key advances in the understanding of these synergistic biotransformation

124 process for which hyporheic zones present a unique capacity.

2.1 Organohalide respiration

126 Under anoxic conditions, CEs are terminal electron acceptors (TEAs) for energy conservation

127 (Fig. 2) during oxidation of an electron donor (usually hydrogen) in the presence of a carbon source

128 (usually acetate) and a nitrogen source (ammonium) (Mohn and Tiedje., 1992; Smidt and de Vos, 2004).

129 OHR is an energy-conserving process where the chemically stable carbon-halide bond of organohalides

130 is unlocked by replacing the halogen atom with hydrogen and liberating it as a halide (Fig. 2) (Maymo-

131 Gatell et al., 1997; McCarty, 1997; Aulenta et al., 2002). The reductive cleavage of carbon-halide bonds

132 is catalyzed by membrane-bound, cobalamin-containing enzymes known as reductive dehalogenases

133 (Bommer et al., 2014). OHR is mediated by organohalide-respiring bacteria (OHRB) belonging to

134 district bacterial groups i.e., *Chloroflexi*, *Proteobacteria* and *Firmicutes* (Atashgahi et al., 2016).

135 Members of the genus *Dehalococcoides mccartyi, Dehalogenimonas* and *Dehalobacter* spp. are

136 restricted to OHR for their metabolism, although fermentative growth is also shown in the latter (Justica-

137 Leon et al., 2014; Lee et al., 2012; Yang et al., 2017). In contrast, *Geobacter*, *Desulfuromonas*,

138 *Anaeromyxobacter*, *Desulfomonile*, *Desulfoluna*, *Desulfovibrio*, *Sulfurospirillum* and

139 *Desulfitobacterium* genera are facultative OHRB with versatile metabolisms including but not restricted

140 to OHR (Maphosa et al., 2010; Atashgahi et al., 2016).

2.2 Potential heterotrophic food webs in anoxic zones

2.2.1 Carbon flow in the hyporheic zone

143 Dissolved organic matter (DOM) is a key source of electron donors and nutrients for OHR and

144 other heterotrophic terminal electron accepting processes (TEAPs) in hyporheic zones (Pusch and

145 Schwoerbel, 1994; Fischer et al., 2005; Zarnetske et al., 2011; Stegen et al., 2016). Anaerobic DOM

146 mineralisation is coupled to the reduction of TEAs as follows (Donn and Barron, 2013):

147     $DOM + TEA \rightarrow DIC + metabolite + NH_4$

148   Where TEA (in descending potential metabolic energy yield) is $O_2 > NO_3 > PCE > Fe(III) > TCE >$

149     $cDCE > SO_4 > VC > CO_2$, DIC is dissolved inorganic carbon, metabolite is N_2, CO_2, Mn^{2+}, LCEs, Fe^{2+},

150   ethene, ethane, S^{2-}, CH_4 and acetate) and ammonium (NH_4). Biogeochemical cycling and OHR will take

151   place when DOM concentrations exceed the stoichiometric requirements (non-limiting) for each TEAP

152   where the reaction timescales are less than the exposure time along reactive sediment flow paths (Haest

153   et al., 2011; Zarnetske et al., 2012; Abbot et al., 2016). Hyporheic sediments are often enriched in

154   complex organic matter (Fig. 1) derived from allochthonous terrestrial ecosystem (such as soil particles

155   and leaf litter) and anthropogenic sources of dissolved-phase and suspended particulate material (e.g.

156   urban runoff and wastewater discharges) (Stelzer et al., 2014; Atashgahi et al., 2015). The nature of the

157   organic material that accumulates in riverbed sediments may also influence the effective residence time

158   of the plumes through hydrophobic partitioning to sedimentary organic matter (SOM) (Conant et al.,

159   2004; Allen-King et al., 2002; Wang et al., 2013).

160   Anaerobic decomposition of particulate organic matter (POM) is initiated by extracellular

161   hydrolytic enzymes produced by primary fermenters which release soluble DOM components capable

162   of traversing bacterial cell walls (Fig. 2) (Mani et al., 2016). Hydrolysis is frequently the rate-limiting

163   step in sedimentation zones with high POM loading rates, large particle size ranges or where highly

164   polymerised complex POM (e.g. lignin) are present that are not readily degradable (Gavala et al., 2003;

165   Nogaro et al., 2007; Atashgahi et al., 2014). DOM hydrolysis is a source of labile dissolved organic

166   carbon (DOC) and dissolved organic nitrogen (DON) species including organic acids and the soluble

167   oligomers and monomers of proteins and carbohydrates which influence the onset and rate of anaerobic

168   respiration processes (Jakobsen and Postma, 1999; Zarnetske et al., 2011; Mineau et al., 2013; Helton

169   et al., 2015). Groundwater typically contains <0.15 mM of DOC (Lapworth et al., 2009; Chapelle et al.,

170   2012). Of this, the labile fraction represents only 0.5–5% of carbon present which is often below

171   thresholds required for induction of catabolic genes (Egli, 2010). In contrast, hyporheic zone pore water

172   DOC concentrations often exceed 1 mM (e.g. Lewandowski and Nützmann, 2010) of which 10–30% is

173   labile and bioavailable (Baker et al., 1999; Vidon and Hill, 2004; Romani et al., 2006).

174   **2.2.2**   **Molecular hydrogen**

175   Free dissolved hydrogen (H_2) is produced from internal proton reduction during primary and

176   secondary DOC fermentation (Fig. 2) (Nath and Das 2004; Hallenbeck et al., 2009). H_2 serves as the

177   strict electron donor for known obligate OHRB (such as *D. mccartyi*) as well as many facultative OHRB

178   (Jugder et al., 2016). A generalised reaction for H_2 production during fermentation of DOC is given as

179   follows:

180     $DOC + 2H_2O \rightarrow DIC + 2H_2 + SCFA + H^+$
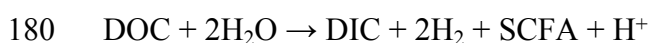
181   Where SCFA is a short-chain fatty acid which is discussed in the next section. In anoxic sediment zones,

182   DOC fermentation is a metabolic process utilised by common fermentative bacteria such as

183   *Desulfovibrio* (Walker et al., 2009), *Syntrophomonas* (Sieber et al., 2010), and *Clostridium* (Wu et al.,

184   2012), among others. H_2 is also produced by homoacetogens (Diekert and Wohlfarth, 1994), certain

185   acetoclastic methanogens (Heimann et al., 2007) and *Geobacter* spp. (Loffler and Sanford, 2005). Low

186   extracellular H_2 concentrations are required for DOC fermentation to be energetically favourable. This

187   is often maintained by unique syntrophic interspecies H_2 transfer relationships (Fig. 3) amongst

188   anaerobes which express membrane-bound hydrogenases (Walker et al., 2009; Morris et al., 2013;

189   Sieber et al., 2014; Jugder et al., 2015). As a result, the reaction timescales for H_2 uptake are typically

190   on the order of minutes (Heimann et al., 2009). Under electron-donor limiting conditions, ambient H_2

191   concentrations in sediments are often at steady-state that is determined by the hydrogenotrophs with the

192   lowest threshold concentration (Lovely and Goodwin, 1988; Fennel et al., 1997; Loffler et al., 1999;

193   Hoelen and Reinhard, 2004). H_2 threshold ranges have been reported for specific TEAPs (Fig. 4) ranging

194   from <0.1 nM for denitrification to >350 nM for homoacetogenesis. From Fig. 4 it can be observed that

195   HCE reduction (0.6–0.9 nM) occurs at an overlapping H_2 threshold to Fe(III) reduction (0.1–0.8 nM)

196   whereas LCE reduction overlaps with sulfate reduction and methanogenesis (Paul et al., 2016). Overlap

197   of TEAP H_2 thresholds are commonly observed in experimental studies (e.g. Jakobsen and Postma,

198   1999; Aulenta et al., 2008; Paul et al., 2016) with lower levels than predicted from thermodynamic

199   considerations (Heimann et al., 2009). This phenomenon has been attributed to kinetic effects arising

200   from the relative efficiencies in H_2 producers and consumers in microbial consortia (Richardson, 2016).

201 ### 2.2.3   Short chain fatty acids

202   Short-chain fatty acid (SCFAs) are labile DOC species such as butyrate, propionate and lactate

203 produced as intermediates during primary DOC fermentation (Fig. 2) and are accompanied by an

204 increase in pore water H^+ ion concentrations. SCFAs readily undergo secondary fermentation to produce

205 electron donors (Giovannini et al., 2016). SCFA turnover is often the rate-limiting step in the anaerobic

206 decomposition of complex DOM (Mani et al., 2016) and release of reducing equivalents for OHRB and

207 other heterotrophs (Atashgahi et al., 2014). The most frequently reported SCFA in hyporheic zones is

208 acetate which is both a terminal metabolite and intermediate of DOC fermentation (Fig. 2) (Baker et al.,

209 1999; Baker and Vervier, 2004; Hlaváčová et al., 2005). It is an easily assimilated carbon source and

210 possible direct electron donor for OHRB which may be independent of extracellular H_2 (Richardson,

211 2016). Under high pore water H_2 concentrations (>350 nM) acetate is also produced from other labile

212 DOC compounds (i.e. methanol) or H_2/CO_2 by homoacetogens such as *Acetobacterium*, *Sporomusa*,

213 *Spirochaeta* (Ziv-El et al., 2012).

214   Acetate has been widely shown to sustain respiration of HCEs as far as cDCE by facultative

215 OHRB groups such as *Geobacter* and *Desulfuromonas* (Sharma and McCarty, 1996; Krumholz et al.,

216 1996; Loffler et al., 2000; Wagner et al., 2012). For example, in contaminated aquifer media, Lee et al.

217 (2007) found that whilst H_2 could sustain dechlorination of PCE as far as ethene, with uptake of acetate,

218 dechlorination stalled at cDCE and proceeded at less than half the rate of H_2 alone. However, the role

219 of acetate in OHR appears to be more complex. He et al. (2002) found that several organohalide-

220 containing sediment microcosms amended with acetate alone were capable of sustaining PCE

221 dechlorination to ethene. Their findings suggest a potential syntrophic partnership between cDCE and

222 VC dechlorinators and acetate-oxidising bacteria (such as *Clostridium ultunense*) because of the low H_2

223 concentrations maintained by OHRB. Wei and Finneran (2013) also showed that acetate could sustain

224 TCE, cDCE and VC dechlorination to ethene. Little difference in dechlorination rates were observed

225 between amendments with 10 times more acetate. Similarly, higher production of acetate from solid

226 organic polymeric materials used to stimulate cDCE in riverbed sediment microcosms did not lead to

227 higher dechlorination rates as the excess organic loading was channelled to methanogenesis (Atashgahi

228 et al., 2014). These results highlight the complex role acetate plays in anaerobic food webs with

229   sometimes contradictory effects observed in relation to OHR rates.

2.3   Interactions between OHR and alternative anaerobic TEAPs

231   Interspecies H_2 transfers between fermenters and anaerobic TEAPs (Fig. 3) including denitrification,

232   metal reduction, sulfate reduction and methanogenesis (Fig. 4) may influence the rate and extent of OHR

233   in anoxic zones (Aulenta et al., 2007). Toxic metabolites generated may also inhibit steps of the

234   dechlorination process (Berggren et al., 2013). In this section, we review the current understanding of

235   these competitive interactions between OHR steps and biogeochemical cycling in the hyporheic zone.

2.3.1   Nitrate reduction

237   Groundwater nitrate has become elevated over the last century as a result of the 6.4-fold increase

238   in global inorganic fertiliser production between 1961 and 1999 (Peng et al., 2002). Heterotrophic

239   denitrification producing nitrous oxide (N_2O) or dinitrogen (N_2) and dissimilatory nitrate reduction to

240   ammonium (NH_4) occur in suboxic conditions (<125 μM O_2) with ambient H_2 levels (<0.1 nM) which

241   are below the level at which OHR becomes favourable for HCEs (~0.6 nM) (Fig. 4). Nitrate respiration

242   has only a slightly lower energy yield than that of oxygen with reaction timescales in the order of hours

243   to days (Rivett et al., 2008a; Jahangir et al., 2012) and has been widely documented in hyporheic

244   sediments (e.g. Ullah et al., 2014; Heppell et al., 2014). The interactions between OHR and nitrate

245   respiration is apparently complex. Yu et al. (2014) have shown that nitrate stimulated OHR of

246   pentachlorophenol at concentrations below 1 mM (likely by providing OHRB with a nitrogen source for

247   growth) but became inhibitory at concentrations greater than 1 mM. This was supported by Chen et al.

248   (2002) where 3 mM of nitrate inhibited hexachlorobenzene respiration under lactate-fermentation

249   conditions in river sediments. Nelson et al. (2002) found PCE reduction inhibition from only 0.6 mM of

250   nitrate and sulfate in a mixed culture where H_2 was continuously fed at non-limiting aqueous

251   concentrations (0.4–0.8 mM). In that study, the metabolite N_2O was found to be an inhibitor of OHR at

252   13 μM. These results are consistent with capture of available electrons by denitrifiers and reduction of

253   ambient H_2 levels below the minimum threshold necessary for OHRB as a key inhibition mechanism

254   which is also in line with thermodynamic predictions and observed H_2 thresholds (Fig. 4).

2.3.2 Iron(III) reduction

Iron oxides and oxyhydroxides such as haematite (Fe_2O_3) and goethite ($FeO(OH)_3$) are abundant primary and secondary minerals in streambed sediments which contain Fe(III) as a solid or colloidal phase (Liu et al., 2014). Similarly, manganese oxide minerals such as pyrolusite (MnO_2) contain the TEA Mn(IV) which behaves comparably to Fe(III) in redox processes (Ng et al., 2016). Some facultative OHRB including *Desulfuromonas michiganensis* and *Geobacter lovleyi* (Sung et al., 2006) can also use Fe(III) as a TEA. Dissimilatory Fe(III) reduction to Fe(II) has been observed at a similar H_2 threshold range (0.1–0.8 nM) to HCE reduction (Fig. 4) and direct competition for electron equivalents is therefore possible over reaction timescales measured months. Inhibition of TCE reduction as a result of Fe(III) was demonstrated in microcosm studies prepared from dolomite aquifer materials (Yager et al., 1997). Dupont et al. (2003) observed general inhibition of TCE reduction due to a large pool of bioavailable Fe(III) minerals in microcosms prepared from contaminated aquifer sediment. In contrast, Wei and Finneran (2011) demonstrated simultaneous reduction of TCE to ethene and Fe(III) to Fe (II) in microcosms dominated by *D. mccartyi* and *Geobacter* with an excess of electron donor. Fe(III) was observed to facilitate OHR by maintaining the ambient H_2 concentration at a level favourable for the OHRB. Using the 'KB-1' organohalide-respiring enrichment culture (Duhamel et al., 2002), Paul et al. (2013) showed that pH and iron mineral structure play an important role in inhibition of OHR in assays of 14 different synthetic Fe(III) minerals using an excess of formate as an electron donor. High-surface area, poorly crystalline minerals (such as ferrihydrite) have a greater Fe(III) fraction which is bioavailable for electron transfer than highly crystalline minerals such as haematite and goethite. These studies suggest the inhibition of OHR by Fe(III) in hyporheic sediments is complex and will be strongly linked to pH, specific bioavailable iron mineral composition, surface area and sediment particle size rather than total sediment Fe(III) content (Paul et al., 2013; 2016).

2.3.3 Sulfate reduction

Sulfate is a major oxyanion in groundwater derived from chemical denudation of sedimentary rocks, oxidation of sulfide minerals and anthropogenic point sources. Dissimilatory sulfate reduction to sulfide occurs at similar observed H_2 thresholds (1–15 nM) to HCE and LCE respiration (Fig. 4). The presence of sulfate has been widely implicated where incomplete dechlorination leads to the

283   accumulation of LCEs (e.g. Bagley and Gossett, 1990; Pavlostathis and Zhuang, 1993; Lorah et al.,

284   2007) although laboratory studies of interactions between sulfate reduction and OHR are inconclusive.

285   For example, sulfate reduction was reported to affect rates of VC respiration and to a lesser extent cDCE

286   respiration (Boopathy and Peters, 2001; Aulenta et al., 2008; Pantazidou et al., 2012). Hoelen and

287   Reinhard (2004) observed respiration of TCE at sulfate concentrations of 1–2.6 mM in sediment

288   microcosm containing CEs and alkyl benzenes. TCE reduction was found to inhibit sulfate reduction

289   under an ambient H_2 concentration of 0.7 nM. Sulfate reduction at higher H_2 levels (2.5 nM) was

290   observed to compete with reduction of cDCE and VC which was sustained at a H_2 concentration of 1.6

291   nM. With constant H_2 concentrations (2–4 nM) sustained by lactate fermentation, Berggren et al. (2013)

292   found that the addition of sulfate affected the cDCE to VC respiration step the greatest (67%) in

293   comparison to VC to ethene (25%) and TCE to cDCE (8%). This work showed that sulfate addition

294   could produce a shift in the community structure of *D. mccartyi* leading to a reduction in OHR rates. In

295   contrast, an absence of any inhibitory effect under non-limiting electron donor conditions were shown

296   (e.g. Hoelen et al., 2006; Aulenta et al., 2007) and even sulfate addition was reported to strongly enhance

297   TCE respiration (Harkness et al., 2012). This may be a result of syntrophic partnership between sulfate

298   reducers and OHRB through provision of essential nutrients such as corrinoids (Men et al., 2012, Sutton

299   et al., 2015; Lu et al., 2017 Atashgahi et al., 2017a). A recent study on isolates, constructed consortia

300   and enrichments containing *D. mccartyi* showed that rather than sulfate, sulfide inhibited the growth of

301   *D. mccartyi* and its syntrophic partner, *Syntrophomonas wolfei* (Mao et al., 2017). Interestingly, in a co-

302   culture of *D. mccartyi* and sulfate-reducing *Desulfovibrio vulgaris* Hildenborough, a high sulfate

303   concentration (5 mM) was not inhibitory to OHR under electron donor (lactate) limitation likely due to

304   strong electron scavenging capacity of *D. mccartyi*. Moreover, at low sulfate levels (2 mM), sulfate

305   reduction was not inhibitory even under excess of electron donor (Mao et al., 2017).

306   **2.3.4   Methanogensis**

307   Methanogenesis represents the final phase of anaerobic DOM metabolism (Figs 2 and 4) with

308   methane production timescales in riverbed sediments reported in the order of 200 days (Sela-Adler et

309   al., 2017). Its importance as a metabolic process in hyporheic zones is becoming increasingly recognised

310   (e.g. Sanders et al., 2007; Shelley et al., 2014; Brablcová et al., 2015). Methanogens are dependent on

311   the same key substrates and electron donors as OHRB including acetate (via acetoclastic

312   methanogenesis) and H_2 (through hydrogenotrophic reduction of CO_2). The effects of acetoclastic

313   methanogenesis on OHRB is less well understood than hydrogenotrophic methanogenesis. Heimann et

314   al. (2006) suggested that acetoclastic methanogenesis may enhance VC dechlorination by providing

315   additional H_2 during acetate cleavage. The observed H_2 thresholds for hydrogenotrophic methanogenesis

316   (5–100 nM) are considerably greater than OHR (0.1–2.5 nM) (Paul et al., 2016) and OHRB have been

317   shown to out-compete hydrogenotrophic methanogens for reducing equivalents at low H_2 concentrations

318   (Ballapragada et al., 1997; Yang and McCarty, 1998; Azizian et al., 2010). CEs on the other hand have

319   been shown to have direct inhibitory effect on methanogenesis through enzyme inactivation although

320   the effect is less pronounced than for other organohalides such as chlorinated methanes (Chan and

321   Radom, 2011). Yu and Smith (2000) found that 137 µM of TCE could inhibit methanogenesis but no

322   effect was observed from PCE at 87 µM. In enrichment cultures containing fermenters, *D. mccartyi*

323   strain 195, homoacetogens and methanogens, Men et al. (2013) found that 22 µM of TCE caused

324   inhibition of hydrogenotrophic methanogenesis with an increase of electron flow to OHR by an average

325   of 7% under non-limiting H_2 concentrations (7–12 nM). Their results suggested that TCE could inhibit

326   homoacetogenesis at H_2 levels well above the optimal range for OHRB. That study showed that non-

327   methanogenic cultures generated significantly more ethene faster than methanogenic cultures and

328   suggested that fermenters (*Clostridium* spp.) may also supply essential nutrients (corrinoids) in addition

329   to H_2.

330   **2.3.5   Co-mingled organohalides**

331   An important consideration in the anaerobic transformation potential of hyporheic zones for CEs

332   is the presence of other organohalides in co-mingled groundwater plumes. The chlorinated ethane 1,1,1-

333   trichloroethane (1,1,1-TCA) introduced as an alternative to TCE and the chlorinated methane,

334   chloroform, (TCM) are frequent co-contaminants (Scheutz et al., 2011; Simsir et al., 2017). TCM is also

335   produced from natural processes in soils at low concentrations (<10 µM) (Laturnas et al., 2002). Both

336   1,1,1-TCA and TCM have been implicated in the accumulation of LCEs and VC in particular (Duhamel

337   et al 2002; Chung and Rittmann, 2008; Chan et al., 2011). TCM has also shown to inhibit dechlorination

338   of cDCE by *D. mccartyi*. (Maymo-Gatell et al., 2001). Conversely, Grostern et al. (2009) showed that

339   VC and cDCE to lesser extent can cause inhibition in the respiration of chlorinated ethanes (CAs) (1,1,1-

340   trichloroethane and 1,1-dischloroethane) Mayer-Blackwell et al. (2016) found that prolonged exposure

341   to 1,2-dichloroethane (DCA) in a VC-respiring culture caused selective changes in *D. mccartyi*

342   community structure and reduced VC-respiration capacity. Strong inhibition of DCA respiration was

343   observed by cDCE. At field scale, Simsir et al. (2017) observed no inhibitory effect on the CE

344   dechlorination capacity of hyporheic sediments due to the spatial segregation of CE and CA

345   dechlorinators (demarcated by expression of the *cfrA* gene associated with CA respiration) along

346   discharging groundwater flow paths. CEs themselves may also competitively inhibit steps of the OHR

347   sequence (Chambon et al., 2013). In a kinetic study using two mixed anaerobic cultures, Yu et al. (2005)

348   found that HCEs generally inhibited the respiration of LCEs whilst the LCEs weakly inhibited the

349   dechlorination of the HCEs. These studies highlight the complex inter-relationship between different

350   organohalide respiration steps and their metabolites.

351   **2.4   Development and implications of anoxic-oxic interfaces**

352   The higher halogen substitution of HCEs produces electrophilicity with resistance to electrophilic

353   attack by oxygenase enzymes of aerobic bacteria. Thus, oxidation of the carbon backbone of the ethene

354   molecule is energetically unfavourable (Vogel et al., 1987). Schmidt et al. (2014) recently reported

355   aerobic TCE oxidation as a sole carbon source which may be evidence of a novel metabolic process not

356   witnessed before. Nonetheless, susceptibility to the reductive pathway decreases proportionately with

357   the ratio of chlorine to carbon substituents. As such, LCEs may be amenable to transformation via

358   aerobic mineralisation pathways during hyporheic exchange flows (Atashgahi et al., 2013). From a

359   natural attenuation perspective, mineralisation reactions are attractive in that no stable toxic

360   intermediates generated (Tiehm and Schmidt, 2011).

361   **2.4.1   Hyporheic oxygen gradients and mixing zones**

362   A unique feature of hyporheic zones is the establishment of oxic domains located where river water

363   infiltrates bringing dissolved O_2 into riverbed pore water at saturation concentrations (Knapp et al.,

364   2015). A common scenario illustrating development of a shallow oxic zone within a predominantly

365   anoxic groundwater regime is depicted in Fig. 5 based on the conceptual framework of Figs 1 and 2. In

366 this situation, river water advection is induced by bed topography in a process known as hyporheic

367 exchange flow where water currents infiltrate on the upstream side of a topographic feature

368 (downwelling zone) and return back to surface water a short distance downstream along with

369 groundwater discharge (upwelling zone) (Harvey and Bencala, 1993; Gomez-Velez et al., 2014; Trauth

370 et al., 2015). These bi-directional hyporheic flow cells (HFCs) will develop at various scales around bed

371 and channel topographic features such as gravel bars, riffles, weirs, debris dams and meanders (Krause

372 et al., 2014; Boano et al., 2014; Fox et al., 2014).

373 A thin mixing zone (Fig. 5); governed by dispersive processes is thought to develop along the

374 boundary between HFCs and the surrounding groundwater-dominated pore water regime which is

375 characterised by unidirectional vertical and lateral flow paths (Zarnetske et al., 2012; Briggs et al., 2013;

376 Binley et al., 2013; Boano et al., 2014). Numerical simulations of conservative tracer transport through

377 hydrologically gaining hyporheic zones (Hester et al., 2013) have suggested that the thickness of this

378 mixing zone is greatest with homogeneous high K sediment (e.g. well-sorted sand or gravel) and reduced

379 groundwater discharge (Fig. 5). Conversely, in zones of elevated groundwater discharge, low stream

380 velocity and smooth bed topography the mixing zone may be reduced or absent entirely (Trauth et al.,

381 2015). During high river flows, HFCs and associated mixing zones may transiently penetrate into deeper

382 sediment layers which is accompanied by a temporary surface water invasion of the groundwater regime

383 at depth (e.g. Cuthbert et al., 2010; Hamonts et al., 2012; Byrne et al., 2013; Freitas et al., 2015). These

384 transient flow events may be particularly acute for urban rivers where peak flows are accentuated by

385 impervious catchment areas and channel canalisation (Meyer et al., 2005).

386 Advection of river water solutes including O_2 and DOM into bed sediments induces zones of

387 aerobic respiration (Fig. 5) where the reaction timescale is often rapid with zero-order rates reported

388 from 9 and 75 µM/h in a gravel bar (Vieweg et al., 2016). Where mixing occurs, an anoxic-oxic interface

389 will develop which shifts dynamically in response to the balance of surface water downwelling and

390 groundwater discharge pressures. Approximately 0.25–0.33 mM of DOC is required to deoxygenate

391 downwelling river water and other reductants from the anoxic zone (Fig. 4) such as Fe(II) and reduced

392 sulfur species may also consume O_2. Hence, conditions for a spatially fluctuating 'reactive fringe'

393 (Trauth et al., 2014) where complementary aerobic co-metabolic or growth-coupled assimilation of

394 cDCE and VC may be favoured. As a result, hyporheic zones can offer a spatially variable secondary

395 aerobic 'treatment' zone facilitating complete mineralisation of LCE metabolites (Atashgahi et al., 2013;

396 2017a).

2.4.2 cDCE mineralisation

398 Although cDCE is amenable to aerobic biotransformation, cDCE oxidation is not widely reported.

399 Bradley and Chapelle (1998a) were among the first to examine the potential for aerobic cDCE

400 mineralisation in microbial communities indigenous to riverbeds. Their study demonstrated biological

401 aerobic mineralisation of [^{14}C]-DCE (4:1 *cis*-to-*trans* isomers) to CO_2 in microcosms with recovery of

402 [^{14}C]-CO_2 ranging from 17–100% after just eight days. However, this study used natural sediment media

403 potentially containing a range of other substrates where co-metabolic processes could not be ruled out

404 (Bradley and Chapelle, 2000). Other work has shown aerobic cDCE mineralisation to proceed much

405 more slowly in aquifer media e.g. Klier et al. (1998) where just 3−10% mineralisation was achieved

406 after 180 days incubation. Moreover, Abe et al. (2009) found no evidence of cDCE removal in aerobic

407 riverbed microcosms incubated over 1.5 years. Such patchy metabolic cDCE aerobic degradation is

408 illustrated by the fact that only a single isolate, *Polaromonas* sp. JS666 has unequivocally shown to

409 grow by cDCE assimilation as the sole carbon and energy source (Coleman et al., 2002). With the advent

410 of compound-specific stable isotope analysis (CSIA), kinetic isotope fractionation of carbon (δ^{13}C-

411 cDCE) has provided an additional line of evidence that cDCE mineralisation may be a growth-linked

412 assimilatory process. Reported δ^{13}C enrichment factors for aerobic uptake of cDCE range from −7.1‰

413 to −22.4‰ for both mixed and pure cultures (Tiehm et al., 2008; Abe et al., 2009; Schmidt et al., 2010).

414 Biomass yields from cDCE uptake are reported to range from 6.1 to 12.5 g/M cDCE (Coleman et al.,

415 2002; Schmidt et al., 2010).

2.4.3 VC mineralisation

417 VC mineralisation is the most thermodynamically attractive CE oxidation process due to the

418 presence of only one chlorine substituent. Reaction timescales tend to be relatively rapid as a result. In

419 a radiolabelled microcosm study prepared from riverbed sediments (Bradley and Chapelle, 1998a), it

420 was shown that greater than 90% of VC was consumed after 12 days over a concentration range of 0.2

421 – 57 µM with uptake rates most adequately described by Michaelis-Menten kinetics. Bacteria capable

422 of aerobic VC oxidation as a sole carbon and energy source belong to a range of genera associated with

423 aerobic ethene assimilation (Mattes et al., 2010; Atashgahi et al., 2017a). Aerobic degradation of VC

424 and ethene are initiated by an alkene monooxygenase (EtnABCD) and an epoxyalkane-coenzyme

425 enzyme (EtnE) (Mattes et al., 2010). Various isolates have been identified including *Mycobacterium*

426 (Hartmans et al., 1985; Hartmans and DeBont, 1992; Fullerton et al., 2014; Le and Coleman, 2011),

427 *Pseudomonas* (Verce et al., 2000; 2001; Atashgahi et al., 2017a), *Ochrobactrum* (Danko et al., 2004,

428 Atashgahi et al., 2017a), *Nocardioides* (Coleman et al., 2002) and a *Ralstonia* (Elango et al., 2006).

429 Besides these classical VC-assimilators, stable isotope probing coupled to high-throughput 16S rRNA

430 gene sequencing and quantitative polymerase chain reaction (qPCR) extended the potential range of VC

431 assimilators to *Sediminibacterium*, *Aquabacterium*, *Variovorax*, *Brevundimonas*, *Tissierella*, and

432 *Rhodoferax*. (Paes et al., 2015; Wilson et al., 2016). Gossett (2010) found that biological VC respiration

433 could be sustained by *Mycobacterium* at extremely low (microaerophillic) extracellular O_2

434 concentrations (e.g. 0.3–0.6 µM). This finding raises the important question as to whether anoxic

435 conditions classically defined by O_2 concentrations less than 3 µM (Chapelle and McMahon, 2006) are

436 in fact still capable of supporting aerobic metabolism. Microaerophilic VC oxidation may also play a

437 role in the apparent 'stalling' of the dechlorination sequence at cDCE. A lack of VC detection may in

438 fact reflect scenarios where VC mineralisation rates equal or exceed the production rate under hypoxic

439 conditions (Bradley and Chapelle, 2011). In a recent microcosm study by Fullerton et al. (2014) subject

440 to inadvertent O_2 contamination, VC disappearance was observed without ethene formation. A strictly

441 aerobic *Mycobacterium* was identified in high numbers in the groundwater microcosm which was

442 inadvertently exposed to O_2. Atashgahi et al. (2013, 2017a) used 16S rRNA of *D. mccartyi* and genes

443 encoding for reductive dehalogenase enzymes (*vcrA* and *bvcA*) and the genes *etnC* and *etnE* involved in

444 aerobic mineralisation of VC and ethene to track the fate of VC in microcosms prepared from riverbed

445 sediments obtained from anoxic-oxic interface. Using a combination of chemical analysis and qPCR,

446 the authors demonstrated co-occurrence and co-activity of aerobic VC degraders and anaerobic *D.*

447 *mccartyi* in hyporheic sediments of the eutrophic Zenne River in urban Belgium (Atashgahi et al.,

448   2017a).

2.4.4   Aerobic co-metabolism

450   CEs can be transformed under aerobic conditions by co-metabolic reactions initiated by common

451   enzymes produced by a wide range of aerobes (Arp, 1995). No known benefit is gained through

452   acquisition of metabolic energy or biomass production (Horvath 1972; Wacket, 1988; Semprini et al.,

453   1994). Co-metabolism is a widespread process by microbes containing non-specific monooxygenase or

454   dioxygenase enzymes which catalyze the initial step in oxidation of a growth-supporting substrate in the

455   presence of O_2 (Suttinun et al., 2013). A lack of enzyme specificity leads to competition between CEs

456   and growth-supporting substrates for active sites of oxygenases leading to incorporation of O_2 into the

457   chloroethene molecule via reactive epoxide formation (Mattes et al., 2010). TCE, cDCE and VC are all

458   amenable to co-metabolic aerobic oxidation by the action of oxygenases. PCE was originally thought to

459   be recalcitrant to aerobic oxidation under environmental conditions, however it is shown to be

460   degradable by the enzyme toluene-o-xylene monooxygenase (Ryoo et al., 2000; Shim et al., 2001).

461   In hyporheic sediments, methane and ammonium are often the most abundant potential substrates

462   for co-metabolism produced from the in-situ anaerobic decomposition of DOM (Figs 2 and 5) (Donn

463   and Barron, 2013; Atashgahi et al., 2013; Brablcová et al., 2015; Simsir et al., 2017). Extensive

464   hyporheic methanotrophy has been documented in association with fine-grained sediments under

465   vegetation stands in streams (Sanders et al., 2007) and in well-oxygenated coarse-grained sediments

466   (Trimmer et al., 2010). The enzyme methane monooxygenase is produced by the methanotroph

467   *Methylosinus trichosporium* and has been shown to oxidise TCE, cDCE, VC and ethene (Oldenhuis et

468   al., 1991; Forrester et al., 2005; Findlay et al., 2016). Conrad et al. (2010) observed evidence of co-

469   metabolic TCE oxidation at field-scale in a contaminated aquifer with an increase in *M. trichosporium*

470   abundance accompanied by a decline in TCE and dissolved methane concentrations. A co-metabolic

471   process was verified in a supporting microcosm study prepared from aquifer media and isotopically-

472   labelled TCE (^{13}C-TCE). Simsir et al. (2017) presented evidence of possible hyporheic zone co-

473   metabolism with declines in cDCE and VC associated with strong vertical methane gradients (Fig. 6)

474   although no O_2 data are presented for comparison. Their study also documented the presence of

475 methanotrophs (*Methylococcaceae* spp.) coincident with the change in methane concentrations in the

476 sediment sequence. Nitrifying bacteria which oxidise ammonium produced from DOM metabolism (Fig.

477 2) such as *Nitrosomonas europaea* can also inadvertently oxidise TCE and LCEs (Arciero et al., 1989;

478 Vannelli et al., 1990; Rasche et al., 1991; Kocamemi and Cecen, 2005). Nitrification is well documented

479 in association with hyporheic flow cells (Jones, 1995; Briggs et al., 2013). Kocamemi and Cecen (2005)

480 found that ammonium and TCE compete for the same active sites of the enzyme ammonia

481 monooxygenase. In a mixed culture study containing *Nitrosomonas europaea,* Kocamemi and Cecen

482 (2010) found that the transformation yield for TCE was strongly dependent on initial ammonium and

483 TCE concentrations, with a minimum initial level of 3 µM of ammonium required to transform the

484 maximum initial TCE concentration of 7 µM. We are not aware of any studies which have considered

485 hyporheic nitrification as a natural attenuation process for CEs under field conditions.

486 ### 2.4.5   Interaction between aerobic and anaerobic VC transformation pathways

487 Ethene-assimilating bacteria (ethenotrophs) can oxidise ethene as the sole carbon and energy

488 source and simultaneously transform VC co-metabolically whereas VC-assimilating bacteria can use

489 VC as the sole carbon and energy source (Mattes et al., 2010). Therefore, when anoxic groundwater

490 containing VC and ethene migrates through the mixing zones around HFCs, there is a potential for

491 assimilation of both substrates by ethenotrophs and VC-assimilators and also co-metabolic VC

492 degradation by ethenotrophs growing on ethene (Atashgahi et al., 2017a). Such synergetic interactions

493 can be further enhanced by methanotrophs that can oxidize both ethene and VC. In line with this, a

494 recent study using groundwater microcosms showed that when methane, ethene and VC were added to

495 microcosms, the rate of VC removal was faster than with either methane or ethene alone, consistent with

496 the idea that methanotrophs stimulate ethenotrophic removal of VC (Findlay et al., 2016). Moreover,

497 recent studies have shown that reductive dechlorination can impact aerobic VC degradation pathways

498 in hyporheic zones. Microcosms prepared from fine-grained, SOM-rich sediments with high reductive

499 dechlorination capacity did not have the potential for metabolic aerobic VC oxidation (Atashgahi et al.,

500 2013, 2017a). Under atmospheric O_2 conditions, *D. mccartyi* was protected from O_2 toxicity by the

501 sediment structure and grew by reductive dechlorination of VC. No ethene accumulation was noted

502 indicating the activity of ethenotrophs (Atashgahi et al., 2013, 2017a) which can, in turn, co-

503 metabolically oxidise VC (Findlay et al., 2016). In contrast, metabolic aerobic VC activity was observed

504 from microcosms prepared from sediments with low SOM, coarse grain size and low reductive

505 dechlorination potential. The study showed that local sediment geochemistry and reductive

506 dechlorination highly impact metabolic versus co-metabolic aerobic VC degradation pathway. Where

507 low reductive potential was noted, ethenotrophs exposed to continuous VC flow in oxic sediment layers

508 might adapt to metabolic aerobic VC degradation (Atashgahi et al., 2017a). These findings are consistent

509 with upon extended exposure to VC, ethenotrophic strains of *Mycobacterium* (Jin and Mattes, 2008) and

510 *Pseudomonas* (Verce et al., 2001) transit from co-metabolic to growth-linked VC mineralisation.

511 **3   Field-scale conceptualisation challenges and potential solutions**

512 **3.1   Spatial variability of potential transformation zones**

513 **3.1.1   OHRB occurrence in riverbed sediments**

514 qPCR approaches targeting 16S rRNA genes and reductive dehalogenation catabolic genes (*rdhA*)

515 (Lu et al., 2015) have been used to track the presence and activity of OHRB in hyporheic zones

516 (Vandermeeren et al., 2014). These biomarkers have revealed a close association with OHRB, SOM

517 content and grain-size distribution in organohalide-impacted hyporheic sediments (Atashgahi et al.,

518 2013; Kranzioch et al., 2013; Hamonts et al., 2014; Atashgahi et al., 2015). In a survey of OHRB in

519 relation to hexachlorobenzene (HCB) fate in 15 riverbed sediment locations across four European

520 catchments, Taş et al., (2011) detected *D. mccartyi* 16S rRNA genes at 80% of sites though their counts

521 did not surpass 0.1% of total bacterial 16S rRNA gene copies. HCB half-lives ranged from a few days

522 to up to one month. A relatively weak correlation was observed between decay rate and number of *D.*

523 *mccartyi* suggesting additional OHR activity from other OHRB. Atashgahi et al. (2013) used 16S rRNA

524 of *D. mccartyi* and genes encoding for VC reductive dehalogenase enzymes (*vcrA* and *bvcA*) to track

525 dominant VC biotransformation processes in microcosms prepared from hyporheic sediments. *D.*

526 *mccartyi* had a dominant role on VC removal in microcosms prepared from coarse-grained bed

527 sediments with an abundance of SOM. A subsequent in situ study at the same site showed a reduced VC

528 respiration potential due to construction of a wastewater treatment plant upstream of the study area that

529 resulted in a reduction of organic matter loading to the riverbed sediments (Atashgahi et al., 2015).

530 Similarly, PCR detection of *D. mccartyi* was reported to be closely linked with SOM content and

531 sediment particle size (Abe et al., 2009). This work showed that *D. mccartyi* presence in the sediment

532 profile was spatially correlated with SOM where it exceeded 2% (g/g) with stronger signals occurring

533 at greater depth. A recent study by Simsir et al. (2017) investigated CEs discharging to a creek in

534 Tennessee from organohalide-contaminated fractured bedrock (Table 1). Over a 300 m river section,

535 16S rRNA genes of *D. mccartyi* and *Dehalobacter* were found to be more abundant in deeper sediment

536 layers (50 cm) coincident with higher concentrations of cDCE and VC. Their study revealed the presence

537 of multiple strains of *D. mccartyi* occurring together and expressing functional genes involved in

538 respiration of CE (e.g. *tceA*, *bvcA* and *vcrA)* and chlorinated ethane/methane (*dcpA*).

539 ### 3.1.2 Plume-scale discharge and transformation controls

540 Groundwater plume discharge and transformation patterns are controlled by the spatial

541 distribution and continuity of sediment domains with distinct hydraulic properties along river corridors

542 (Fleckenstein et al., 2006; Wondzell, 2011). The available field investigations (Table 1) have shown that

543 low-K sediments (e.g. clay and silt) are the most important local controls on contaminant discharge in

544 heterogeneous sediment sequences. For example, at the Pine River PCE plume site in Ontario (Conant

545 et al., 2004; Abe et al., 2009) and for a TCE plume discharging to the River Tern in the UK (Weatherill

546 et al., 2014), the lateral and longitudinal continuity of semi-confining silty deposits were shown to

547 control the location and magnitude of plume discharge (Table 1, Fig. 6). Where these low-K sediments

548 also contain SOM, conditions for enhanced biotransformation may also occur where residence times are

549 greatly extended due to hydrophobic portioning to the organic carbon present. This is demonstrated by

550 Conant et al. (2004) and Abe et al. (2008) for the Pine River site with extensive in-situ transformation

551 of PCE or TCE to cDCE (Fig. 6) reported. The importance of low-K layers in contaminant

552 transformation has been illustrated in a modelling study by Gomez-Velez et al. (2014). Their work has

553 shown that the interfaces of sediments with large K contrasts can sequester upwelling groundwater by

554 creating stagnation zones which facilitate mixing of waters with different residence times. Hyporheic

555 exchange patterns arising from riverbed topography have also been shown to influence patterns of in-

556 situ transformation of discharging groundwater plumes. With the benefit of chloride as a conservative

557 surface water tracer, Freitas et al. (2015) observed TCE transformation associated with coarse grained

558   sediment along a riffle sequence of the River Tame (Birmingham, UK) where river water infiltration

559   carrying DOC was inferred from chloride profiles. In one location (Fig. 6) up to 80% TCE conversion

560   to ethene was reported. In-situ dechlorination was also shown to be largely absent along the investigation

561   corridor where riverbed topography was subdued.

562   The presence of preferential pathways with short residence times (such as springs) afford little

563   opportunity for in-situ natural attenuation (e.g. Fryar et al., 2000; LaSage et al., 2008; Rønde et al.,

564   2017). This mode of contaminant discharge arises where discontinuities occur in low-permeability

565   alluvial architecture that act as high-flux 'geological windows' (Conant et al., 2004; Weatherill et al.,

566   2014). For example, through temperature mapping, Conant (2004) found that nearly a quarter of the

567   baseflow accretion along a reach of the Pine River in Angus occurred from just 7% of the riverbed area.

568   The available case studies (Table 1) highlight the patchy nature of natural attenuation zones where the

569   mass fluxes to surface water are dominated by small areas through which most groundwater baseflow

570   discharge is concentrated.

571   ### 3.1.3   Pore-scale mass transport and processing

572   Microbial metabolic activity and diversity in riverbeds has been reported to be concentrated in

573   the uppermost 50–60 cm of hyporheic sediment sequences (Franken et al., 2001; Freixia et al., 2016).

574   Heterotrophic microbes (including OHRB) are predominantly found attached to sediment surfaces in

575   epilithic biofilms (Davey and O'Toole, 2000). Hyporheic biofilms comprise diverse assemblages of

576   fungi, algae and microbial consortia enveloped by a matrix of extracellular polymeric substances (EPS)

577   through which diffusive transport of solutes predominates (Storey et al., 1999; Battin et al., 2016).

578   Biofilm structures have been shown to retain and process complex DOM in sediments (Fischer et al.,

579   2005; Bengtsson et al., 2014) and may be important sites of SCFA fermentation (Rulík and Hereka,

580   1998; Rulík et al., 2000). Biofilm accumulation in interstitial spaces is known to decrease the bulk K

581   and effective porosity of sediments in downwelling zones, thereby reducing hydrodynamic exchange

582   with the overlying water column (Battin et al., 1997; 2003; 2008; Mermillod-Blondin et al., 2005) and

583   inducing redox gradients (Boano et al., 2014). Biofilm development may be facilitated in downwelling

584   streambed zones during the clogging of coarse sediment pore spaces (colmation) by non-settlable POM

585   originating from surface water sources (Navel et al., 2012).

586   The convolution of flow paths through heterogeneous biofilm-enveloped sediments produces a

587   scale dependence in chemical reaction rates where microbial processes are superimposed on larger-scale

588   advective solute fluxes that deliver reactants (Sobczak and Findlay, 2002; Findlay et al., 2003; Nogaro

589   et al., 2013). Mendoza-Lera et al. (2017) propose a conceptual framework unifying hyporheic flow and

590   biogeochemical transformation capacity in riverbeds which is applicable for both anaerobic and aerobic

591   biotransformation of CEs. They propose that biogeochemical processing is controlled by the mass

592   transfer of key reactants (TEAs, fermentable DOM and nutrients) and the sediment surface area

593   available for biofilm colonisation. Mass transfer is characterised in three stages, the slowest of which

594   will be the rate-limiting step for biotransformation: (1) transfer from bulk groundwater or surface water

595   to the hyporheic environment (2) transfer from sediment pore water into microbial biofilms and (3)

596   membrane transfer from the extracellular biofilm matrix into individual cells. The sediment particle size

597   controls the surface area for colonisation which is inversely proportional to K. High K sediments (such

598   as gravel bars) exhibit high mass transfer rates but low processing capacities. Conversely, low K

599   sediments (e.g. silts and clays) have very high areas for colonisation due to the huge surface area to

600   sediment volume ratio present but mass transport is slow and diffusion-limited. Their model assumes

601   that solute uptake rates follow Michaelis-Menten kinetics where microbial activity increases

602   proportionally with the mass transfer rate until a saturation point is reached (Ribot et al., 2013). This

603   framework supports field observations where enhanced dechlorination activity and OHRB are reported

604   in sediment domains where diffusive transport is locally dominant (Conant et al., 2004; Abe et al., 2009;

605   Damgaard et al., 2013; Atashgahi et al., 2015; 2017a). Biofilm colonisation is also influenced by changes

606   in pore water temperature, pH and light exposure as well as sediment surface roughness (Gette-Bouvarot

607   et al., 2015; Voisin et al., 2016).

608   Delivery of CEs into reactive diffusion-dominated pore water systems will be dependent on the

609   presence, connectivity and residence time of macropore flow paths which facilitates advective transfer

610   from bulk upwelling groundwater (e.g. Bohlke et al., 2007; Chambon et al., 2010; Menichino and Hester,

611   2015). Heterogeneity in grain size distribution, particle packing and sphericity can induce locally anoxic

612   or hypoxic microzones where mass transfer rates from bulk water to biofilms vary at the scale of

613   sediment pores (Briggs et al., 2013; 2014). Anoxic microzones have been widely implicated where

614   mixed TEAPs and fermentation reactions appear to occur simultaneously (e.g. Storey et al., 1999; Baker

615   et al., 2000; Mermillod-Blondin et al., 2005) and may explain observed co-activity of aerobes and

616   anaerobes (e.g. Atashgahi et al., 2017a). Using two-dimensional dual-domain pore network models of

617   advective-diffusive transport, Briggs et al. (2015) showed that pore-scale heterogeneity (variation in

618   pore throat size) leads to zones of increased and decreased solute mobility in hyporheic in hyporheic

619   sediments. In the less mobile domain (which is responsible for observed tracer breakthrough tailing),

620   anoxia develops where O_2 consumption exceeds the re-supply rate. These microzones may be

621   cumulatively important in determining the overall transformation capacity of hyporheic sediments. For

622   example, LCEs produced anaerobically in the low-mobility domain are mineralised in the more mobile

623   aerobic domain. This is particularly important in the case of VC which can require only trace oxygen

624   levels to sustain aerobic uptake (Gossett, 2010; Fullerton et al., 2014).

625   **3.2   Multi-scale in-situ characterisation technologies**

626   **3.2.1   Evaluating hydrological connectivity**

627   The natural flow regime heterogeneity of most plume discharge zones makes the design and

628   implementation of effective in-situ monitoring programs non-trivial (Kalbus et al., 2006; Rivett et al.,

629   2008b; Roy and Bickerton, 2010; Burk and Cook, 2015). Heterogeneous sediment sequences present a

630   common challenge for the development of conceptual models of plume fate (Ellis and Rivett, 2007).

631   Direct sediment core sampling can provide intact samples in cohesive sequences but is impractical where

632   granular deposits occur. Freeze-coring methods have been applied to characterise fluvial sediment

633   sequences with mixed grain size distributions (Freitas et al., 2015). However, this extractive technique

634   necessitates disruption of sediment microstructure and associated transport pathways (Descloux et al.,

635   2010). Non-invasive hydrogeophysical 'imaging' technologies have emerged which offer promising

636   insights into the in-situ spatial variability of mass transport pathways at <1 m to >10 m scales. For

637   example, Mermillod-Blondin et al. (2014) used ground-penetrating radar (GPR) to map biologically

638   active domains within a gravel bar sequence of the River Rhone in France. Electrical resistivity imaging

639   technology has been used to image flow-controlling sedimentary structures and electrically conductive

640 solute plume migration at reach-scale (Nyquist et al., 2008; González-Pinzón et al., 2015). Recently,

641 Briggs et al. (2014) applied geoelectrical hysteresis techniques to characterise pore-scale rate-limiting

642 mass transport through dual domain porous media (mobile and less mobile porosity fractions).

643   Temperature mapping using contact probes or continuous optical fibres and has been used to

644 evaluate patchy sub-metre scale connectivity between groundwater and surface water in plan-view when

645 a seasonal thermal gradient is present (Conant, 2004; Tristram et al., 2014; Rosenberry et al., 2016).

646 Groundwater baseflow fluxes through riverbeds have been modelled using one dimensional advective-

647 diffusive heat flow as a surrogate for Darcian flow in grids of riverbed temperature profiles (Schmidt et

648 al., 2006; 2007) where upwelling flow predominates. Both upwelling and downwelling fluxes have more

649 recently been obtained from vertical profiles of riverbed temperature time series (e.g. Anibas et al.,

650 2011; 2016; Munz et al., 2016). Heat pulse injections have been used to understand three-dimensional

651 water fluxes through shallow sandy hyporheic sediments at centimetre scales (Angermann et al., 2012).

652 Recently, point-velocity probing (PVP) has been used to map hydrological connectivity with a

653 comparable resolution to temperature mapping (Rønde et al., 2017). Overall, hydrological connectivity

654 will be highly site-specific and the most useful results will be obtained where multiple techniques are

655 employed with overlapping nested scales (e.g. Milosevic et al., 2012).

3.2.2   Resolving in-situ biogeochemical gradients

657   Delineation of in-situ vertical chemical gradients has been accomplished using depth-discrete

658 profiling in riverbeds (Fig. 6) over vertical length scales of 0.1 to 3 m (Conant et al., 2004; Rivett et al.,

659 2008b; Krause et al., 2013; Heppel et al., 2014). Multi-port monitoring wells have achieved finer vertical

660 sampling resolution (<0.1 m) in contaminated aquifers (e.g. Jobelius et al., 2011). However, technical

661 challenges and cost implications are often likely to prohibit monitoring installations in riverbeds which

662 require a drilling rig. Instead, hand-portable lightweight installations utilising drive-point and direct-

663 push methods usually offer the most cost-effective solutions (Rivett et al., 2008b; Roy and Bickerton,

664 2010). CE fate in hyporheic systems has been investigated using the 'waterloo profiler' (Pitkin, et al.,

665 1999; Conant et al., 2004) and other pore water samplers (e.g. Hamonts et al., 2012), multi-level sampler

666 bundles (Conant et al., 2004), nested piezometers (Lorah and Olsen, 1999) and drive-point mini-

667   piezometers (Conant et al., 2004; Ellis and Rivett, 2007; Roche et al., 2008). Rivett et al. (2008b) and

668   Typical drive-point methods utilise a temporary casing to protect screens and sampling inlets during

669   installation. Collapse of adjacent sediments is relied upon to provide sufficient sealing action to prevent

670   any preferential flow along this 'short-circuit' pathway. This limitation is overcome with direct-push

671   methods such as that described by Roy and Bickerton (2010). They developed a miniaturised waterloo

672   profiler optimised for rapid acquisition of riverbed sediment pore water using a portable hammer drill.

673   From Fig. 6 and Table 1, it can be seen that the selection of appropriate vertical monitoring scales is

674   site-specific and must be informed by an initial conceptual understanding larger scale water fluxes.

675   Recent field experience suggests vertical resolutions of 0.1 – 0.2 m may be sufficient to capture bulk

676   gradients in pore water chemistry (Fig. 6).

677   A potential monitoring gap exists at µm–mm scales below the sediment-water interface where

678   preservation of natural stratification in pore water chemistry is impossible with extractive sampling

679   techniques. Transformation boundary zones at thin fine-grained sediment strata and critical mixing

680   zones may be overlooked by sampling at coarser resolutions (Hester et al., 2017). In-situ passive

681   sampling technology utilising diffusive pore water equilibration offers a possible means to address this

682   scale gap. Dialysis 'peepers' can provide high-resolution solute profiles using a chambered sampler

683   design (Lewandowski et al., 2002; Tan et al., 2005) and have been used to investigate CE plume fate at

684   centimetre-scale (Lorah and Olsen, 1999; LaSage et al., 2008). Peepers require up to several weeks for

685   concentration equilibration to be achieved between ambient pore water and dialysis cell water during

686   which the sampler is vulnerable to disturbance and transient effects (Tan et al., 2005). Simsir et al.

687   (2017) used direct push diffusion samplers (Fig. 6) comprised of glass vials with polyethene or non-

688   woven porous fabric deployed into shallow creek sediment in Tennessee which were equilibrated for

689   two weeks. Passive sampling technology using hydrogel media requires much shorter equilibration

690   times (24-72 hours) and offers a promising way forward to address the mm monitoring scale gap.

691   Diffusive equilibration or gradient sampling in thin films (DET/DGT) has provided mm-scale depth

692   profiles of inorganic nitrogen species, phosphorus and metals in stream sediments (Davison et al., 1994;

693   Palmer-Felgate et al., 2010; Ullah et al., 2012). This technology has recently been applied to studies of

694   enhanced attenuation of groundwater nitrate in riverbeds influenced by in-stream vegetation growth

695   (Ullah et al., 2014). As Briggs et al. (2015) duly point out, extractive sampling procedures tend to always

696   bias the mobile porosity domain of reactive sediment zones which is overcome by diffusive thin film

697   approaches. This technology offers the best opportunity to capture vertical scales approaching the size

698   of pore networks where key reactive transport features such as anoxic microzones may be resolved. As

699   such, this technology can provide a nested observational capability when combined with larger scale

700   solute profiles for a more integrated understanding of plume behaviour.

701   Given that microbial biomass occurs predominantly as biofilms, microbial sampling in the

702   hyporheic zone is sensitive to particle size distribution given that the surface area available for

703   colonisation decreases with increasing particle size (Mendoza-Lera et al., 2017). Direct sampling of fine

704   sediment fractions for molecular analyses of microbial composition have been taken from surficial

705   sediment layers using piston corers (Atashgahi et al., 2012; Hamonts et al., 2014; Vandermeeren et al.,

706   2014) as well as high resolution sub-sampling of deeper cores (Abe et al., 2009; Atashgahi et al., 2014).

707   These methods are well suited to homogenous silt and clay sediments but are subject to the same

708   limitations for coring in granular deposits discussed above. Microbial 'trapping' using an emplaced

709   porous medium for biofilm colonisation is an established approach in aquifer media (Voisin et al., 2016).

710   At the Tennessee creek, Simsir et al. (2017) were able to measure in-situ changes in microbial

711   community structure at high spatial resolution using Bio-Sep bead traps (www.microbe.com) arranged

712   along a 55 cm vertical profile. By combining microbial and pore water sampling (Fig. 6), their work

713   permitted a comprehensive picture of microbial community structure and biogeochemical gradients to

714   be formed. This approach can offer unique insights on linkages between chemical and microbial factors

715   which govern in-situ biotransformation capacity and is suitable for deployment at a range of scales.

716   4   Conclusions and outlook

717   4.1   Conclusions

718   Our review has established the degree to which hyporheic zones can serve as natural bioreactors

719   capable of reducing mass fluxes of CEs to surface water from groundwater sources. Although OHRB

720   appear to be relatively widespread in riverbeds, field experience to date has shown that in-situ anoxic

721   biotransformation tends to be patchy (Table 1) where fine-grained or SOM-rich sediments occur (Fig

722   1). Observed transformation extents range from partial TCE to cDCE dechlorination to near-complete

723   stoichiometric conversion of TCE to ethene (Fig. 6). In DOM-dominated foodwebs, OHR requires

724   residence times along discharging groundwater flow paths to exceed the reaction timescales for higher-

725   energy TEAs such as O_2 and nitrate. Here, extracellular H_2 will be closely regulated by interspecies

726   transfers between fermenters and heterotrophs (Fig. 3) such as OHRB and methanogens. Competitive

727   interactions between dechlorinators and mineral reducing TEAPs (particularly sulfate) may inhibit OHR

728   steps, particularly cDCE to VC and VC to ethene (Fig. 4). When H_2 is limiting, OHR may compete with

729   these TEAPs for reducing power. Toxic metabolite formation (e.g. sulfide) may cause inhibition when

730   H_2 is unlimited. Conversely, syntrophic partnerships may develop between OHRB and other

731   hydrogenotrophs which may enhance in-situ biotransformation.

732   Thin mixing zones developing between discharging groundwater and downwelling surface water

733   carrying O_2 may create a dynamic anoxic-oxic interface potentially supporting ethenotrophs,

734   methanotrophs and ammonia oxidisers (Fig. 5). Here, a secondary niche environment for VC

735   mineralisation and to some extent, cDCE via direct metabolism and co-metabolism (from the action of

736   mono-oxygenase enzymes) can complete a coupled anoxic-oxic natural 'treatment' process. The oxic

737   potential of CE attenuation in hyporheic zones has only in the last few years become appreciated but is

738   not widely demonstrated at field scale. The efficacy of this step in the process will be dependent on the

739   presence of suitable aerobes and the thickness of mixing zones relative to discharging groundwater flow

740   paths transporting LCEs and co-substrates such as methane and ammonium produced from DOM

741   metabolism.

742   For both anoxic and oxic zone processes, mass transformation will be controlled by the sediment

743   surface area available for biofilm colonisation, delivery rates of critical reactants (e.g. H_2 or O_2), their

744   reaction timescales and exposure time in reactive zones. Diffusion-limited mass transfer from bulk pore

745   water to biofilm domains is likely to be the most important rate-limiting step governing in-situ

746   biotransformation efficacies. At pore scales, heterogeneity in pore throat size can induce dual mobility

747   porosity domains. Solute exposure timeframes are increased in the less mobile domain where anoxic

748   microsite formation may play an important role in the cumulative impact of processes.

749   Effective in-situ monitoring of hyporheic processes in plume discharge zones necessitates a

750   detailed conceptual understanding of the spatial and temporal variability of water flows and residence

751   times. This is best accomplished using non-invasive hydrogeophysical technologies and the application

752   of heat as a surrogate for advection where overlapping spatial scales may be observed simultaneously.

753   Bulk chemical gradients may be delineated using multi-level profiles with vertical resolutions of 0.1–

754   0.2 m often proving sufficient. Nested monitoring is warranted at critical boundary zones where the

755   local properties of pore networks may exert important cumulative controls on observed

756   biotransformation capacities. Simultaneous microbial and pore water sampling can demonstrate that

757   niche conditions for biotransformation are occurring in-situ and is advocated further.

758   **4.2   Future research opportunities**

759   Additional field experience is required to better address spatial and temporal biotransformation

760   variability and allow predictive models to be developed. Directions to improve the current conceptual

761   understanding are outlined as follows:

762   •   Integrated field studies addressing coupled flow and reactive transport along pathways at

763   multiple scales are needed (Oldham et al., 2013; Pinay et al., 2015; Abbot et al., 2016). New in-

764   situ tools have emerged which can quantify residence times at high spatial resolution (e.g. Rønde

765   et al., 2017). Dual reactive-conservative 'smart' tracer studies such as acetate-bromide (Rinehart

766   et al., 2015), acetate-resazurin-resorufin (Briggs et al., 2013) which can quantify both residence

767   time and reaction timescales for TEA-DOC couples offer an integrated way forward.

768   •   Capturing in-situ spatial and temporal dynamics of O_2/H_2 gradients along discharging

769   groundwater flow paths. These key end-members alone can provide the critical diagnostic

770   information on redox status and indications of key tipping points and priming events which

771   govern in-situ anaerobic-aerobic biotransformation efficacies.

772   •   Finally, these recent advances should be coupled with microbiological studies to better

773   understand the ecophysiology of anaerobic and aerobic CE-transforming microbes and their

774   interaction with biotic and abiotic factors. Overall, interdisciplinary efforts are necessary to

775   enhance understanding of hydrological, chemical, physical and microbial interactions and

776   critical gradients that influence the fate of CE in hyporheic zones.

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1586   **Fig. 1.** Generic conceptual model of a groundwater CE plume discharging through heterogeneous

1587   hyporheic zone sediments with a range of solute residence times. The plume may be composed of HCEs,

1588   LCEs or a mixture of both. Dashed line indicates the approximate boundary between an upper oxic zone

1589   and lower suboxic/anoxic zone induced by buried sedimentary organic matter. OM: organic matter; K:

1590   sediment hydraulic conductivity. Brown areas in inset circles represents buried sedimentary organic

1591   matter (adapted from Gomez-Velez et al., 2014).

1592   **Fig. 2.** Schematic representation of hyporheic zone electron flow derived from complex organic matter

1593   metabolism (adapted from Heimann et al., 2009) to organohalide-respiring bacterial communities in

1594   anoxic riverbed sediments. Processes depicted include: orange arrow: hydrolysis; green arrows:

1595   fermentation; brown arrows: carbon assimilation; dashed red arrow: ammonification; dashed green

1596   arrow: methanogensis; blue arrows/lines: electron transport and red arrows: organohalide respiration.

1597   Ac: acetate; DIC: dissolved inorganic carbon; ETH: ethene. Representative Gibb's free energy ($\Delta G°$)

1598   values (kJ/M) indicated for sequential reduction of PCE to ethene taken from Dolfing and Beurskens

1599   (1995) with representative Eh values from Wiedemeier et al. (1998) under standard conditions reported

1600   therein.

1601   **Fig. 3.** Influence of extracellular hydrogen concentrations on growth rates of fermenters and

1602   hydrogenotrophs (redrawn from Ter Meer et al., 1999).

1603   **Fig. 4.** Idealised ecological succession of terminal electron acceptors and their reduced metabolites in

1604   hyporheic zones based on observed molecular hydrogen (H_2) threshold concentrations. Ac: acetate;

1605   ETH: ethene. Gibb's free energy ($\Delta G°$) values (kJ/M) from Heimann et al. (2009) with representative

1606   Eh values from Wiedemeier et al. (1998) under standard conditions reported therein. Observed minimum

1607   hydrogen threshold ranges: (1) Cord-Ruwish et al. (1988); (2) Haring et al. (1991); (3) Sung et al. (2006);

1608   (4) Lu et al. (2001); (5) Mazur et al. (2001); (6) Lovely et al. (1989); (7) Caccavo et al. (1992); (8)

1609   Luijten et al. (2004); (9) Lovely et al. (1985).

1610   **Fig. 5.** (a) Conceptual illustration of potential metabolic (black arrows) and co-metabolic mineralisation

1611   of cDCE and VC (red arrows) at the dispersive mixing zone between upwelling anoxic groundwater

1612   containing DOM by-products (ammonium and methane) and oxic hyporheic flow cells (blue arrows).

1613   DIC: dissolved inorganic carbon; ETH: ethene. Note that reactive intermediates from co-metabolism are

1614   omitted for clarity. (b) Numerical simulations of oxygen concentration and uptake rate under gaining

1615   flow conditions in a dune bedform (adapted from Trauth et al., 2014). (c) Simulation of bedform mixing

1616   zone thickness with varying sediment hydraulic conductivity (K) and vertical groundwater flux (q_z)

1617   using a conservative groundwater tracer dilution approach (adapted from Hester et al., 2013).

1618   **Fig. 6.** Vertical pore water chemical gradients from selected case studies illustrating a range of in-situ

1619   transformation capacities (refer to Table 1 for profile context and supporting information). Black

1620   diamonds: TCE; open squares: cDCE; open triangles: VC; green circles: ethane; blue circles: ethane;

1621   orange squares: methane. Redrawn from: (a) Weatherill et al. (2014); (b) Abe et al. (2009); (c) Freitas

1622   et al. (2015) and (d) Simsir et al. (2017).

Table 1: Field experience of chlorinated ethene plume discharging to surface water receptors illustrating a range of in-situ transformation potentials from selected case studies (continued overleaf).
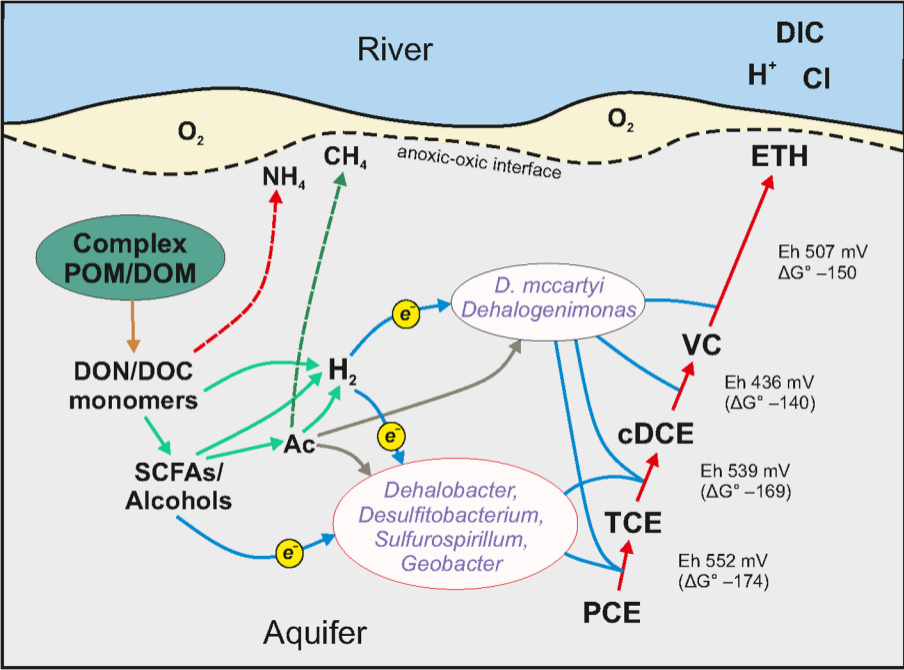
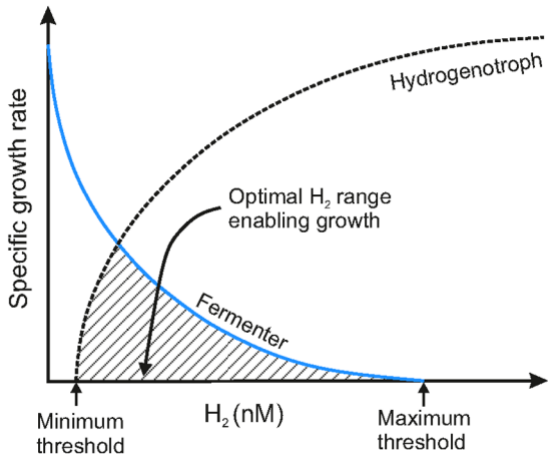| Receptor	CE source	Aquifer	Plume	Mass flux to surface water (kg/y)	Residence time (d/m)	In-situ biotransformation potential	References
Grindsted Stream, Jutland, Denmark.	Pharmaceutical factory in operation from 1914 to 1999.	Quaternary and Tertiary sands, underlain by thick regional clay layer, natural gradients.	cDCE (47 μM) and VC (72 μM). Co-mingled hydrocarbons.	45–123 (cDCE), 42–123 (VC)	0.4–10	**Low**, although not specifically addressed. CE-impacted stream, also impacted by metals and other organic contaminants.	Rønde et al., 2017; Sonne et al., 2017.
Third Creek, Knoxville, Tenessee, USA.	Spills associated with historic metal processing facility operated for over a century.	Fractured/karstic dolomite. Overlain by silty-clay saprolite, natural gradients.	PCE and TCE (6-60 μM). Co-mingled 1,1,1-TCA, TCM, hydrocarbons.	16.1 (TCE + cDCE)	58	**High.** No CEs detectable in surface water. Stochiometric cDCE transformation in SO_4-reducing conditions and potential CH_4 cometabolism. Hydrocarbon co-contamiants source of SCFAs. Well distributed *D. mccartyi* population.	Simsir et al., 2017.
River Tame, Birmingham, UK.	Multiple historic urban/industrial sources, City of Birmingham.	Unconfined Permo-Triassic sandstones. Complex urban environment. Gradients historically affected by abstractions.	TCE + cDCE (0.25–9.5 μM). Measured in riverbed piezometers.	20–200 (TCE) over 7.4 km city reach	12–100 (sorption modified)	**Transiently high**, low-moderate overall. Stochiometric TCE transformation to ethene in riffle sequence under Mn(IV) reduction. HEF and buried SOM a source of DOC driving redox. Inhibition by high SO_4 (5 mM) suspected.	Freitas et al., 2015; Roche et al., 2009; Ellis and Rivett, 2007.

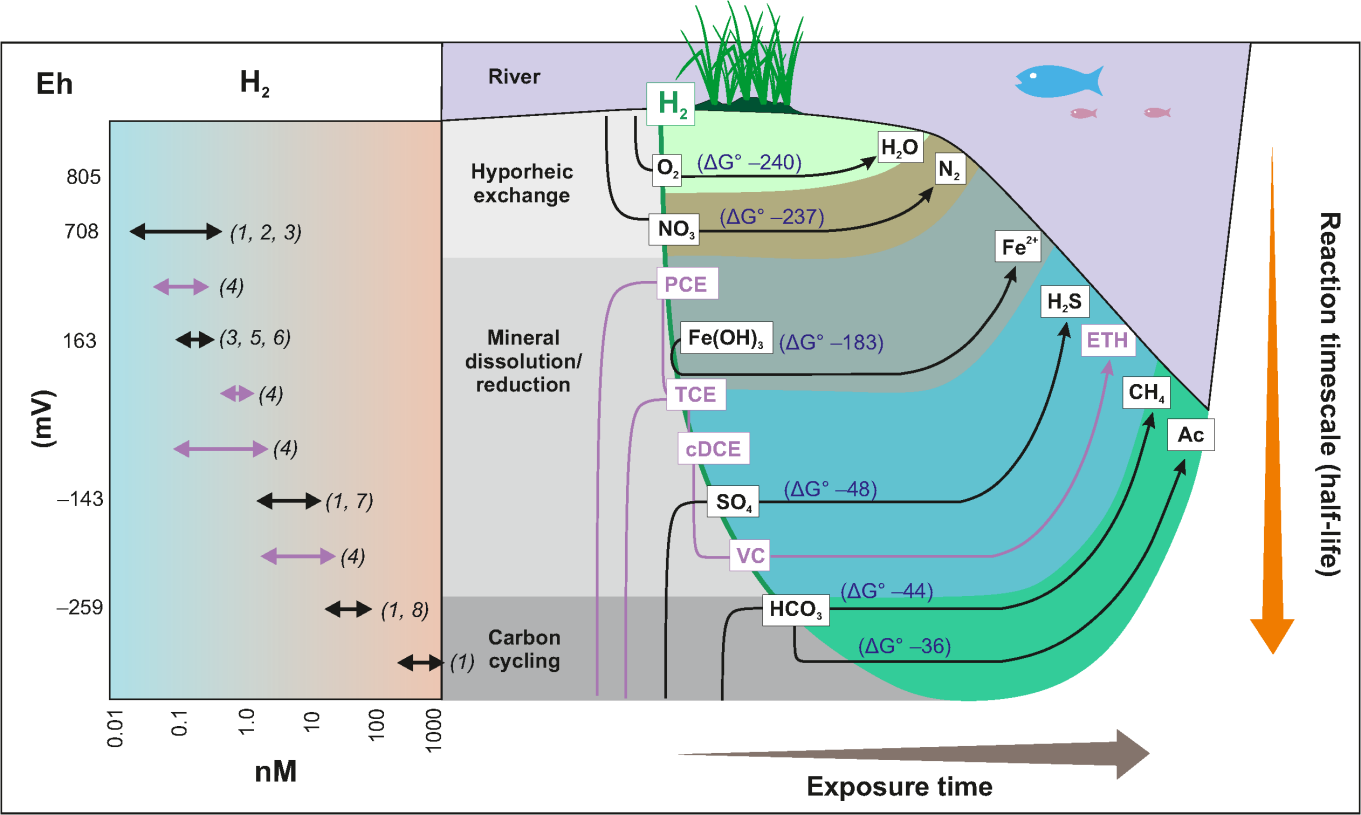|							
Zenne River, Vilvoorde, Belgium	Multiple industrial sources, Vilvoorde area.	Unconfined superficial gravel aquifer in urban setting. Gradients affected by retaining walls and canal water seepage.	VC (35 µM), cDCE (12 µM). 1.4 km wide plume.	Not addressed.	12–29	**High**. Transformation of cDCE and VC in up to 75% of locations. SOM-enriched sediments due to eutrophic conditions. *D. mccartyi* populations associated with SOM. VC/Ethene oxidisers present in SOM-poor zones.	Hamonts et al., 2009; 2012; Atashgahi et al., 2013; 2017a.
River Tern, Shropshire, UK.	Unknown. Suspected to be associated with a millitary airfield 1.5 km away.	Unconfined Permian sandstone. Natural gradients, occasional abstractions, agricultural setting.	TCE (0.1–1.4 µM).	0.15–0.55 (TCE)	0.5–1081 (sorption modified)	**Low** to **moderate** (patchy). Mass flux dominated by preferential pathways. Local TCE to cDCE hotspots associated peat lenses and surficial sediments. Elevated NO_3 in aerobic aquifer.	Weatherill et al., 2014; Weatherill, 2015.
Skensved Stream, Sjaelland, Denmark.	Leaking TCE storage tank, auto lacquer shop in operation since 1974. Leak discovered in 1993.	Interbedded soil, sand and gravel overlying bryozoan limestone. Pump-and-treat plume control in operation.	Max TCE (0.9 µM)	33 (TCE modelled from DNAPL source)	Not addressed	**Low,** aerobic to denitrifying conditions in hyporheic zone. Stream CE impact controlled by pump-and-treat system (as of 2009).	McKnight et al., 2010.
Pine River, Ontario, Canada.	Dry cleaning facility 195 m from river. PCE DNAPL contamination originated in the 1970s.	Semi-confined interbedded sand and clays in urban setting. Natural gradients.	Max PCE (0.25 mM)	3–53 (PCE) during 1995-1999.	858–71175 (sorption modified in silt/clay). Locally 3–23.	**High.** 52% of plume transformed to cDCE in riverbed. Stoichiometric ethene production in Fe(III)/SO_4 reducing SOM zones supporting *D. mccartyi* populations . Widespread aerobic VC mineralisation potential.	Conant et al., 2004; Abe et al., 2009.
Little Bayou Creek, McCracken County, Kentucky, USA.	Uranium enrichment facility, TCE used from 1953 to1996 for cleaning. Residual DNAPL volume of 795 m^3.	Regional gravel aquifer locally confined by silt and clay. Channelised river flow regime.	TCE near solubility limit near source (8.4 mM). 0.12–13 µM in riverbed. Co-mingled ^{99}Tc.	3-57 (TCE)	Not addressed	**Low**, TCE impacted surface water. No LCEs detectable in aerobic riverbed conditions with high mass flux through springs. Strong seasonal variability in stream water TCE concentrations.	LaSage et al., 2008; Fryar et al., 2000.

Unnamed Triburatary Streams, Industrial Facility, Conneticut, USA.	Manufacturing facility, 1970s underground degreaser tanks. TCE DNAPL source isolated in 1994. Plumes sustained by back-diffusion.	Unconfined surficial aquifer underlain by extensive aquitard. Steep topography, visible groundwater seepage zones.	TCE (18–24 µM), cDCE (5 µM) at Transect 2.	10 (TCE + cDCE) from Transect 2.	Not addressed	**Unknown.** Hyporheic zone processes in the streambeds not specifically investigated.	Chapman et al., 2007.

Atmosphere

OM O₂

River

Hyporheic Zone

Aquifer

High K **sand/gravel** migrating bedforms in close hydraulic contact with surface water. Short residence times.

Oxic zone

Low K **silt/clay** lenses with diffusion-limited transport and OM-rich laminations. Long to very long residence times.

Suboxic/anoxic zone

Moderate K heterogenous **alluvial** sediments with variable grainsize & OM content. Intermediate residence times.

G r o u n d w a t e r   C E   p l u m e

Specific growth rate

H_2 (nM)

Hydrogenotroph

Optimal H_2 range
enabling growth

Fermenter

Minimum
threshold

Maximum
threshold

(a) Atmosphere — O₂ — Surface Water — DIC, NO₂, Cl — Aerobic respiration — Mixing zone — Complex POM/DOM → DON/DOC monomers → SCFAs/Alcohols — NH₄, CH₄ — cDCE → VC → ETH — Anaerobic respiration

(b) O₂ mM — 0.3 0.2 0.1 0 — O₂ uptake mM/L/d — 0.4 0.2 0

(c) Groundwater Tracer — 0 3 6 9 mg/L — K = 25 m/d — qz = 0.55 m/d — K = 84.4 m/d — qz = 1.09 m/d — K = 150 m/d — qz = 5.47 m/d
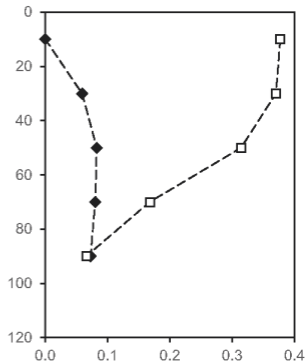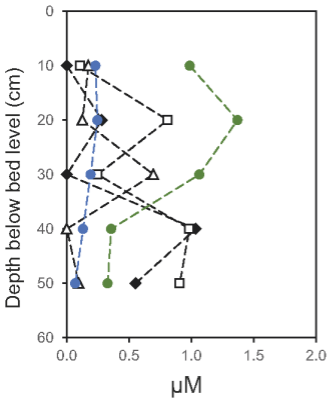
(a) **River Tern**, Shropshire, UK

(b) **Pine River**, Ontario, Canada

(c) **River Tame**, Birmingham, UK.

(d) **Third Creek**, Tennessee, US.