A review of methods for measuring groundwater-surface water exchange in braided rivers

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Abstract. Braided rivers, while uncommon internationally, are significant in terms of their unique ecosystems and as vital freshwater resources at locations where they occur. With an increasing awareness of the connected nature of surface water and groundwater, there have been many studies examining groundwater-surface water exchange in various types of waterbodies, but significantly less research has been conducted in braided rivers. Thus, there is currently limited understanding of how characteristics unique to braided rivers, such as channel shifting; expanding and narrowing margins; and a high degree of heterogeneity affect groundwater-surface water flow paths. This article provides an overview of characteristics specific to braided rivers, including a map showing the regions where braided rivers are mainly found at the global scale: Alaska, Canada, the Japanese and European Alps, the Himalayas, Russia and New Zealand. To the authors' knowledge, this is the first map of its kind. This is followed by a review of prior studies that have investigated groundwater-surface water interactions in braided rivers and their associated aquifers. The various methods used to characterise these processes are discussed with emphasis on their effectiveness in achieving the studies' objectives and their applicability in braided rivers. We also discuss additional methods that appear promising to apply in braided river settings. The aim is to provide guidance on methodologies most suitable for future work in braided rivers. In many cases, previous studies found a multi-method approach useful to produce more robust results and compare data collected at various scales. Given the challenges of working directly in braided rivers, there is considerable scope for the increased use of remote sensing techniques. There is also opportunity for new approaches to modelling braided rivers using integrated techniques that incorporate the complex river bed terrain and geomorphology of braided rivers explicitly. We also identify a critical need to improve the conceptual understanding of hyporheic exchange in braided rivers; rates of recharge to/from braided rivers; and historical patterns of dry and low-flow periods in these rivers.

1   Introduction

Until recently, groundwater and surface water systems were often considered separately both in research and in the way they were managed as resources (Kalbus et al., 2006; Winter et al., 1998). However, understanding the interactions between groundwater and surface water is now recognised as crucial to effective water resource management (Brodie et al., 2007). These systems are connected, so the development or contamination of either groundwater or surface water will often affect the other (Rosenberry and LaBaugh, 2008). Pumping from wells that are hydraulically connected to surface water bodies can result in, for example, reduced flows in rivers or diminished lake levels, or cause surface water inflow to groundwater (Stefania et al., 2018). Locations where groundwater and surface water interact can serve as contaminant transport pathways (Chadwick et al., 2002).

Groundwater seepage into surface water can provide important nutrients and temperature regulation for aquatic organisms (Hayashi and Rosenberry, 2002). Key questions in groundwater-surface water investigations are the location and flux of groundwater discharge to surface water bodies, and conversely, surface water recharge to groundwater. These questions can be considered at various spatial and temporal scales (Fleckenstein et al., 2010; González-Pinzón et al., 2015; Magliozzi et al., 2018).

This paper often refers to groundwater-surface water exchange, which in this context may include regional groundwater exchange with river water, as well as hyporheic zone exchange. The hyporheic zone consists of the sediments surrounding a river that are permeated by stream water (Boano et al., 2014). Hyporheic flow consists of river water that enters the hyporheic zone and re-emerges at the surface at some location downstream (Boano et al., 2014). Groundwater also may mix with surface water in the hyporheic zone (Boano et al., 2014). Hyporheic zone flow is multi-directional and may occur at multiple time and spatial scales (Cardenas, 2015). It is critical to note that there have been very few studies examining the hydrology and conceptualisation of the hyporheic zone in braided rivers. This is a crucial gap in knowledge, as this often limits our ability to interpret data collected from braided rivers relating to groundwater-surface water exchange.

This article investigates the methods that have previously been used for examining groundwater-surface water exchange in braided rivers and discusses scope for new methods to be applied. Braided rivers are a highly dynamic type of river with meandering channels, wide bars and variable flow levels. Globally, braided rivers are relatively rare; they are mainly found in the Canadian Rockies, Alaska, the Himalayas, New Zealand, Russia and the European and Japanese Alps (Figure 1) (Tockner et al., 2006; Alexeevsky et al., 2013). There are instances of braided rivers at locations outside of these regions (e.g., the U.S., Scotland, Iceland, China, Poland, Belarus, Colombia, Congo, Brazil, Paraguay, Argentina, and the Touat Valley in Africa); however these locations are not shown in Figure 1 because, at a global scale, they are not where braided rivers are mainly found. The regions displayed in Figure 1 are regularly cited in literature on braided rivers as the main regions where this river type can be found (Hibbert and Brown, 2001; Tockner et al., 2006). Braided rivers generally occur in mountainous areas with a large sediment source (such as glacial outwash), high river discharge rates and a steep topographic gradient (Charlton, 2008). These high-energy environments enable the rivers to carry large sediment loads. When these rivers reach their capacity to carry sediment, they form gravel braids, which branch out and re-join, creating gravel islands and shallow bars (Figures 2 & 3). Bars and islands are often

80 referred to as distinct features, with bars existing at periods of low flow, while islands are generally more

81 permanent features that may be vegetated (Charlton, 2008). Braided rivers can completely change their

82 geometry over a few decades. They undergo expansion and contraction phases in which their channels widen or

83 narrow, depending on sediment supply and river flows (Piégay et al., 2006). The wetted channels of the river

84 can shift, abandoning channels and re-occupying old channels (Charlton, 2008). Relatively erodible

85 streambanks, which allow for wide channels to form and meander, are a key characteristic of braided rivers.

86 These rivers generally have gravel beds but sand-bed rivers such as the Brahmaputra-Jamuna, which begins in

87 the Himalayas and flows through India and Bangladesh (and is the world's largest braided river), can also form

88 braided patterns (Sarker et al., 2014). The Brahmaputra-Jamuna is the only braided river in this review that is

89 not a gravel-bed braided river. Also, it is important to note, the specific rivers discussed in this article are all

90 braided rivers unless otherwise mentioned.

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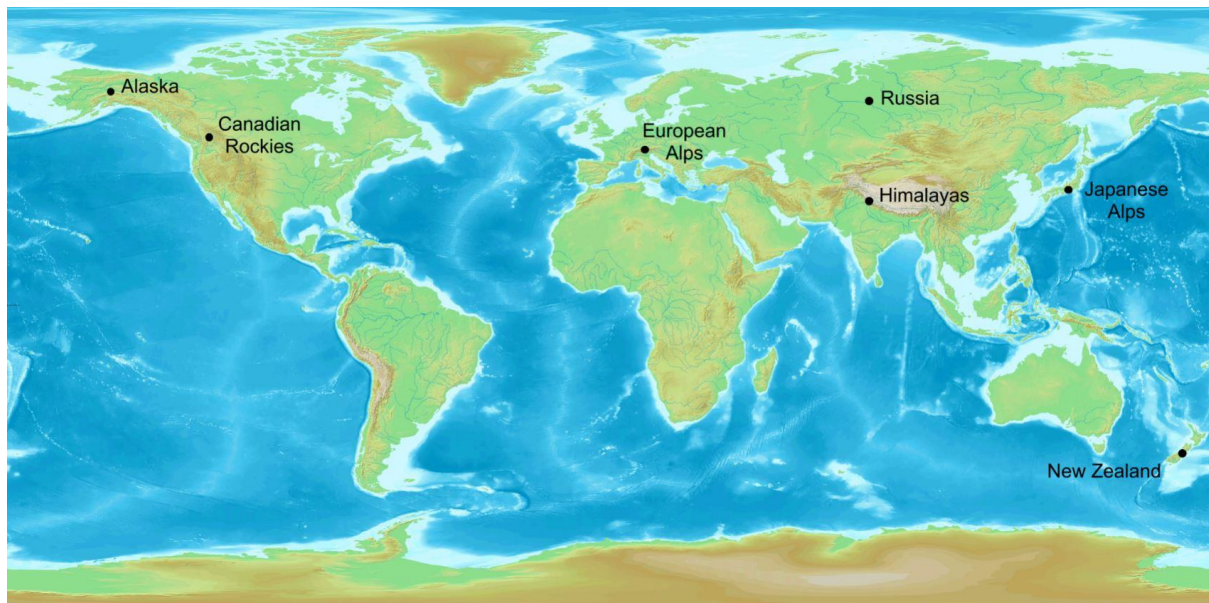
92 Figure 1. Locations where most braided rivers occur globally. Map base layer image attribution: "World Map-A

93 non-Frame" is licensed under CC BY-SA 3.0.

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96   Figure 2. Rakaia River in New Zealand displaying a classic braided pattern. Image reproduced with permission

97   by Andrew Cooper.

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100   Figure 3. The Rakahuri/Ashley River in New Zealand displaying a typical braided river consisting of multiple

101   channels, gravel bars and vegetated islands. Photo: Katie Coluccio.

102

103   Braided river deposits have formed extensive aquifers throughout the world including many in the regions

104   shown in Figure 1 (Brown, 2001; Huggenberger and Regli, 2006). The complex depositional processes of

105   braided rivers create heterogeneous aquifer properties (Huggenberger and Regli, 2006), and a significant portion

106   of flow occurs at varying scales in preferential flow paths formed by previous river flow channels (Close et al.,

107   2014; Dann et al., 2008; White, 2009). The complexity of braided rivers and their underlying heterogeneous

108   aquifers makes managing these systems in an integrated manner, that accounts for surface water-groundwater

109   interaction, challenging. For example, there is significant uncertainty surrounding rates of groundwater recharge

110   from large braided rivers in New Zealand, which complicates the sustainable allocation of water extraction

111   rights from surface water and groundwater sources (Close et al., 2014). There is also limited knowledge of how

112   hyporheic flow processes operate and how they impact river flow levels and water quality in braided rivers.

113   Braided rivers also often have reaches that become dry or have very low flows at the surface. The historical

114   patterns of these drying and low-flow periods, and the impact of groundwater-surface water exchange on this, is

115   an area of research where improved knowledge is needed. For example, many irrigation schemes have

116   artificially raised groundwater levels due to land surface recharge, or lowered groundwater levels due to

117   abstraction in comparison to their pre-irrigation states. In some rivers this has affected their losing/gaining

118   patterns (Burbery and Ritson, 2010; Riegler, 2012).

119

120   Braided rivers around the world have ecological, cultural, social, economic and recreational importance. They

121   provide habitat for many plant and animal species specifically adapted to survive in the dynamic, nutrient-poor

122   environment of the rivers' gravel bars and their margins (Kilroy et al., 2004; Tockner et al., 2006). In New

123   Zealand, the rivers are some of the last remaining native habitat on the heavily modified Canterbury Plains of

124   the South Island, thus serving a vital ecological purpose for plant and animal species, many of which are

125   critically endangered (Caruso, 2006; Williams and Wiser, 2004). Braided rivers and their associated aquifers are

126   also important freshwater resources used for drinking water supplies, irrigation, stock water and hydropower. In

127   many areas, these rivers hold significant cultural, social and recreational value for their importance for food

128   gathering, boating and swimming, and as places of outstanding natural character.

129

130   However, braided rivers face pressure from many angles. In many places they are subject to damage from

131   vehicles, gravel extraction, invasive plant species, development on river margins, damming, land encroachment,

132   containment through flood engineering, low-flow levels and poor water quality (Caruso, 2006; Larned et al.,

133   2008; Tockner and Stanford, 2002). These factors can influence river processes in many ways, including

134   altering the rate of sedimentation or changing the flow regime, which may impact various uses of these rivers, as

135   well as riparian ecosystems (Piégay et al., 2006).

136

137   Much braided river research has focused on understanding their geomorphological structures and processes,

138   such as sediment transport (e.g., Ashmore, 1993; Chalov and Alexeevsky, 2015; Huggenberger and Regli, 2006;

139   Nicholas et al., 2006). The majority of studies up to the early 1990s consisted of laboratory-based modelling of

140   the braiding process (e.g., Ashmore, 1982; Young and Davies, 1991) and field studies of small reaches of

141   valley-confined systems (Ferguson et al., 1992). Beginning in the mid-1990s, there were advances in numerical

142   models to estimate the braiding process in reaches, remote sensing, and the quantification of river morphology

143   and morphological change using digital elevation models (e.g., Bernini et al., 2006; Copley and Moore, 1993;

144   Doeschl et al., 2006; Huggenberger, 1993). This allowed, for the first time, the visualisation and analysis of the

145   morphology of large braided rivers (e.g., Hicks et al., 2006; Huggenberger, 1993; Lane, 2006). A number of

146   studies have looked at the surface water features of braided rivers (e.g., Davies et al., 1996; Meunier et al., 2006;

147   Young and Warburton, 1996), as well as aquifers created by braided river deposits (e.g., Huber and

148   Huggenberger, 2016; Pirot et al., 2015; Vienken et al., 2017). However, the connections between the two have

149   been less explored, particularly in regard to hydrology and conceptualisation of the hyporheic zone.

150

151   This article addresses this gap in the literature by reviewing methods previously used in braided rivers

152   internationally to characterise groundwater-surface water interactions, as well as recommendations for new

153   methods that can be applied in this type of river environment. The objective is to provide guidance for future

154   braided river studies. As described in this section, braided rivers have many features that may make it difficult

155   to apply techniques used in different river environments. While many of these features are found in other river

156   types, they exist in a particular combination in braided rivers, which make it problematic to investigate

157   groundwater-surface water exchange. The rapidly shifting channels of braided rivers make it difficult to

158   establish, maintain and access study sites. The typical coarse gravel substrate makes it challenging to install

159   instruments in the riverbed. Large braided rivers can be several kilometres wide, resulting in data collection

160   across the width of the river difficult or impossible. The very permeable gravel streambeds are often highly

161   gaining or losing in respect to groundwater, and these interactions can have large temporal variability. The

162   mixed sand and gravel substrate makes it nearly impossible to take undisturbed samples for sediment structure

163   analysis. The heterogeneous nature of the river substrate and structures—largely mixed sand and gravel, with

164   some clay and silt layers, and open framework gravels—make upscaling point-scale data difficult. A significant

165   portion of river flow occurs within the streambed; and in aquifers, the open framework gravels (i.e., paleo river

166   channels) serve as preferential flow paths. In relation to the methods used in previous studies, this article

167   examines the equipment and study design; cost; issues of temporal and spatial scales; and ultimately the

168   techniques' effectiveness. For general overviews of methodologies not specific to braided river applications,

169   refer to Kalbus et al. (2006); Brodie et al. (2007); Rosenberry and LaBaugh (2008); Lovett (2015); Rosenberry

170   et al. (2015); and Brunner et al. (2017).

171

172   **2   Methodologies for assessing groundwater-surface water interactions in braided rivers**

173   Various types of methods have been used to investigate groundwater-surface water exchange in braided rivers

174   such as mass balance approaches; hydrochemical tracers; direct measurement of hydraulic properties; and

175   modelling. Many of these studies employed multiple methods to meet their objectives. To thoroughly and

176   clearly assess each method, the techniques, and their advantages and limitations will be discussed individually in

177   the following section, and the discussion section will review the merits and limitations of multi-method studies.

178   This information is then summarised in Table 1.

179

180   **2.1   Water budgets**

181   Some of the most commonly used methods for identifying gains and losses to braided rivers have been based on

182   a mass balance approach. The underlying principle of this method is that any gain or loss of surface water can be

183   related to the water source, therefore the groundwater component can be identified and quantified (Kalbus et al.,

184   2006). Many of these mass balance approaches have used water budgets to separate groundwater and surface

185   water components both on river-reach and catchment-wide scales.

186

187   River-reach water budgets involve estimating the net flux of seepage in a defined river reach by measuring

188   stream discharge in cross-sections and then calculating the difference in flow between the cross-sections

189   (Kalbus et al., 2006). If there is an increase or decrease in discharge, this can be considered as a gaining or

190   losing reach, respectively, provided any surface inflows or outflows (e.g., tributary inflows, abstractions) are

191   accurately quantified. Measurements should generally be taken in low flow conditions to avoid errors caused by

192   river flow recession after rainfall or snowmelt (Brodie et al., 2007).

193

194   Several studies have used river-reach water budgets to identify gaining and losing reaches of braided rivers. The

195   Selwyn River in New Zealand, which has losing and gaining reaches, and annually dries in parts, has been the

196   focus of several studies (Larned et al., 2008; Larned et al., 2015; Vincent, 2005). Both Larned et al. (2008) and

197   Vincent (2005) used flow gauging data to classify gaining and losing reaches of the river. Larned et al. (2015)

198   used a 30-year gauging record from two flow recorder sites on the river to calculate groundwater level lag times.

199   In another study, Farrow (2016) characterised gaining and losing reaches of the four major rivers in the Ashley-

200   Waimakariri zone in New Zealand using historic flow gauge records, however they cited the need for additional

201   concurrent flow gauging under mean flow conditions to more accurately characterise long-term gaining and

202   losing reaches. In an attempt to determine the causes of the perennial drying of the Ashburton River in New

203   Zealand, Riegler (2012) conducted flow gauging along the river in conjunction with groundwater well

204   measurements, mapping of dry reaches and regression analysis. White et al. (2012) used a steady-state

205   groundwater budget to estimate groundwater outflow from the riverbed based on the mean daily flow at a

206   recorder site on the Waimakariri River and groundwater level observations in a monitoring well array beside the

207   river. The authors found that river channel area rather than channel position was most important in their

208   calculations; however, they recommended that future research examine the effects of channel position and area

209   on groundwater outflow. This is particularly relevant in braided rivers, as their channel positions often change.

210   Both Simonds and Sinclair (2002) and Doering et al. (2013) used flow gauging as part of multi-method studies

211   for estimating groundwater-surface water interactions in the Dungeness River (Washington State, U.S.) and

212   Tagliamento River (northeastern Italy), respectively. These authors conducted concurrent gauging to calculate

213   the net loss or gain of flow along river reaches and compare to data collected from other methods.

214

215   A smaller number of braided river studies (e.g., Burbery & Ritson (2010)) have used catchment-scale water

216   budget calculations to estimate the inflow and outflow from braided river catchments and distinguish

217   groundwater from surface water sources. The underlying relationship is provided below (modified from Scanlon

218   et al. (2002)):

219    $inflow = outflow \pm \Delta S$ (1)

220   Here, inflow is the sum of precipitation, surface water inflow and groundwater inflow. Outflow is comprised of

221   actual evapotranspiration, surface water outflow and groundwater outflow. ΔS is the change in water storage in

222   the catchment. This also considers artificial changes to water levels in the catchment such as industrial

223   discharges to surface water or water abstraction. Burbery and Ritson (2010) calculated a water budget for the

224   Orari River catchment in Canterbury, New Zealand, which was based on field observations from various

225   methods including flow gauging and groundwater well observations, climate data and water use data. The

226   authors used the flow gauging data to classify gaining and losing reaches in four of the rivers in the catchment.

227   They noted that in order to obtain a greater level of detail about groundwater-surface water connectivity at the

228   local scale, shorter spaced flow gauging coupled with high-resolution piezometric surveys and aquifer pumping

229   tests should be carried out (Burbery and Ritson, 2010).

230

231   **Advantages and Limitations**

232   River-reach water budgets are useful for identifying hotspots of river gains and losses at a broad scale. However,

233   there are several issues regarding their effectiveness in braided rivers. As detailed in Section 1, these types of

234   rivers are typically comprised of heterogeneous materials and thus there may be small-scale interactions of

235   groundwater and surface water within reaches, which flow gauging is poor at identifying (Hughes, 2006). For

236   example, Larned et al. (2015) noted that lag time calculations can only highlight generalised flow paths, whereas

237   predicting more specific groundwater flow paths or residence times would require studies using additional

238   techniques such as tracers or potentiometric data. Also, accurate measurements of flow rates can be

239   compromised by several factors including interference of macrophytes in the streambed, low flow, imprecise or

240   shifting river margins, high sediment load, or unstable streambeds that permit parafluvial flow (i.e., flow in the

241   area of riverbed that is to some extent annually scoured by flooding (Stanford, 2007)). As noted by Close et al.

242   (2014), there is significant uncertainty around estimates of river to groundwater flows solely based on hydraulic

243   measurements, particularly for large braided rivers, as these environments provide various challenges for

244   accurate flow measurements. These systems are difficult to measure because precise flow gauging can only be

245   carried out during low flows and measurement errors can be considerable (Close et al., 2014). Often the

246   measurement error is greater than the net exchange of groundwater and surface water (LaBaugh and

247   Rosenberry, 2008).

248

249 Catchment water budgets can be a useful method at a larger scale but are generally not appropriate for assessing

250 small-scale groundwater-surface water interactions, as the accuracy of recharge rates to or from rivers is limited

251 by the accuracy of the measurement of the other components in the budget (Scanlon et al., 2002). They can be

252 simple and quick to calculate, but this depends on how time consuming or expensive the data collection is. Also,

253 this method can have low resolution because of the limited number of flow gauging stations on rivers (Kalbus et

254 al., 2006). Thus, when calculating budgets for large catchments, the errors can be significant.

255

256 **2.2   Hydrochemistry**

257 There are various natural physical and chemical properties of groundwater and surface water that can serve as

258 indications of interaction between the two systems. A variety of tracers have been used in braided rivers to

259 investigate groundwater-surface water exchange including geochemical tracers such as conductivity, chloride or

260 alkalinity; stable isotopes; and radioactive isotopes such as radon. At sites where there is a discernible difference

261 between the groundwater and surface water concentrations of one of these parameters, the influence of

262 groundwater or surface water may be able to be detected. This type of analysis assumes there is an evenly

263 distributed groundwater concentration between sampling locations and that there is complete mixing of water

264 sources (Lovett, 2015). To separate surface water or groundwater components, mixing models based on

265 conservation of mass are used (Kalbus et al., 2006), such as End Member Mixing Analysis (EMMA) or

266 hydrograph separation. The methods presented below represent the majority of known braided river applications

267 to date, and thus this is not a complete list of all tracers used in previous studies. Some additional tracers applied

268 in braided river settings not discussed in detail here include dissolved oxygen (e.g., Larned et al., 2015; Rodgers

269 et al., 2004), silica (e.g., Botting, 2010; Rodgers et al., 2004; Soulsby et al., 2004), nitrate (e.g., Burbery and

270 Ritson, 2010; Larned et al., 2015; White et al., 2012) and sulphate (e.g., Acuña and Tockner, 2009; Botting,

271 2010).

272

273 **2.2.1 Stable isotopes**

274 Oxygen, which is a key component of water, naturally occurs in three stable isotopic forms: mainly as oxygen-

275 16 (^{16}O), and in smaller proportions as oxygen-17 (^{17}O) and oxygen-18 (^{18}O) (Sharp, 2007). Due to the

276 difference in mass between these isotopes, they undergo fractionation during evaporation and condensation

277 (Taylor et al., 1989). The process is largely driven by temperature, humidity and salinity, whereby precipitation

278 is increasingly depleted in ^{18}O at colder temperatures (which tend to occur at higher elevations) (Sharp, 2007).

279   The ratio of ^{16}O to ^{18}O (referred to as $\delta^{18}O$) is used to identify the relative concentrations of the two most

280   abundant stable oxygen isotopes. This allows for the identification of groundwater recharged by alpine sources

281   and lowland rainfall (Burbery and Ritson, 2010) and can shed light on groundwater flow paths in aquifers.

282

283   Several studies have used $\delta^{18}O$ to characterise groundwater-surface water exchange in braided rivers and their

284   associated aquifers. Blackstock (2011) found their isotopic model for the Christchurch, New Zealand

285   groundwater system matched well with previous physical mass balance calculations and that stable isotope

286   analysis was useful, especially in shallow groundwater. Botting (2010) found that stable isotope analysis was the

287   most effective technique for distinguishing surface water from groundwater amongst the multiple methods that

288   they used (including hydrochemical sampling, pumping tests, and groundwater well observations) in a study of

289   the north bank of the braided Wairau River in New Zealand. In addition, Vincent (2005) successfully used $\delta^{18}O$

290   analysis to identify groundwater recharge sources in the upper Selwyn River catchment. Burbery and Ritson

291   (2010) used $\delta^{18}O$ analysis to determine alpine versus lowland recharge sources for groundwater in the Orari

292   River catchment. Of the various methods used in the study (which also included flow gauging, a catchment-

293   scale water budget, chemical tracers and groundwater level observations), the authors found $\delta^{18}O$ analysis to be

294   highly effective for understanding groundwater-surface water interactions in the catchment. Given $\delta^{18}O$ varies

295   seasonally, they recommended sampling be carried out at various times during the year to obtain better temporal

296   resolution, as well as on a long-term basis to consider climatic variations. Hanson and Abraham (2009) carried

297   out $\delta^{18}O$ and other hydrochemical analyses along two transects across New Zealand's Canterbury Plains. The

298   authors found $\delta^{18}O$ to be the most reliable tracer to differentiate between land surface recharge and alpine river

299   water. However, they pointed out that a suite of tracers would be needed to characterise groundwater flow paths

300   and groundwater recharge sources. They also noted that $\delta^{18}O$ can be significantly altered where alpine water is

301   used for irrigation.

302

303   **2.2.2 Radon**

304   Radon-222 (Rn-222) is another useful tracer for identifying groundwater-surface water interactions. It is a

305   chemically and biologically inert radioactive gas part of the Uranium-238 decay process and is present in nearly

306   all rocks and soils (LaBaugh and Rosenberry, 2008). As water flows through rocks and soils it becomes

307   enriched in Rn-222. In surface waters, radon quickly degasses, so groundwater generally has Rn-222

308   concentrations three to four orders of magnitude higher than surface waters, thus making it an effective tracer in

309 many environments (Burnett et al., 2001). For example, an area of high radon concentrations in surface water

310 would suggest groundwater inflow. It is a cost-effective, simple technique that is suitable for study areas ranging

311 in size (Martindale, 2015).

312

313 Rn-222 analysis can address many questions related to groundwater and surface water interactions. In a multi-

314 method study in the braided Tagliamento River in northeast Italy, Acuña and Tockner (2009) used Rn-222 to

315 assess the residence time of upwelling groundwater in the hyporheic zone. Moore (1997) analysed Rn-222 to

316 estimate groundwater inflow to the Brahmaputra River in the Bay of Bengal. Close et al. (2014) used Rn-222

317 sampling to calculate the velocity of groundwater recharge from the Waimakariri River to groundwater in the

318 Canterbury Plains in New Zealand using the ingrowth (i.e., the rate of build-up in a closed system) equation for

319 Rn-222. The authors recommended that a high-resolution study with closely spaced sampling sites could be

320 useful for highlighting preferential flow paths in the riparian zone. In addition, Close (2014) sampled Rn-222

321 amongst other hydrochemical parameters in the Wairau River in Marlborough and in groundwater wells within

322 five kilometres of the river to better understand the groundwater-surface water interactions in the river and the

323 amount and variability of recharge to the groundwater system. Close (2014) found that temperature correlated

324 well with the spatial distribution of the radon but added that there could be significant errors with estimating

325 groundwater flow paths due to local heterogeneity and the meandering nature of the alluvial deposition process

326 in the area. Close (2014) recommended analysing temperature and data collected from piezometers in

327 conjunction with radon to resolve these uncertainties.

328

329 There are some limitations of Rn-222 analysis, as it requires several assumptions, including that stream water is

330 well mixed downstream of groundwater discharge areas; water fluxes are constant; the radon activity in the

331 stream water and groundwater are known and constant; and that there is no additional surface recharge from

332 sources such as streams or stock water races (Kraemer and Genereux, 1998). It also may be difficult to

333 distinguish between regional groundwater discharge and hyporheic zone exchange using radon analysis (Lovett,

334 2015; Martindale, 2015). Rn-222 concentrations will also vary with different mineral compositions in the rocks

335 present (Close et al., 2014).

336

337 **2.2.3 Chloride**

The chloride ion (Cl⁻) can be used as an indicator for groundwater and surface water mixing in locations with sufficiently distinct chloride concentrations in groundwater and surface waters. For example, the groundwater surrounding the Bow River in the Canadian province of Alberta has elevated levels of chloride from road salting. This allowed Cantafio and Ryan (2014) to measure chloride levels in an urban reach of the river and assess water quality impacts and baseflow sources. They found that nearly all river flow originates in the Rocky Mountains and there is little contribution from groundwater.

Chloride is frequently sampled amongst a suite of hydrochemical parameters to investigate groundwater and surface water interactions, as groundwater often becomes enriched in chloride as it passes through soil and rocks (Dommisse, 2006). Burbery and Ritson (2010) measured chloride concentrations in the Orari River catchment in New Zealand, specifically looking at chloride-to-sulphate ratios to delineate groundwater-surface water interactions and examine recharge sources in the catchment. They found that basic ion chemistry was useful for determining the extent of the Orari River water but noted that results can be complicated by hydrochemical changes due to land use activities. Several other studies measured chloride to determine recharge sources and quantities in braided rivers and their associated aquifers including Acuña and Tockner (2009), Larned et al. (2015), Botting (2010) and Domisse (2006).

2.2.4 Alkalinity

Alkalinity can serve as an effective indicator for determining catchment water sources. In a study of the braided River Feshie, in the Cairngorms in Scotland, Rodgers et al. (2004) used alkalinity as a tracer to investigate temporal changes in stream water hydrochemistry and characterise sources of river flow. The authors noted that Gran alkalinity is particularly useful as it serves as a directly measurable, close approximation to the acid neutralising capacity, which is considered a conservative chemical tracer. Gran plots are commonly used to determine alkalinity and acid neutralising capacity in water with low alkalinity or low conductivity. A Gran function plot identifies the point at which all alkalinity has been titrated in a strong acid-strong base titration (Rounds and Wilde, 2002). Rodgers et al. (2004) used EMMA to estimate different hydrological sources of River Feshie water. The authors were reasonably confident of their estimates because of the extensive temporal and spatial components of their study. Because of the relative simplicity and low cost of the Gran alkalinity method, these types of longer term and detailed spatial surveys are becoming increasingly feasible (Rodgers et al., 2004), though may be costly in terms of human resources required. In another study in the Feshie catchment,

368   Soulsby et al. (2004) conducted a geochemical tracer study to improve large-scale flow path understanding. The

369   authors carried out chemical-based hydrograph separations to separate baseflow from storm event sources. They

370   analysed for Gran alkalinity, which they noted was simple and inexpensive to measure. Alkalinity has proven to

371   be a useful parameter in the United Kingdom (UK) to distinguish between water sourced from acidic, organic

372   soils (which are common in the UK at shallow depths) and deep, older groundwater. Soulsby et al. (2004) found

373   their study provided valuable information at the sub-catchment scale, but more information was needed at finer

374   spatial scales.

375

376   **Advantages and Limitations**

377   Hydrochemistry can provide significant insight into both catchment-wide hydrology, as well as provide

378   estimations of seepage flux on the point scale (Close, 2014; Dommisse, 2006; Lovett, 2015). Even considering

379   catchment heterogeneity, some tracers can behave predictably enough to serve as effective tracers for studies of

380   braided rivers (Soulsby et al., 2004). Environmental tracers are useful in settings where there is a sufficient

381   difference between tracer concentrations in the groundwater and surface water, and some parameters can be

382   easily incorporated in long-term routine monitoring programs. Disadvantages of these methods include that

383   hydrochemistry of the baseflow and storm event water composition may be too similar, or that hydrochemistry

384   may not be constant in time or space (Genereux and Hooper, 1998). Importantly, various tracers such as

385   dissolved oxygen, pH, nitrate and sulphate may be affected by biogeochemical processes, so to be effective, the

386   tracers must be conservative at the scale of the investigation. Also, land use activities may alter hydrochemistry

387   in catchments, for example from fertiliser application or mixing of water sources through irrigation (Soulsby et

388   al., 2004). Additionally, some low tracer concentrations may cause analysis errors (e.g., in the case of radon)

389   (Close, 2014).

390

391   **2.3   Temperature studies**

392   Temperature has been used in a number of studies to characterise groundwater-surface water interactions in

393   braided rivers. In most locations, during winter and summer months, there is a discernible difference in

394   groundwater and surface water temperatures. In general, groundwater temperature is more stable, whereas

395   surface water temperatures change diurnally and seasonally (Kalbus et al., 2006). In summer, groundwater is

396   typically colder than surface water, whereas in winter, groundwater is generally warmer. Heat tracer methods

397   can be used to identify discharge and recharge zones as well as quantify the flux of water moving between

398 groundwater and surface water systems (Andersen, 2005). There are various methods involving temperature

399 sensing that range in complexity, scale and cost. One-off temperature readings can be taken using probes, or

400 sensors and data loggers can gather time-series data both in-stream or in groundwater wells. Vertical and

401 horizontal temperature profiles can also be measured by arranging sensors in a series either in-stream or in wells

402 on river margins. Temperature profiles can be analysed using various methods such as VFLUX (Gordon et al.,

403 2012) or the steady state approach (Schmidt et al., 2006). Some temperature methods, such as thermal infrared

404 imaging and fibre-optic temperature sensing (both of which are discussed further in Section 4), are best suited

405 for identifying patterns, such as temperature differences in surface water that may indicate areas of recharge or

406 discharge. Other methods such as temperature depth profiles can be used to quantify the flux of water through

407 the streambed.

408

409 The following studies demonstrate various applications of temperature measurement that have been used to

410 characterise groundwater-surface water exchange in braided rivers. Passadore et al. (2015) conducted thermal

411 monitoring to characterise the temporal and spatial variability of streambed water fluxes in the Brenta River in

412 Italy. They used heat as a tracer in conjunction with water level measurements and found this combination of

413 methods to be effective in estimating groundwater-surface water interactions. Two studies of the Wairau River

414 in Marlborough, New Zealand analysed temperature (Close, 2014; Close et al., 2016). Close (2014) measured

415 temperature in the river and in groundwater wells located near the river to characterise river recharge to the

416 aquifer. The author compared the data to Rn-222 analysis and found that the temperatures correlated well with

417 the spatial distribution of radon. Close et al. (2016) used the daily mean temperatures in groundwater wells to

418 estimate the lag time between the river and the observation wells. Lastly, Coluccio (2018) used VFLUX to

419 analyse diurnal temperature signals to characterise seepage through the streambed of a braided river. The study

420 determined the direction and magnitude of vertical seepage through the streambed using temperature probes in

421 the Ashburton River in New Zealand. The results were compared with hydrochemistry and water level

422 measurements in the river and shallow groundwater to better inform the interpretation of the temperature data.

423 Coluccio (2018) found that it was difficult to distinguish between shallow groundwater and hyporheic flow and

424 also noted that further studies would benefit from combining a point-scale method like temperature probe

425 analysis with broader scale techniques.

426

427 **Advantages and Limitations**

428   Heat tracers offer many techniques at varying spatial and temporal scales. Broad-scale methods like aerial

429   thermal infrared imaging can be used to obtain large-scale data, and they can offer the advantage of remote

430   collection of data in areas that are difficult to access. Point-scale techniques using temperature sensors can

431   indicate surface water-groundwater interactions at a specific location. Some methods of temperature analysis

432   can also quantify seepage flux (e.g., using diurnal signal analysis). The methods range in cost and complexity,

433   and thus can be tailored to suit a study's needs. There are some limitations including that a temperature gradient

434   between groundwater and surface water might not always be present (e.g., this may be affected by

435   environmental conditions such as season, wind, shade from vegetation or rapidly changing river levels)

436   (Johnson, 2003). Also, for certain types of analysis, temperature needs to be measured continuously (Irvine et

437   al., 2017). In addition, due to the dynamic nature of braided rivers and their associated sediments, heat transfer

438   within the heterogeneous materials may be non-linear.

439

440   **2.4   Darcy approach**

441

442   **2.4.1 Hydraulic gradient**

443   Groundwater levels are often used to aid in the understanding of groundwater-surface water interactions, and

444   there have been several studies conducted in braided rivers using this technique. Groundwater level data can be

445   used to identify the hydraulic gradient (i.e., the difference in hydraulic head over a given distance) at a location,

446   which can reveal groundwater discharge to a river or river recharge into an aquifer. The underlying principle is

447   that if groundwater levels in a well are higher than the river level, the river is gaining (i.e., groundwater is

448   flowing into the river). Conversely, where river levels are higher than the groundwater level in a nearby well,

449   the river is losing (i.e., river water is flowing into groundwater). It is worthwhile to note that it is important to

450   obtain a conceptual understanding of the relationship of the river to the water table, as the river might be

451   connected, disconnected or in a transitional state between the two (Brunner et al., 2009). Groundwater levels are

452   most typically measured using pressure transducers or electronic water level indicators.

453

454   The hydraulic gradient is calculated as $\Delta h/\Delta l$, where Δh [L] is the difference in hydraulic head [L] and Δl is the

455   distance between the points where the hydraulic head was measured. Hydraulic gradient can be measured in the

456   horizontal direction to characterise flows into or out of a river through the sides of the river. Here, Δh [L] is the

457   difference between the groundwater level in a well at the edge of the river and a well a distance Δl [L] away

458   from the edge of the river. Hydraulic gradient can also be measured in the vertical direction, to characterise

459   vertical flows into or out of the river through the streambed. In this case, Δh [L] is the difference between the

460   groundwater level in an in-river piezometer and the river level at that location; and Δl [L] is the distance from

461   the riverbed to the top of the well screen (Doering et al., 2013).

462

463   Once the hydraulic gradient has been measured, the magnitude of groundwater flow into or out of a river can be

464   estimated using the Darcy equation:

465   $Q = -KA\frac{\Delta h}{\Delta l}$ (2)

466   Where Q [L^3/T] is the volume of flow; A [L^2] is the cross-sectional area perpendicular to flow through which the

467   water passes; and K [L/T] is hydraulic conductivity (Schwartz and Zhang, 2003). For calculating the horizontal

468   flow magnitude, a horizontal hydraulic conductivity of the surrounding aquifer is generally used. To calculate

469   the vertical magnitude of flow, the vertical hydraulic conductivity of the streambed needs to be determined, as

470   does the streambed area over which the water exchange occurs (Simonds and Sinclair, 2002).

471

472   In terms of specific methods that can be used for measurements, existing piezometers (i.e., monitoring wells)

473   near rivers can be useful for conducting these types of studies, particularly given the often high cost of drilling

474   new wells. Please refer to a standard text such as Fetter (2001) for a definition of piezometers. Mini-

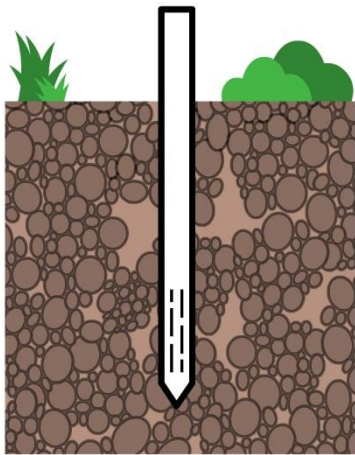
475   piezometers, which are scaled-down versions of piezometers and typically installed no deeper than about two

476   metres (Figures 4 & 5), have been previously used in studies of braided rivers (Acuña and Tockner, 2009;

477   Doering et al., 2013; Malard et al., 2001). We recommend referring to the studies mentioned in this section for

478   piezometer designs for braided river applications, as feasibility of installation into coarse gravel is one of the

479   significant limitations of this technique, and not all designs would be effective in braided rivers for this reason.

480

481   Figure 4. Conceptual diagram of a mini-piezometer (Coluccio, 2018).

482

483   Figure 5. Mini-piezometer installed on the bank of a braided river (Coluccio, 2018).

484

485   Previous studies have examined the correlations between groundwater levels and river levels to establish the

486   degree of connectedness of groundwater systems and braided rivers, for example, attempting to identify the

487   causes of drying reaches and changes in long-term river flows. Prior studies have been carried out in catchments

488   with substantial agricultural surface and/or groundwater abstraction for irrigation. Thus, the questions here are

489   often whether abstraction has caused drying in rivers or decreases in river flows, and what effect future

490 abstraction will have. These studies have often coupled groundwater level measurements with streamflow

491 gauging and physicochemical sampling of river water and groundwater. Riegler (2012) examined groundwater

492 levels, in conjunction with flow gauging, in the North Branch of the Ashburton River in Canterbury, New

493 Zealand to attempt to correlate groundwater levels and decreased flow levels in the river. The study concluded

494 that there were too many uncertainties, particularly around the complex behaviour of the groundwater system, to

495 draw strong conclusions on the causes of the drying riverbed. Several other studies also investigated New

496 Zealand braided rivers that are highly connected to groundwater using these methods (Larned et al., 2008;

497 Larned et al., 2015; Vincent, 2005; Coluccio, 2018).

498

499 A multi-method study was carried out on the Dungeness River in Washington State in the U.S. to characterise

500 groundwater-surface water interactions. Simonds and Sinclair (2002) installed mini-piezometers in the river in

501 which they measured the vertical hydraulic gradient between the stream and water table. They also continuously

502 monitored water levels and temperature in two well transects, providing data on the horizontal hydraulic

503 gradient and temporal changes in groundwater-surface water flows. The authors also conducted flow gauging

504 along "seepage runs" in the river to quantify the net gain or loss of flow over a reach.

505

506 Groundwater level measurements in mini-piezometers have also been applied in studies of European braided

507 rivers. Malard et al. (2001) calculated the difference in hydraulic head between hyporheic water and surface

508 water and between the parafluvial hyporheic zone and the river using mini-piezometers in the Roseg River in

509 Switzerland. Acuña and Tockner (2009) also incorporated groundwater level observations into their multi-

510 method study of the Tagliamento River in Italy. The used PVC mini-piezometers installed to a depth of 50 cm in

511 four reaches of the river. They calculated vertical hydraulic gradient to determine the direction and intensity of

512 surface and subsurface (i.e., hyporheic flow or groundwater) exchange in the streambed. In another study of the

513 Tagliamento River, Doering et al. (2013) installed mini-piezometers along 10 transects in losing and gaining

514 reaches of the river. Five mini-piezometers were installed horizontally across the river at each location and were

515 used to calculate the vertical hydraulic gradient where the piezometers were installed.

516

517 **2.4.2 Hydraulic conductivity**

518 As detailed above, the hydraulic conductivity of riverbeds is needed to calculate the magnitude of flow through

519 the riverbed. There have been a number of studies investigating the hydraulic conductivity of streambeds (e.g.,

520 Landon et al., 2001; Kelly and Murdoch, 2003), though few studies have been conducted in braided rivers.

521 There are many well-established methods for calculating hydraulic conductivity of a porous medium, including

522 grain size analysis, permeameter tests, slug and bail tests, and pumping tests (see Fetter, 2001).

523

524 In an early investigation of the permeability of gravel streambeds, Van't Woudt and Nicolle (1978) extracted

525 gravel from the bed of the braided Waimakariri River in Canterbury, New Zealand. They conducted lab-based

526 tests to determine hydraulic properties of the bed substrate such as porosity and infiltration rates. This study

527 resulted in several conclusions about sub-surface flow in gravel-bed rivers, including that fine sediments

528 flowing through the gravels tend to create a low-permeability clogging layer along the margin of and below the

529 riverbed. The authors also found horizontal permeability to be far higher than vertical permeability (30:1), but it

530 is difficult, if not impossible, to draw conclusions about horizontal and vertical conductivities once the sediment

531 is disturbed.

532

533 Cheng et al. (2010) carried out a study to determine the statistical distribution of streambed vertical hydraulic

534 conductivity at 18 sites along a 300-km reach of the Platte River in Nebraska. They conducted in-situ

535 permeameter tests using falling head tests and found that vertical hydraulic conductivity was normally

536 distributed at all but one of their study sites.

537

538 In a study on the north bank of the Wairau River in Marlborough, New Zealand, Botting (2010) conducted

539 pumping tests to determine groundwater flow paths and origins. The pumping tests were of limited use however,

540 because the pumping did not successfully lower the groundwater levels, most likely due to the high

541 transmissivity of the aquifer.

542

543 On the Ashburton River in New Zealand, Coluccio (2018) conducted slug tests in mini-piezometers installed on

544 the margins of the river. The hydraulic conductivity values calculated from the slug tests were on the low end of

545 the range for expected hydraulic conductivity values in this area, which may have been a reflection of the tests

546 being conducted in localised areas of finer sediments, highlighting the limits of using this point-scale method in

547 heterogeneous environments (Coluccio, 2018).

548

549 **Advantages and Limitations**

550   There are various benefits and drawbacks of the methods described in this section. Use of existing groundwater

551   wells may be very convenient in a study, but the installation of new deep wells generally comes at a high cost.

552   Mini-piezometers offer an inexpensive and simple method for obtaining groundwater level and pressure data

553   (Lee and Cherry, 1978). They are easy and quick to install in most locations, and the analysis of their

554   measurements is generally straightforward (Brodie et al., 2007). They can be used in small-scale applications

555   and in detailed surveys in heterogeneous environments (Fritz et al., 2016). However, measurements at a study

556   site must be taken at the same time to be representative of similar flow conditions (Kalbus et al., 2006). Another

557   important factor to consider is that many data loggers require a certain diameter well. In previous studies,

558   groundwater level observations have rarely been used in isolation and typically have been coupled with other

559   methods.

560

561   The heterogeneous composition of braided rivers complicates the estimation of the hydraulic conductivity of

562   streambeds on a reach or catchment scale. Hydraulic conductivity can vary significantly across an area, even

563   with small changes in sediment composition, thus it is difficult to extrapolate values to represent a large area

564   (Brodie et al., 2007). With grain size analysis, the structure and stratification of the sediment are destroyed

565   during analysis, so the conductivity value does not represent the vertical or horizontal conductivity (Cheng and

566   Chen, 2007) and does not provide any information on preferential pathways (Brunner et al., 2017). This issue is

567   particularly problematic in gravel-bed braided rivers where there is high anisotropy and a large portion of sub-

568   surface flow occurs in preferential channels (Dann et al., 2008). Similarly, when conducting permeameter tests it

569   is difficult to transport sediment samples without disturbing their structure (Kalbus et al., 2006). In particular,

570   taking undisturbed cores of sediments containing unconsolidated gravel, as most braided rivers do, is nearly

571   impossible. However, these tests can be used as a preliminary estimation before conducting further tests. Also,

572   there is the potential for freeze coring, which allows for an intact sediment sample, but there are limitations,

573   such as in rivers with warm water or compacted cobbles (Brunner et al., 2017). Slug tests are quick and simple

574   to carry out and a significant advantage is that they only require one well. Pumping tests on the other hand

575   require a pumping well and an observation well, which can be cost prohibitive. Pumping test results provide

576   average hydraulic conductivity values across a larger area than for slug tests, thus their results may be less

577   sensitive to heterogeneous conditions (Kalbus et al., 2006), whereas slug tests provide information only about

578   the location where the well is installed. Arguably, as vertical hydraulic conductivity is the controlling factor for

579   river losses, there should be more focus on estimating anisotropy values of the braided river substrate. Methods

580 for estimating anisotropy have been demonstrated using aquifer tests (Neuman et al., 1984; Mutch, 2005;

581 Mathias and Butler, 2007) and more recently geophysics (Al-Hazaimay et al., 2016; Fernández-Álvarez et al.,

582 2016).

583

584 **2.5   Modelling**

585 Computer modelling is often used for the estimation of exchange between surface water and groundwater as a

586 complement to field measurements. Such computer models have become irreplaceable tools to gain insight into

587 real-world surface water-groundwater issues ranging from system understanding at the local or regional scale to

588 future projections for management purposes. The complexity of numerical hydrological models used for this

589 purpose range from simple conceptual models that treat subsurface compartments (i.e., groundwater) as

590 reservoirs where inflows or outflows are specified, to highly complex integrated models that have a more

591 realistic physical coupling between surface water and groundwater. MODFLOW (Harbaugh, 2005) is the most

592 commonly used numerical model to simulate surface water-groundwater interactions (Furman, 2008; Barlow

593 and Harbaugh, 2006). As pointed out by Wöhling et al. (2018), MODFLOW is considered to be a good

594 compromise between integrated and conceptual modelling approaches. Several packages are available in

595 MODFLOW for simulating surface water-groundwater interaction and further details about the application and

596 limitations of these can be found in Brunner et al. (2009) and (2010).

597

598 While the modelling of braided rivers is not new, it has been done more often from a geomorphological

599 perspective (e.g., Ashmore, 1993; Copley and Moore, 1993; Meunier et al., 2006; Williams et al., 2016).

600 Nevertheless, a number of published studies detail modelling of braided rivers for the purposes of understanding

601 groundwater-surface water interactions (e.g., Baalousha, 2012; Chen, 2007; Passadore et al., 2015; Scott and

602 Thorley, 2009; Shu and Chen, 2002; Wilson and Wohling, 2015; Wohling et al., 2018).

603

604 Wilson and Wöhling (2015) attempted to improve the understanding of Wairau River recharge into the Wairau

605 aquifer in Marlborough, New Zealand, using a steady-state MODFLOW model and the SFR2 package. The

606 authors noted groundwater monitoring records and pump testing showed the aquifer to be more complex and

607 stratified than previously thought, indicating that groundwater monitoring sites were likely only representative

608 of local conditions. This finding underscores the difficulties of modelling highly heterogeneous, complex river

609 systems and their associated aquifers. This was further highlighted by Close et al. (2016) who used the Wilson

610   and Wöhling (2015) MODFLOW model as a basis for a study using heat as a tracer in the Wairau aquifer. Close

611   et al. (2016) found that including heterogeneity was important when calibrating the model to observed

612   temperature data.

613

614   In a subsequent study of the Wairau Plain aquifer and the Wairau River, Wöhling et al. (2018) developed a

615   transient MODFLOW model that was calibrated using targeted field observations as well as "soft" information

616   from experts of the local water authority. The uncertainty of simulated river-aquifer exchange flows was

617   evaluated using Null Space Monte Carlo methods. The study suggested that the river is hydraulically perched

618   (losing) above the regional water table in its upper reaches and is gaining in the downstream section. It was

619   found that despite large river discharge rates (i.e., regularly reaching 1000 m^3/s), the net exchange of flow rarely

620   exceeded 12 m^3/s and seemed to be limited by the physical constraints of unit-gradient flux under disconnected

621   rivers. An important finding for the management of the aquifer was that changes in aquifer storage are mainly

622   affected by the frequency and duration of low-flow periods in the river.

623

624   **Advantages and Limitations**

625   Field methods are often time consuming and expensive, and they may not be at the targeted spatial or temporal

626   scale. Therefore, the estimation of exchange between braided rivers and groundwater is often complemented by

627   hydrological modelling. It is also possible to integrate a range of data types at varying spatial and temporal

628   scales with modelling. MODFLOW is commonly used to model surface water-groundwater interaction,

629   including in braided rivers. Complex flow channel geometry, which changes over time, is not explicitly

630   incorporated into modelling efforts, at least in the studies identified by the authors listed above. As such, the

631   impact of complex and temporally variable flow channel geometry on surface water-groundwater exchange is

632   not well understood. More complex integrated modelling approaches than that possible using the MODFLOW

633   suite of packages is likely required to incorporate this level of detail. A future integrated approach that considers

634   channel geometry in a more physically realistic manner may be facilitated by the recent development of braided

635   river terrain models (e.g., Williams et al., 2016) and methods for simulating the heterogeneity of braided river

636   sediments (e.g., Ramanathan et al., 2010).

637

638   **3   Discussion**

639   There are many factors to consider when selecting the appropriate method(s) for studying groundwater-surface

640   water interactions, and there are special considerations relevant to braided river environments. The most

641   appropriate method will depend on physical and hydrological conditions in the setting and scale of interaction to

642   be measured (LaBaugh and Rosenberry, 2008). As a result of this review of studies investigating groundwater-

643   surface water exchange in braided rivers, a summary table has been developed (Table 1) that summarises the

644   literature discussed in this paper and the advantages and disadvantages of the various methods used in these

645   studies.

646   **Table 1.** Advantages and disadvantages of various methodologies for measuring groundwater-surface water interactions in braided rivers

Method	Advantages	Disadvantages	Applications of these methods in braided rivers*
Water budgets	- Better suited for relatively homogeneous aquifers - Good for large-scale studies - Useful for identifying hotspots of river gains and losses - Can be simple and relatively quick to calculate	- Errors can be greater than the amount of groundwater-surface water flux - Not well suited for sub-reach scale - Not very accurate in highly heterogeneous systems - Does not consider streambed throughflow - Multiple sites on a river must be gauged concurrently - Errors can be significant in large catchments - Uncertainties of land surface recharge and offshore flow rates can result in errors - Can be expensive and time consuming depending on how data is collected	Acuña & Tockner (2009); Burbery & Ritson (2010); Doering et al. (2013); Farrow (2016); Larned et al. (2008); Larned et al. (2015); Riegler (2012); Simonds & Sinclair (2002); Soulsby et al. (2004); White et al. (2012)
Hydrochemistry	- Good for environments where there is a sufficient difference between tracer concentrations in groundwater and surface water - Useful for identifying interactions on a large scale - Some parameters can easily be included in long-term, routine sampling - Some tracers can be used to quantify seepage rates	- Analysis errors can be an issue when concentrations are low (e.g., radon) - Groundwater and surface water concentrations may be too close to differentiate - Concentrations may not be temporally or spatially consistent - Some tracers (e.g., dissolved oxygen, nitrate) may be affected by biogeochemical processes, so they need to be conservative on the scale of the investigation	Acuña & Tockner (2009); Blackstock (2011); Botting (2010); Burbery & Ritson (2010); Cantafio & Ryan (2014); Close (2014); Close et al. (2014); Coluccio (2018); Doering et al. (2013) Domisse (2006); Guggenmos (2011); Larned et al. (2015); Malard et al. (2001); Moore (1997); Rodgers et al. (2004); Soulsby et al. (2004); Vincent (2005)
Temperature studies	- Variety of methods ranging in complexity, cost, scale - Can be used for both locating areas of discharge/recharge and quantifying flux - Aerial surveys can be faster than in-stream surveys	- Often needs to be measured continuously - Need a sufficient temperature difference between groundwater and surface water - May be less effective in periods of high river flows	Acuña & Tockner (2009); Close (2014); Close et al. (2016); Coluccio (2018); Doering et al. (2013); Lovett et al. (2015); Malard et al. (2001); Passadore et al. (2015)

Modelling	• Acts as a database for field data • Can assist researchers to develop intuition about physical processes and refine their conceptual models • Useful for carrying out regional-scale assessments for management purposes, such as determining streamflow depletion associated with pumping • MODFLOW packages widely accepted for numerical simulation and intuitive to apply • MODFLOW packages considered a good compromise between a simple conceptual modelling approach and a more complex integrated approach	• Some models have high computational and time requirements • Various assumptions required that may not reflect actual hydraulic processes or aquifer properties	Baalousha (2012); Chen (2007); Close et al. (2016); Passadore et al. (2015); Scott & Thorley (2009); Shu & Chen (2002); Wilson & Wöhling (2015); Wöhling et al. (2018)
Darcy approach	• Piezometers are typically easy and quick to install • Wells can be installed in-stream or on land • Can also use existing well networks • Can be used in small-scale or regional applications • Can be used to survey heterogeneous areas • Piezometer measurements are straightforward to analyse	• Deep groundwater wells are expensive to install • All measurements at a study site must be taken at the same time • Hydraulic conductivity can significantly vary spatially, thus difficult to extrapolate to represent a large area	Acuña & Tockner (2009); Botting (2010); Burbery & Ritson (2010); Chen (2007); Cheng et al. (2010); Coluccio (2018); Doering et al. (2013); Domisse (2006); Larned et al. (2008); Larned et al. (2015); Malard et al. (2001); Passadore et al. (2015); Riegler (2012); Shu & Chen (2002); Simonds & Sinclair (2002); Van't Woudt & Nicolle (1978); Vincent (2005); Wilson & Wöhling (2015); Wöhling et al. (2018)

647   *Note: some studies referenced in this table were not discussed in the text.

648

649 The objectives of a study will influence which methods are most applicable. If only qualitative information

650 about groundwater-surface water exchange is required, this could be obtained by methods such as mapping the

651 locations of wet and dry reaches of a river, or identifying where there is mixing between groundwater and

652 surface water based on chemical or heat tracers. Qualitative data will often assist in developing a conceptual

653 understanding of the study site, which is a critical first step in data gathering. Alternatively, if quantitative data

654 is needed, such as the rate of groundwater seepage into a surface water body, this may be obtained by measuring

655 Rn-222, analysing temperature signals, or by calculating the hydraulic gradient. Researchers have developed

656 flux quantification techniques for some of the methods discussed in this paper (e.g., for temperature analysis see

657 Gordon et al., 2012), but it is important to consider inputs required to calculate seepage through a streambed,

658 such as streambed hydraulic conductivity (see section 2.4). If direct water samples are needed, tools to consider

659 could include groundwater wells or mini-piezometers. Water samples and flux rates can also be obtained using

660 seepage meters, a common method used for estimating groundwater-surface water interactions typically based

661 on the design proposed by Lee (1977). However, it does not appear that these devices have been previously used

662 in gravel-bed braided rivers. Seepage meters have various limitations as discussed in previous studies (e.g.,

663 Kelly and Murdoch, 2003; Brodie et al., 2009; Cey et al., 1998), which indicate their application in braided

664 rivers would be difficult and less effective than other methods.

665

666 It is important to match the scale of the data required with the methods being used. This should include the

667 consideration of both spatial and temporal scales. Remote sensing techniques such as airborne thermal infrared

668 imaging and geophysics may prove useful to apply in braided river settings for gathering data on a large scale,

669 as these methods have been used in braided rivers for geomorphological studies (e.g., Huber and Huggenberger,

670 2016) and for investigating groundwater-surface water exchange in other settings (McLachlan et al., 2017). We

671 discuss these approaches further in Section 4. It is important to recognise that it may be difficult to accurately

672 characterise smaller scale groundwater-surface water interactions in highly heterogeneous braided river

673 environments based on broad-scale methods. However, obtaining a broad snapshot of conditions or processes in

674 a location may provide sufficient information to satisfy the study's objectives. Also, using a combination of

675 broad and point-scale techniques at a single study site may help overcome the limitations of the individual

676 techniques, particularly in heterogeneous environments (Kalbus et al., 2006).

677

678   Site-specific characteristics will largely determine the most appropriate methods to use relating to the geology,

679   topography, hydrochemistry, hydrology and hydrogeology of the study site. Large braided rivers with high

680   flows and deep channels may prove difficult to access directly. There is also a reasonable risk of the loss or

681   damage of equipment installed in braided riverbeds due to floodwaters and sediment movement. These practical

682   considerations underline the potential benefits of remote techniques to collect data in this type of river.

683

684   **4   Key gaps and possibilities**

685   This paper has highlighted that there are currently gaps in the knowledge of how groundwater and surface water

686   interact in braided rivers. There is limited conceptual understanding of hyporheic flow processes, and how they

687   impact river flow levels and water quality in braided rivers. The hyporheic zone has been highlighted as a

688   significant area for ecological processes in rivers (Febria et al., 2011; Krause et al., 2011; Malard et al., 2001),

689   but as Kalbus et al. (2006) note, it can be difficult to differentiate between hyporheic exchange and groundwater

690   discharge. In addition, despite the contributions of the studies discussed here, the recharge rates to and from

691   braided rivers continue to be a source of question for water scientists and managers, as this has implications for

692   both water quality and quantity. Measuring seepage rates is still difficult in many gravel-bed braided rivers, and

693   often there is significant uncertainty in the data collected. Lastly, there is still much scope for research on

694   identifying historical patterns of dry and low-flow periods in braided river reaches. This is often an area of

695   significant concern for communities that are seeking answers on the correlations between dry or low-flow

696   periods, and current and historical water use practices and climate.

697

698   There is also room for improvement in the methods available to carry out these investigations. Refinement of

699   techniques that allow for direct measurements of physical or chemical properties in braided rivers would be

700   helpful. While the studies presented here have employed some direct methods, there is still a need for techniques

701   that can be used in braided rivers with coarse gravel substrate, wide active riverbeds, and shifting channels and

702   gravel bars. Methods that can better capture the heterogeneous properties of braided rivers would be ideal. The

703   present paper has shown the promise of using environmental tracers such as Rn-222 and stable isotopes, as well

704   as heat tracers in these settings. Additional techniques that allow for indirect measurements would also be

705   beneficial, given the difficulty of working directly in braided rivers. Geophysical methods (discussed in more

706   detail below) have been used in many other river environments to gather information about hydrogeologic

707   systems that can then be inferred to better conceptually understand groundwater-surface water exchange. There

is also scope for more remote collection of data, and Carbonneau and Piégay (2012) review a range of technique

for use in rivers, while Marcus (2012) provides an overview of remote sensing specifically in gravel-bed rivers.

There is a significant amount of freely available satellite data (e.g., via the Sentinel satellites,

https://sentinel.esa.int/web/sentinel/home) that may be useful in braided river studies. Unmanned aerial vehicles

have become more affordable and advanced in recent years, allowing for remote collection of a range of data on

rivers such as thermal infrared, multispectral and hyperspectral imaging, and photogrammetry (Pai et al., 2017).

Artificial dye, chemical (e.g., salt) or bacterial tracers are often useful for shedding light on processes such as

groundwater velocity and flow paths or hyporheic zone flow (Flury and Wai, 2003). They have been used in

other types of rivers to investigate groundwater-surface water exchange (e.g., Binley et al., 2013; Ferreira et al.,

2018; Stoner et al., 2013; Knöll and Scheytt, 2018; González-Pinzón et al., 2015). Several studies have used

rhodamine dye in a New Zealand well array installed in an alluvial aquifer deposited by braided rivers to

estimate hydraulic properties and examine groundwater flow paths (e.g., Close et al., 2002; Dann et al., 2008;

Sarris et al., 2018). For artificial tracer tests to be time and cost effective, some prior knowledge of water flow

paths and velocities is necessary (Close et al., 2002).

There is scope to use other temperature methods than those described in section 2.3, such as fibre-optic

distributed temperature sensing (FODTS) (Busato et al., 2019; Klinkenberg, 2015; Lovett et al., 2015; Meijer,

2015; Rosenberry et al., 2016; Mwakanyamale et al., 2012) or active heat pulse methods (see Briggs et al.,

2016; Banks et al., 2018). Collection of temperature profiles was briefly mentioned in section 2.3 in the study

conducted by Coluccio (2018), which used 1-D temperature profiles. However, there are several ways

temperature profiles can be collected (1-D, 2-D, 3-D), as well as a range of analysis methods that can be used, as

demonstrated in several previous studies in non-braided river settings (Briggs et al., 2014; Gordon et al., 2013;

Naranjo and Turcotte, 2015; Rosenberry et al., 2016). There is also considerable scope for applying thermal

infrared (TIR) imaging in braided rivers. Handcock et al. (2012) provide a comprehensive review of the use of

TIR imaging in rivers. Using TIR imaging to highlight temperature differences in a braided streambed may be

particularly useful for qualitatively identifying locations of groundwater inflow to rivers. TIR data can be

collected remotely (by UAV, helicopter or fixed wing plane), on the ground or by satellite, and there are

important considerations with each category (e.g., cost, scale of data collected). TIR imaging has been used in

several river environments to identify groundwater-surface water interactions (Culbertson et al., 2013; Eschbach

738   et al., 2017; Hare et al., 2015; Liu et al., 2016; Lovett et al., 2015; Rautio et al., 2018) but does not appear to

739   have been applied in braided rivers to any great extent for this purpose.

740

741   There have been many advances in geophysical techniques in recent years, and these methods do not appear to

742   have been applied in braided river settings for investigations of groundwater-surface water exchange.

743   McLachlan et al. (2017) provide a thorough recent review of geophysical methods for characterising the

744   groundwater-surface water interface such as electrical resistivity tomography (ERT); ground penetrating radar

745   (GPR); seismic methods; and forward and inverse geophysical modelling. These methods allow for river

746   systems to be characterised where factors such as geological, hydrological and biogeochemical heterogeneity

747   make it difficult to make direct measurements (McLachlan et al., 2017). Geophysical methods may be

748   particularly useful for characterising subsurface structures in braided rivers, which are poorly understood. A

749   recent study by Busato et al. (2019), demonstrates the use of ERT and FODTS in a rocky stream with poorly

750   sorted substrate, which may provide useful learnings for braided rivers. Examples of studies in other types of

751   river environments that used geophysics to characterise the groundwater-surface water interface include Singha

752   et al. (2008), Binley et al. (2013) and Steelman et al. (2017). Geophysical data can also be collected remotely in

753   airborne electromagnetic surveys such as in Harrington et al. (2014). As McLachlan et al. (2017) note,

754   geophysical techniques should be used to complement data collected by other hydrological and biogeochemical

755   methods.

756

757   As discussed in the modelling section of this paper, there is also opportunity for new approaches to modelling of

758   braided rivers. Brunner et al. (2017) note that there have been recent advances in hydrologic modelling that

759   incorporate both surface and sub-surface water flow, and there is certainly room to apply some of these

760   techniques to braided river settings. Given the dynamic nature of braided river subsurface hydrology, models

761   needed to fully couple surface and subsurface flow. There are software packages that have been applied

762   elsewhere such as HydroGeoSphere (e.g., Gilfedder et al., 2019; Goderniaux et al., 2009; Tang et al., 2017) and

763   MIKE-SHE (e.g., Butts et al., n.d.; House et al., 2016; Bandini et al., 2017) that appear promising to try in

764   addition to MODFLOW, which has been traditionally used in braided river modelling of the groundwater-

765   surface water interface. As the braided river bedform cannot be sufficiently characterised using the existing

766   MODFLOW SFR functions, there is a need to address this gap. Also, current groundwater models do not allow

767   for changes to bed morphology, which is a key feature of braided rivers, but using script-based models (e.g.,

768   FloPy (Bakker et al., 2016)), may allow for this to be achieved.

769

770   **5   Summary**

771   Braided rivers are unique and dynamic river environments that serve important ecological, cultural, recreational

772   and freshwater resource functions. A critical aspect of their effective management is understanding groundwater

773   and surface water interactions in these rivers and their associated aquifers. This article provides an overview of

774   characteristics specific to braided rivers, which include multiple meandering channels that often shift; temporary

775   and semi-permanent bars and islands; wide active riverbed areas; heterogeneous and (typically) mixed sand and

776   gravel streambeds; and a significant portion of river flow that occurs within the streambed. We present a map

777   showing the regions where braided rivers are mainly found at the global scale: Alaska, Canada, the Japanese and

778   European Alps, the Himalayas, Russia and New Zealand. To the authors' knowledge, this is the first map of its

779   kind. Our review of prior studies of surface water-groundwater interactions in braided rivers showed that most

780   studies have been recent (in the past 10-20 years), and they have investigated a range of questions including

781   calculating seepage rates to/from braided rivers; estimating time lags between rivers and groundwater; and

782   looking at the implications of groundwater-surface water exchange on ecological processes. We also

783   investigated the effectiveness of the various methods used in the studies identified in this review in terms of

784   achieving the studies' objectives and their applicability in braided rivers. A table has been produced

785   summarising these findings and shows that there is a variety of available methods ranging in cost and scale.

786

787   Lastly, this article explored the various considerations one may make when choosing appropriate techniques for

788   investigating groundwater-surface water exchange in braided rivers. The use of multiple methods at varying

789   spatial scales at a single study site may help overcome the uncertainties associated with data gathered in

790   heterogeneous, dynamic braided river environments. Given the challenges of working directly in braided rivers,

791   there is considerable scope for the increased use of remote sensing techniques and geophysics. There is also

792   scope for new approaches to modelling braided rivers using integrated techniques that incorporate the often-

793   complex river bed terrain and geomorphology of braided rivers explicitly. There is presently limited

794   understanding of how hyporheic zone processes operate and impact braided rivers; recharge rates to and from

795   braided rivers; and historic drying and low-flow trends in braided rivers, and thus future research is needed in

796   these areas. While only some of the methods discussed here allow for quantification of groundwater-surface

797   water flux, many of these techniques can improve our conceptual understanding of these systems and processes.

798

799   Author contribution

800   The project was instigated by LM. KC carried out the literature review that formed the content of this

801   manuscript and wrote the initial manuscript draft. KC and LM revised the manuscript together.

802

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808

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