

Electrical Resistivity Dynamics beneath a Fractured Sedimentary Bedrock Riverbed in Response to Temperature and Groundwater/Surface Water Exchange

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1 **Abstract.** Bedrock rivers occur where surface water flows along an exposed rock surface. Fractured sedimentary
2 bedrock can exhibit variable groundwater residence times, anisotropic flow paths, heterogeneity, along with
3 diffusive exchange between fractures and rock matrix. These properties of the rock will affect thermal transients in
4 the riverbed and groundwater-surface water exchange. In this study, surface electrical methods were used as a non-
5 invasive technique to assess the scale and temporal variability of riverbed temperature and groundwater-surface
6 water interaction beneath a sedimentary bedrock riverbed. Conditions were monitored on a semi-daily to semi-
7 weekly interval over a full annual period that included a seasonal freeze-thaw cycle. Surface electromagnetic
8 induction (EMI) and electrical resistivity tomography (ERT) methods captured conditions beneath the riverbed
9 along a pool-riffle sequence of the Eramosa River in Canada. Geophysical datasets were accompanied by
10 continuous measurements of aqueous specific conductance, temperature and river stage. Time-lapse vertical
11 temperature trolling within a lined borehole adjacent to the river revealed active groundwater flow zones along
12 fracture networks within the upper 10 m of rock. EMI measurements collected during cooler high-flow and warmer
13 low-flow periods identified a spatiotemporal riverbed response that was largely dependent upon riverbed
14 morphology and seasonal groundwater temperature. Time-lapse ERT profiles across the pool and riffle sequence
15 identified seasonal transients within the upper 2 m and 3 m of rock, respectively, with spatial variations controlled
16 by riverbed morphology (pool versus riffle) and dominant surficial rock properties (competent versus weathered
17 rock rubble surface). While the pool and riffle both exhibited a dynamic resistivity through seasonal cooling and
18 warming cycles, conditions beneath the pool were more variable largely due to the formation of river ice during the
19 winter season. We show that surface electrical resistivity methods have the capacity to detect and resolve electrical
20 resistivity transience beneath a fractured bedrock riverbed in response to porewater temperature and specific
21 conductance fluctuations over a complete annual cycle.

22

23 **1 Introduction**

24 Fractured sedimentary bedrock represents an important source of water for many communities around the world.
25 Although the effective porosity of rock is low relative to unconsolidated sediment, the existence of dense networks
26 of interconnected fractures, dissolution-enhanced conduits, and karst features, can result in productive yet
27 heterogeneous and anisotropic flow systems. An exposed bedrock surface may exhibit greater variability in flow
28 and transport properties as it is subjected to weathering and erosional processes, leading to very complicated
29 groundwater recharge and discharge patterns. Fractured sedimentary rock is best conceptualized as a dual porosity
30 system where fractures dominate flow, but remain connected to water stored in the porous matrix through advection
31 and diffusion. Such conceptualizations of fracture flow and transport are routinely applied to groundwater resource
32 (e.g., Novakowski and Lapcevic, 1998; Lemieux et al., 2009; Perrin et al., 2011) and contaminant transport studies
33 (e.g., Zanini et al., 2000; Meyer et al., 2008; McLaren et al., 2012). Recent studies have extended these concepts to
34 fluvial river environments (e.g., Singha et al., 2008; Toran et al., 2013a).

35 Groundwater-surface water interactions at the reach-scale are conceptualized through gaining, losing and flow-
36 through interactions (Woessner, 2000). At the channel scale, micro-to-macro bedform variations result in variably-
37 scaled zones of surface water downwelling (recharge) and groundwater upwelling (discharge) (e.g., Binley et al.,
38 2013; Käser et al., 2013). Groundwater temperature measurements are frequently used to monitor spatiotemporal
39 variations in groundwater-surface water exchange or flux across riverbeds (e.g., Anderson, 2005; Irvine et al., 2016).
40 Yet, very little is known about the existence and nature of hyporheic and groundwater-surface water mixing zones in
41 fractured sedimentary bedrock, largely because these systems are very difficult to instrument using direct methods
42 (e.g., drive point monitoring wells, seepage meters, thermistors), and the scale of the interaction may be very small
43 or heterogeneous relative to unconsolidated sediment.

44 Hydrologic processes along a fractured bedrock river were explored by Oxtobee and Novakowski (2002), who
45 concluded that groundwater-surface water interaction was restricted by poor vertical connectivity and limited
46 exposure of horizontal bedding plane fractures. A subsequent numerical sensitivity analysis by Oxtobee and
47 Novakowski (2003) confirmed that groundwater-surface water connectivity through discrete fractures would be
48 highly variable in space and time, and would largely depend on fracture size or aperture, river stage, and the
49 distribution of hydraulic head within the flow system. Fan et al. (2007) numerically explored the influence of
50 larger-scale fracture orientations and geometries on the groundwater flow system near a stream; they concluded that
51 the base flow to a stream would be higher for streams aligned with fracture dip than those aligned with fracture
52 strike. Therefore, groundwater-surface water interaction in a fractured bedrock environment will depend on stream-
53 fracture alignment. Although these previous studies offered valuable insights into the magnitude of groundwater-
54 surface water exchange, they were based on idealized fracture network conceptualizations, and did not consider the
55 role of matrix porosity and potential exchanges between fractures and the porous matrix.

56 Electrical and electromagnetic methods such as ground-penetrating radar, electromagnetic induction and electrical
57 resistivity imaging are commonly used to characterize fluvial deposits (e.g., Naegeli et al., 1996; Gourry et al., 2003;
58 Froese et al., 2005; Sambuelli et al., 2007; Rucker et al., 2011; Orlando, 2013; Doro et al., 2013; Crosbie et al.,

59 2014). The capacity of time-lapse electrical resistivity imaging for conceptualization of groundwater transients in
60 sediment is well-documented in the literature (e.g., Nyquist et al., 2008; Miller et al., 2008; Coscia et al., 2011;
61 Cardenas and Markowski, 2011; Musgrave and Binley, 2011; Coscia et al., 2012; Dimova et al., 2012; Wallin et al.,
62 2013). Electrical imaging of natural river systems perturbed by solute tracers has resulted in unprecedented
63 visualizations of fluid flow (e.g., Ward et al., 2010a, 2010b; Doetsch et al., 2012; Ward et al., 2012; Toran et al.,
64 2013a; Toran et al., 2013b; Harrington et al., 2014). More recent applications of electrical resistivity in karst
65 undergoing surface water transients have shown how surface geophysics can unravel complex hydrologic processes
66 in sedimentary bedrock environments (e.g., Meyerhoff et al., 2012; Meyerhoff et al., 2014; Sirieix et al., 2014),
67 especially when site conditions limit the use of more invasive direct measurement methods.

68 While a variety of geophysical tools and techniques can measure flow and water chemistry in space and time
69 (Singha et al., 2015), the most appropriate tool and approach will depend on the scale of interest. The vast majority
70 of geophysical work within shallow river environments has utilized discrete temperature monitoring below the
71 riverbed to detect vertical fluxes (e.g., White et al., 1987; Silliman and Booth, 1993; Evans et al., 1995; Alexander
72 and Caissie, 2003; Conant, 2004; Anderson, 2005; Hatch et al., 2006; Keery et al., 2007; Schmidt et al. 2007;
73 Constantz, 2008). Recent advancements in distributed fiber optic cables have improved spatial and temporal
74 resolution of groundwater-surface water interactions (e.g., Slater et al., 2010; Briggs et al., 2012; Johnson et al.,
75 2012).

76 Groundwater and surface water interaction can be monitored through changes in thermal gradient or electrolytic
77 concentration (e.g., Norman and Cardenas, 2014), yet the scale and magnitude of these interactions will vary as a
78 function of riverbed architecture and subsurface hydraulic conditions (Crook et al., 2008; Boano et al., 2008; Ward
79 et al., 2012; Tinkler and Wohl, 1998) resulting in spatially dynamic exchange. These processes are further
80 complicated by diel (e.g., Swanson and Cardenas, 2010) and seasonal (e.g., Musgrave and Binley, 2011)
81 temperature fluctuations across a range of spatial scales, local transients such as precipitation events (e.g.,
82 Meyerhoff et al., 2012), river stage fluctuations (e.g., Bianchin et al., 2011) and controlled dam releases (e.g.,
83 Cardenas and Markowski, 2011). Relative to other non-invasive geophysical methods, electrical resistivity methods
84 are more robust in their ability to provide information about temperature and solute fluctuations beneath actively
85 flowing surface water bodies (e.g., Nyquist et al., 2008; Cardenas and Markowski, 2011; Ward et al., 2012;
86 Meyerhoff et al., 2014) particularly in a time-lapse manner. Unlike conventional hydrogeological methods (e.g.,
87 screened or open coreholes), which may bias conduction in the fractures, surface electrical methods are sensitive to
88 the bulk electrical conductivity of the formation, making them more suited for detection of processes between the
89 open fractures/conduits and the porous matrix.

90 A hypothetical groundwater-surface water mixing zone in a porous fractured rock system (Fig. 1) will exhibit
91 transience in water temperature and specific conductance as a result of changes in groundwater flow (recharge vs.
92 discharge), mixing of surface water and groundwater, seasonal atmospheric temperature fluctuations and/or
93 biogeochemical changes within the riverbed. Ward et al. (2010a) demonstrated how surface electrical methods
94 sensitive to changes in water conductivity could be used to detect and quantify diffusive mass transport (exchange)

95 between a mobile and immobile storage zone in a shallow riverbed. Our study uses a similar geophysical
96 monitoring approach to assess the magnitude and scale of groundwater-surface water transience beneath a fractured
97 bedrock riverbed based on the detection and characterization of geoelectrical dynamics over a complete annual
98 cycle. Seasonal freezing and thawing of surface water bodies is an important process in mid-latitude climates;
99 winter freeze-up will reduce base flow contributions, while spring snow melt will result in a sudden and large
100 increase in base flow to a river. Therefore, understanding the transient behaviours of the hydrogeologic system,
101 including water-phase transformations, over a complete annual cycle will be critical to our understanding of the
102 magnitude and spatial extent of groundwater-surface water exchange along fractured bedrock rivers. Given the
103 challenges and costs associated with installing sensors within rock and heterogeneous flow system characteristics,
104 minimally invasive surface and borehole geophysical methods offer an ideal alternative, and possibly, more
105 effective approach for long-term groundwater-surface water monitoring of bedrock environments.

106 Our study examines the capacity of electrical imaging methods (i.e., electromagnetic induction and electrical
107 resistivity) to monitor geoelectrical transients within a fractured sedimentary bedrock river to better understand
108 groundwater-surface water transience over a complete annual cycle. To achieve this, seasonal variations in
109 electrical resistivity distribution were measured across a 200 m reach of a bedrock river using a ground conductivity
110 meter and time-lapse electrical resistivity measurements along two fixed transects intersecting a pool-riffle
111 sequence. These geophysical surveys were supported by continuous measurements of groundwater and surface
112 water temperature, specific conductance and river stage. Our study shows that spatiotemporal resistivity dynamics
113 were largely controlled by riverbed morphology in combination with seasonal changes in water temperature, and to
114 a lesser-degree electrolytic concentration. The formation of ground frost and basal ice strongly affected the
115 electrical resistivity beneath the riverbed compared to intraseasonal dynamics (spring, summer and fall). Observed
116 geoelectrical changes beneath the riverbed appear primarily dependant on seasonal temperature trends exhibiting
117 varying zones of influence (vertical and horizontal) across the pool and riffle section of the river. The riverbed was
118 strongly susceptible to seasonal atmospheric temperature fluctuations, which might have implications to
119 biogeochemical processes or benthic activity.

120 **2 Background**

121 The Eramosa River – a major tributary of the Speed River within the Grand River Watershed in Ontario, Canada –
122 resides upon a regional bedrock aquifer of densely fractured dolostone with dissolution-enhanced conduits and karst
123 features (e.g., Kunert et al., 1998; Kunert and Coniglio, 2002; Cole et al., 2009). Although this aquifer represents
124 the sole source of drinking water for the region, the potential effects of increased groundwater pumping on the
125 overlying bedrock river and surrounding ecosystems, are not yet understood. This is largely due to a gap in our
126 conceptual understanding of groundwater-surface water interaction in rivers that flow directly along sedimentary
127 bedrock surfaces with exposed fracture networks. The fractured sedimentary bedrock exhibits a complex flow
128 system due to variably connected fracture networks, dissolution-enhanced features, and variable bedrock exposure
129 (Steelman et al. 2015a, 2015b). The bulk of the flow will occur along the fracture networks, with highly-variable

130 head distributions; matrix storage could support equally complex biogeochemical processes and thermal dynamics
131 through convective or diffusive exchange with open fractures or dissolution-enhanced features.

132 A focused geophysical investigation was carried out along a 200 m reach of the Eramosa River (Fig. 2). The study
133 area was positioned at a bend in the river with relatively cleared vegetation along the south shoreline and adjacent
134 floodplain with exposed rock at surface. A network of coreholes (continuously cored boreholes) and streambed
135 piezometers were installed across the site. Locally, the water table elevation corresponds to the surface water or
136 stage elevation, resulting in vadose zone thicknesses from <0.5 m to 2.0 m along the shorelines. The temperate
137 southern Ontario climate subjects the river to a wide-range of seasonal conditions, including high precipitation
138 periods in spring and fall, hot and dry summers, and variable degrees of ground frost and surface water freeze-up
139 during the winter months (Fig. 3a).

140 Locally, the river incises the Eramosa Formation by 2 m to 3 m exposing abundant vertical and horizontal fractures
141 with little to no alluvial sediment deposited along the riverbed (Fig. 3b). Regionally the Eramosa acts as a
142 discontinuous aquitard unit (Cole et al. 2009); however, core logs collected at the study site show bedding plane and
143 vertical joint set fractures spanning the entire 11 m sequence of Eramosa. This upper formation is underlain by
144 approximately 3 m of cherty, marble-like Goat Island formation, which exhibits high-angle fractures along cherty
145 nodules near the Eramosa contact. The Goat Island is underlain by more than 15 m of Gasport formation, which
146 exhibits coral reef mounds of variable morphology. The rock matrix of the Gasport is visually more porous with
147 well-defined vugs, dissolution-enhanced features, and fewer fractures than the overlying Goat Island and Eramosa.
148 A full-description of these bedrock sequences can be found in Brunton (2009).

149 In this region, the winter season may be accompanied by the formation of ground frost and surface water freezing.
150 Seasonal freeze-up will consist of an ice crust layer on the surface of the water and the possible formation of basal
151 ice along the riverbed (Stickler and Alfredsen, 2009). The latter phenomenon can occur during extreme atmospheric
152 cooling over turbulent water bodies, resulting in super-cooled water (<0°C) that rapidly crystalizes to form frazil
153 (i.e., tiny ice particles with adhesive characteristics); these crystals can flocculate to form slush, which adheres and
154 accumulates on the substrate forming a basal ice layer.

155 **3 Methods**

156 **3.1 Electrical Properties**

157 Electrical resistivity methods are based upon Ohm's Law ($R = \Delta V / I$). In the case of a homogeneous half-space,
158 the electrical resistance (R) of the subsurface is determined by measuring the potential difference (ΔV) across a pair
159 of 'potential' electrodes due to an applied current (I) across a pair of 'current' electrodes some distance away. The
160 measured R (Ω) across a unit volume of the earth can be converted to apparent resistivity (Ω m) using a specific
161 geometric factor that compensates for varying electrode array geometry (Reynolds, 2012). Apparent resistivity
162 measurements are commonly interpreted using tomographic inversion techniques, whereby measured data is
163 reconstructed from forward models of an optimized physical parameter distribution (Snieder and Trampert, 1999;
164 Loke et al., 2013). Although data inversion techniques are standard practice in the interpretation of most

165 geophysical data, the model that best matches the measured data is not necessarily an exact representation of the
166 subsurface. Here, the inversion process ultimately yields a smoothed representation of the true parameter
167 distribution.

168 The bulk electrical resistivity (i.e., inverse of conductivity) of a fluid-saturated medium can be calculated through a
169 simple empirical relationship known as Archie's Law (Archie, 1942):

$$170 \quad \rho_b = \phi^{-m} \rho_w, \quad (1)$$

171 where the resistivity of the bulk formation (ρ_b) is related to the porosity of the medium (ϕ) raised to the negative
172 power (m), which represents the degree of pore cementation, and the resistivity of the pore fluid (ρ_w) (Glover 2010).
173 This relationship carries a number of simplifying assumptions: the most significant being that the current flow is
174 entirely electrolytic. While more sophisticated formulations of Archie's Law incorporating fluid saturation and
175 interfacial conduction can be found in the literature (e.g., Rhoades et al., 1976; Waxman and Smits, 1968), Eq. (1) is
176 considered to be a reasonable approximation for this environment. Equation (1) is used in this study to evaluate the
177 impact of observed groundwater and surface water aqueous conductivity variations on the bulk formation resistivity.
178 Here, a value of 1.4 was used for the constant m , which is considered reasonable for fractured dolostone. It should
179 be noted that the relative impact of aqueous conductivity changes on the bulk formation resistivity may vary with
180 clay content and pore connectivity due to intrinsic deviations in the m value (Worthington, 1993). Furthermore,
181 orientated fracture networks may result in an anisotropic resistivity response (Steelman et al., 2015b); however,
182 these static properties of the rock will not impact relative changes in resistivity at a fixed location.

183 The electrolytic (fluid phase) resistivity will depend on the concentration and composition of dissolved ions, and
184 viscosity of the pore water (Knight and Endres, 2005). Increasing ion concentrations and temperature will lead to a
185 reduction in formation resistivity. Empirical evidence has shown that resistivity can decrease anywhere from 1 % to
186 2.5 % per °C (Campbell et al., 1948; Keller, 1989; Brassington, 1998). The effects of temperature on our resistivity
187 signals were determined using Arp's law (Arps, 1953):

$$188 \quad \rho_{w2} = \frac{\rho_{w1}(T_1 + 21.5)}{(T_2 + 21.5)}, \quad (2)$$

189 where ρ_w (Ω m) and T ($^{\circ}\text{C}$) represent the resistivity and temperature of the water at two points. This formulation
190 was developed from a least-squares fit to the conductivity of a NaCl solution ranging from 0°C to 156°C; however,
191 the exact relationship between fluid conductivity and temperature will depend on the composition of the electrolytic
192 solution (Ellis, 1987).

193 **3.2 Bedrock Lithology, Fractures and Porosity**

194 The geology was characterized through a series of vertical and angled coreholes along the southern shoreline that
195 were advanced into upper Gasport formation. These drilling activities were part of a broader hydrogeological
196 investigation of groundwater flow and fluxes along the river. A network of riverbed piezometers, bedrock stage

197 gauges, and flux measurement devices were installed between 2013 and 2014 within the pool. Locally, the riverbed
198 morphology can be distinguished in terms of the amount of bedrock rubble or weathered rock fragments covering
199 the exposed rock surface. Figure 2 shows the transition from a rubble dominated riverbed (RDR) to a more
200 competent rock riverbed (CRR); this boundary roughly corresponds to the riffle-pool transition.

201 Geophysical measurements were supported by temperature, specific conductance of the fluid and river stage
202 elevation collected at nearby monitoring points (Fig. 2). The geologic and hydrogeologic data were obtained from a
203 river stage gauge (RSG4), a vertical corehole (SCV6) drilled to a depth of 10.9 m, and an angled corehole (SCA1)
204 drilled to a vertical depth of 31.8 m. The angled corehole plunges at 60° and is orientated at 340°, and therefore,
205 plunges beneath the river with a lateral displacement of 21 m. Coreholes were drilled using a small-diameter
206 portable Hydrocore Prospector™ drill with a diamond bit (NQ size: 47.6 mm core and 75.7 mm corehole diameter)
207 and completed with steel casings set into concrete to a depth of 0.6 m below ground surface (bgs). All coreholes
208 were sealed using a flexible impermeable liner filled with river water (FLUTe™ Flexible Liner Underground
209 Technologies, Alcalde, New Mexico, USA) (Keller et al., 2014).

210 The SCA1 rock core was logged for changes in lithology, vugs, and fracture characteristics, intensity and
211 orientation, including bedding plane partings. Rock core subsamples were extracted for laboratory measurements of
212 matrix porosity using the following procedure: sample was oven dried at 40°C; dimensions and dry mass recorded;
213 samples evacuated in a sealed chamber and imbibed with deionized water; sample chamber pressurized to 200 psi to
214 300 psi for 15 minutes; samples blotted and weighed to obtain saturated mass. Open coreholes were logged using an
215 acoustic (QL40-ABI) and an optical (QL40-OBI) borehole imager (Advanced Logic Technologies, Redange,
216 Luxembourg), to characterize the fracture network.

217 **3.3 Pressure, Temperature, Specific Conductance and River Flow**

218 Temperature, specific conductance and hydraulic pressure data were recorded using a CTD-Diver™ (Van Essen
219 Instruments, Kitchener, Canada) deployed within RSG4 (surface water) and SCV6 (groundwater) at a depth of 10.5
220 m bgs. The transducer in SCV6 was placed near the bottom of the open corehole prior to being sealed with an
221 impermeable liner, thereby creating a depth-discrete groundwater monitoring point. Surface water data were
222 recorded through the full study period while deeper bedrock conditions were recorded from early-September 2014
223 through late-May 2015. All measurements were collected at 15 minute intervals.

224 Vertical temperature profiles were additionally collected along the inclined sealed corehole water column of SCA1
225 from 4-Sep-2014 to 22-May-2015 using an RBRsolo™ temperature logger paired with a RBRsolo™ pressure logger
226 (RBR Limited, Ottawa, Canada). These data were recorded at 0.5 second intervals while the sensors were manually
227 lowered into the water column using a fiberglass measuring tape at a rate of 0.02 m s⁻¹ to 0.03 m s⁻¹. Barometric
228 pressure was collected at the site using a Baro-Diver™ (Van Essen Instruments, Kitchener, Canada). These
229 temperature sensors can resolve changes to 0.5×10⁻⁴ °C with a full-scale response time of 99 % in 2 seconds.

230 Rainfall was recorded at the University of Guelph Turfgrass Institute, located 2 km northwest of the site, while
231 snowfall accumulation was obtained from the Region of Waterloo Airport roughly 18 km south west of the site.

232 Hourly mean river flux was recorded 900 m upstream at the Watson Road gauge operated by the Grand River
233 Conservation Authority. A summary of the weather and river flux data are provided in Fig. 4.

234 **3.4 Riverbed Electrical Resistivity**

235 **3.4.1 Spatial Electrical Resistivity Mapping**

236 Spatial riverbed resistivity distribution was measured using a Geonics EM-31 ground conductivity meter (Geonics,
237 Mississauga, Canada) during a seasonally cool and warm period: mid-summer/low-stage conditions on 7-Jul-2014
238 and early-spring/high-stage conditions on 3-Apr-2013. Measurements were collected at a rate of 3 readings per
239 second with the device operated in vertical dipole mode held ~1 m above the riverbed. The effective sensing depth
240 of this instrument in vertical dipole mode is approximately 6 m (cumulative depth), and is minimally sensitive to
241 conditions above the ground surface (McNeil, 1980). Data was recorded along roughly parallel lines spaced ~1.75
242 m apart orthogonal to the river orientation, with the coils aligned parallel to surface water flow direction. Water
243 depths over the investigated reach varied from <0.1 m in the riffle during low-flow to nearly 1 m in the pool during
244 high-flow conditions. Data sets were filtered for anomalous outliers prior to minimum curvature gridding.

245 **3.4.2 Time-Lapse Electrical Resistivity Imaging**

246 Surface electrical resistivity measurements were collected along two fixed transects orientated orthogonal to the
247 river (Fig. 2), capturing conditions within a pool and riffle sequence (Fig. 3). Line 1 was positioned downstream
248 over a deeper pool section with more substantial bedrock incision into a competent bedrock surface (Fig. 3b, i and
249 ii), while line 2 was situated upstream over a shallower riffle section blanketed by bedrock rubble fragments with less
250 bedrock incision (Fig. 3b, iii).

251 For this study, resistivity cables were constructed using a pair of 25 multicore cables (22 gauge stranded wire, 600V
252 rating) wound within a PVC jacket. The PVC jacket was split open every meter to expose and cut out a single wire
253 that was connected to an audio-style banana plug. Spliced sections of outer PVC jacket were resealed using heat
254 shrink tubing and silicon. This process yielded two 24 channel cables each connected to a single multi-pin
255 connector for direct data logger communication. Electrodes were constructed from half-inch diameter stainless steel
256 rod cut to 6 inch lengths. A hole was drilled on one end of the electrode to receive the banana plug connector.
257 Given the exposed bedrock across the site, a half-inch hole was drilled into the rock at 1 m intervals along the
258 ground surface. In some cases, electrodes were buried beneath a rubble zone of the riverbed, or were pushed into a
259 thin layer of sediment. On the shorelines electrodes were fully implanted into the rock along with a few teaspoons
260 of bentonite clay to minimize contact resistance. Each monitoring line was instrumented with dedicated electrodes
261 and cables that remained in place for the duration of the study.

262 Resistivity measurements were recorded using a Syscal Junior Switch 48 (Iris Instruments, Orléans, France)
263 resistivity meter. A Wenner array was selected for its higher S/N ratio. A dipole-dipole array was tested, but found
264 to be very susceptible to noise (i.e., bad data points resulting from low potentials); this can be attributed to the high-
265 contact resistances with rock combined with the instruments more moderate power capability (max 400 V, 1.3 A).

266 Although the Wenner array geometry resulted in a stronger signal (i.e., where potentials are measured across a pair
267 of electrodes located between the current electrodes), it was less sensitive to lateral variations across the riverbed
268 and did not permit the collection of a reciprocal dataset, which could have been used to assess potential
269 measurement errors (e.g., Slater et al., 2000). Surface resistivity data were recorded on a semi-daily to semi-weekly
270 interval from 18-Jul-2014 to 3-Jul-2015 covering a complete annual cycle, which included a seasonal freeze-thaw
271 cycle, and numerous wetting-drying events accompanied by large river stage fluctuations. The timing of resistivity
272 measurement events are shown along the river flow data in Fig. 4. Resistivity measurements were generally
273 recorded between 8 AM and 1 PM.

274 Measured apparent resistivity data was manually filtered to remove erroneous data points prior to being inverted
275 using RES2DINV v.3.59 (Geotomo Software, Malaysia), which uses the Gauss-Newton least-squares method (Loke
276 and Dahlin, 2002). For this study, a robust inversion scheme was used with a moderate dampening factor applied at
277 the riffle and a slightly higher dampening factor at the pool. These dampening factors were chosen based on the
278 resistivity contrasts observed along the rock surface, along with intermittently noisy measurements at the pool; the
279 width of the model cells were set to half the electrode spacing (i.e., model refinement) to help suppress the effects of
280 large surface resistivity variations on the inversion process. All other parameters within the program were optimized
281 to compensate for high noise and large resistivity contrasts, while achieving the lowest possible model root mean
282 squared (RMS) error (Loke, 2002). Each model was independently inverted with a defined surface water boundary
283 (i.e., stage height above the submerged electrodes) and true aqueous resistivity, both of which were fixed for each
284 model inversion. The same set of initial inversion parameters was applied to all datasets collected along a given
285 line. Model convergence typically occurred within 8 iterations.

286 Temporal variations in bedrock resistivity were assessed within 5 m wide \times 2.5 m deep zones beneath the riverbed
287 based on a resistivity index (*RI*) calculation. This enabled comparison of datasets with different magnitudes of
288 resistivity variation. The *RI* was defined as follows:

$$289 \quad RI_{i,j} = \frac{MZR_{i,j} - MAR}{MAR}, \quad (3)$$

290 where $RI_{i,j}$ = resistivity index for the i^{th} zone on the j^{th} sample date; $MZR_{i,j}$ = mean zone resistivity for the i^{th} zone
291 on the j^{th} sample date; MAR = mean annual resistivity of the entire profile across the full time series for the pool or
292 riffle.

293 **4 Results**

294 **4.1 Bedrock Fracture Network, Temperature and Specific Conductance**

295 Formation contacts of the Eramosa–Goat Island and the Goat Island–Gasport formations were identified in core at
296 depths of 8.6 and 13.0 m bgs, respectively (Fig. 5a). Fractures beneath the river were predominantly horizontal to
297 slightly dipping ($<10^\circ$), and most abundant in the Eramosa and Goat Island. Although vertical and sub-vertical
298 fractures ($>10^\circ$) were relatively less abundant than bedding plane fractures, they were more uniformly distributed

299 with depth. These high-angle fractures terminate at surface as vertical joint sets along two regional orientations: 10°
300 to 20° NNE and 280° to 290° SNW (Fig. 3b, ii). Matrix porosities from the corehole ranged from 0.5 % to 5 % with
301 the lowest porosities observed along the highly weathered riverbed surface and lower portion of the Eramosa
302 Formation. Hydraulic head data collected in the river and at the base of SCV6 (10.5 m bgs) suggest a seasonally
303 sustained downward gradient ranging from 0.21 to 0.26 at the pool; however, this does not necessarily reflect
304 conditions proximal to the streambed.

305 Vertical temperature profiling within the static water column of the FLUTE™ lined SCA1 corehole from 4-Sep-
306 2014 to 22-May-2015 captured seasonal fluctuations in ambient groundwater temperature to depths up to 20 m (Fig.
307 5b), thereby delineating the vertical extent of the heterothermic zone. Temperatures inside the liner ranged from
308 18°C in late-summer, to 5°C in mid-winter. Although fluctuations were observed along the entire 20 m profile, the
309 bulk of the variations (short and long-period) were observed in the upper 10 m bgs.

310 Previous studies using ambient temperature profiling in lined coreholes (Pehme et al. 2010; Pehme et al. 2014)
311 examined the effects of active groundwater flow around static water columns. Pehme et al. (2010) demonstrated
312 how a lined water-filled corehole in thermal disequilibrium with the surrounding formation would exhibit more
313 short-period temperature perturbations along its vertical profile than an equilibrated water column within zones of
314 active groundwater flow. Here, the onset of winter seasonal conditions (9-Jan-2015 through 31-Mar-2015) cooled
315 the corehole water column near the ground surface resulting in density-driven convection within the column, leading
316 to thermal disequilibrium with respect to the surrounding bedrock resulting in subtle temperature perturbations as
317 the water column cooled toward 5°C. The magnitude and frequency of the perturbations observed in Fig. 5b during
318 these cooler periods correspond to areas of increased fractures (Fig. 5a) supporting active groundwater flow.

319 Specific conductance and temperature of surface water (RSG4) and groundwater (SCV6) corresponding to
320 geophysical sampling events (Fig. 4) are presented in Fig. 6. These data indicate that surface water specific
321 conductivity varied within a much narrower range than the actual (uncompensated) conductivity, which includes the
322 effects of temperature. While the overall impact of temperature and ionic concentration on the specific conductance
323 of surface water were similar (i.e., equivalently dynamic), variations associated with ionic concentration appear
324 more erratic, exhibiting sharper fluctuations over shorter periods of time. For instance, major precipitation events
325 coinciding with measurement events 13, 26 and 31 (refer to Fig. 4) were accompanied by short-period reductions in
326 surface water conductivity and increases in temperature. Seasonal atmospheric temperature trends resulted in more
327 gradual, yet sustained reductions in aqueous conductivity. In comparison, the groundwater specific conductance at
328 10.5 m bgs was comparatively stable during the study period, exhibiting a moderate temperature driven decline
329 superimposed by shorter-period fluctuations associated with ion concentration.

330 Figure 7 shows the potential impact of these observed specific conductivity and temperature variations (based on
331 Fig. 5b and 6) on the bulk formation resistivity using Eq. (1) and Eq. (2) for three representative porosity values.
332 Porosities of 1 % and 5 % correspond to the values obtained in core, while a porosity of 35 % is assumed to
333 represent the maximum porosity of a weathered or broken rubble zone (i.e., unconsolidated porous sediment).

334 These calculations indicate that variations in temperature will likely be the primary driver in formation resistivity
335 dynamics. For instance, water temperature could affect the formation resistivity by as much as 46 % (between
336 isotherms), based on the observed range in groundwater and surface water temperatures, respectively. In
337 comparison, measured aqueous conductivity ranges (along an isotherm) for groundwater and surface water would
338 affect the formation resistivity by 18 % and 36 %, respectively. These maximum temperature effects represent end-
339 member conditions for a given specific conductance and effective porosity estimate.

340 **4.2 Sub-Riverbed Electrical Resistivity Distribution**

341 Two ground conductivity surveys were conducted across the riverbed surface to assess spatial variability in bulk
342 formation resistivity and its relationship to riverbed morphology (e.g., pool vs. riffle): resistivity snapshots were
343 collected on 7-Jul-2014 during low-flow conditions ($1.30 \text{ m}^3 \text{ s}^{-1}$) and on 3-Apr-2013 during high-flow conditions
344 ($6.81 \text{ m}^3 \text{ s}^{-1}$) (Fig. 8a). The percentage change in resistivity from low to high-stage periods is shown in Fig. 8b. The
345 daily average river flows for the years 2013 and 2014 were $3.5 \text{ m}^3 \text{ s}^{-1}$ and $3.3 \text{ m}^3 \text{ s}^{-1}$, respectively.

346 Two main observations can be made from the changes observed between warmer low-flow and cooler high-flow
347 conditions. First, the southern shoreline exhibited a 10% to 15% reduction in resistivity along the southern shoreline
348 within the pool and along the cut bank (outer edge of the elbow). These areas were characterized by more
349 competent rock with exposed bedding plane fractures and vertical joint (Fig. 3b, i). Secondly, a broader and more
350 variable increase in resistivity upwards of 20% to 25% was observed northward into the thalweg, along the slip-off
351 slope (inner edge of the elbow) and the eastern and western portions of the reach; the rock surfaces in these areas
352 were more weathered with large irregular rock fragments and dissolution features, and limited exposure of
353 horizontal bedding plane fractures. A lower resistivity zone (blue area) was identified upstream within the northern
354 portion of the riffle section (Fig. 3b, iii). The riffle portion of the river was also accompanied by a break in the high
355 resistivity trend observed along the south shoreline. A lower average resistivity was observed during warmer low-
356 flow conditions indicating that a portion of the response may be dependent on formation temperature (i.e., 5°C to
357 20°C fluctuations). However, the reduction in resistivity along the cut bank during cooler high-stage period indicates
358 an increase in the specific conductance of the pore fluid, which would be consistent with increased baseflow and
359 groundwater discharge to the river.

360 **4.3 Time-Lapse Electrical Resistivity Imaging**

361 **4.3.1 Electrical Resistivity Models**

362 Figure 9 provides a summary of the inverted model results at the pool (Fig. 9a) and riffle (Fig. 9b) sections for the
363 full study period. The mean inverted model resistivity and data range for each sample event is presented along with
364 the number of apparent resistivity data points removed from the dataset prior to inversion, and the root mean squared
365 (RMS) error of the inverted model. The partially-frozen and frozen ground conditions were accompanied by higher
366 signal noise due to a systematic increase in electrode contact resistance and reduction in pore fluid connectivity
367 (liquid water saturation near the surface); here, lower potentials resulted in more frequent failed measurement, which
368 had to be removed from the dataset prior to inversion. Therefore, the noisier datasets collected in these periods were

369 accompanied by higher model RMS errors. A subset of the inverted resistivity models over the annual cycle (i.e.,
370 samples a–h identified in Fig. 9) are shown in Fig. 10. These snapshots capture the spatiotemporal evolution of
371 predominant geoelectrical conditions beneath the riverbed.

372 Spatial electrical resistivity data were highly variable across the pool (Fig. 10a–h). The highest resistivities were
373 observed along the south shoreline, which coincided with the presence of competent bedrock (Fig. 3b, i) with
374 bedding plane fractured and vertical joint sets. Similarly resistive conditions extended southward onto the
375 floodplain. Subsurface conditions became less resistive toward the north shoreline, which coincided with the
376 presence of increased vertical fracturing and dissolution-enhanced features, mechanically broken or weathered
377 bedrock, and a thin layer of organic rich sediment alongside the north shoreline and floodplain. Initial surveys
378 conducted across the pool on 25-Jul-2014 identified a relatively low resistivity zone ($<1000 \Omega \text{ m}$) extending 2 m
379 beneath the riverbed that spanned the full width of the river. Measurements on 26-Sep-2014 through 24-Dec-2014
380 captured the retraction of this zone toward the north shore. During this period the resistivity across the full transect
381 increased only slightly. The onset of frozen ground and river conditions on 29-Jan-2014 resulted in an abrupt shift
382 in the resistivity distribution. A high resistivity zone formed above the water table across the southern floodplain
383 and was accompanied by an increase in resistivity across the full river profile. It is important to note that these
384 frozen periods were accompanied by higher model RMS errors, and thus, our interpretation of these data focus on
385 long-period trends. The formation of river ice (visually observed basal and surface ice) may have altered the true
386 geometry of the surface water body represented in the model, potentially contributing to the higher RMS errors
387 along with the overall noisier measurements during this time (Fig. 9). The arrival of seasonal thaw conditions on
388 27-Mar-2015 was accompanied by reduced resistivities across the river as rock and river ice progressively thawed
389 and was mobilized by spring freshet. Further seasonal warming on 6-May-2015 and 3-July-2015 resulted in a
390 systematic decrease in riverbed resistivity from the north to south shoreline.

391 Riverbed resistivity across the riffle portion of the river (Fig. 10a–h) was markedly different with respect to the
392 distribution and magnitude of resistivity fluctuations. The riffle exhibited a zone of comparatively low resistivity
393 ($<100 \Omega \text{ m}$) that extended slightly deeper than that at the pool, to a depth of 3 m. The initial survey on 26-Jul-2014
394 identified a zone of very low resistivity that progressively became more resistive over time (26-Sep-2014 through
395 24-Dec-2014). Much like the pool, however, this low resistivity zone reverted back toward the north shoreline. The
396 onset of seasonally frozen river conditions was accompanied by an increase in resistivity across a significant portion
397 of the riverbed. Inverse models during frozen water conditions were again accompanied by higher RMS errors,
398 which we attribute to the formation of river ice. Unlike the pool, which experienced the formation of a substantial
399 zone of ground frost along the south shore, less ground frost was observed at depth along the riverbanks bounding
400 the riffle. Spring thaw brought reduced resistivities across the riverbed with subtle lateral variations, followed by
401 the beginnings of a less-resistive riverbed zone emanating southward from the north shoreline. The bedrock
402 resistivity below 3 m depth remained relatively constant through the monitoring period.

403 **4.3.2 Spatiotemporal Resistivity Trends**

404 A resistivity index (RI) was calculated using Eq. (3) to compare spatiotemporal variations in electrical resistivity
405 within predefined zones of the bedrock beneath the river (Fig. 11); zones A and D represent conditions along the
406 south and north riverbank, while zones B and C represent conditions within the southern and northern portions the
407 river. These zones were defined based on their representative areas and the magnitude of the temporal fluctuations
408 observed over the full monitoring period (Fig. 10). A RI of zero indicates a mean zone resistivity (MZR) that is
409 equal to the mean annual resistivity (MAR) of the whole profile. An index of +1 indicates a resistivity that is twice
410 the annual mean, while an index of -0.5 indicates a resistivity that is half the annual mean.

411 The RI time-series for the pool (Fig. 11a) and riffle (Fig. 11b) capture the magnitude and frequency of the temporal
412 variability observed within these four zones. Relative to the MAR, the pool exhibited larger and more frequent
413 fluctuations in resistivity compared to the riffle. The south shoreline (zone A) at the pool was more dynamic than
414 the corresponding zone at the riffle; zone A at the pool encompasses a larger unsaturated zone, which is more likely
415 impacted by changes in temperature and saturation, especially during the freezing and thawing period. The north
416 shoreline (zone D) at the pool and riffle exhibited lower than average resistivities with relatively minimal transience
417 over the study period, with the exception of the mid-to-late-winter freeze-up. Here, a variable layer of sediment and
418 organic matter with higher water content likely moderated freeze-thaw fluctuations relative to sections of exposed
419 rock. Conditions below the riverbed (zones B and C) exhibited both longer-period (seasonal) and shorter-period
420 (intra-seasonal) fluctuations. While the relative changes observed at the pool were larger than the riffle, similar
421 seasonal trends were observed at each location. Zones B and C at the pool were mutually consistent, while those at
422 the riffle were less consistent.

423 Although perturbations were observed in the resistivity beneath the riverbed before and after winter freeze-up (e.g.,
424 zones B and C), the responses were dampened relative to the winter period. Events 13, 26 and 31 (Fig. 6), which
425 correspond to periods of increase precipitation, may coincide with observed perturbations in the RI; however, based
426 on these data it is not clear whether the riverbed resistivity and surface water responses are mutually consistent.
427 Here, a limited relationship may suggest that groundwater-surface water interaction does not occur in this section of
428 the river or that groundwater discharge is strong enough to limit potential groundwater-surface water mixing at this
429 particular location. Therefore, these observed geophysical dynamics within the riverbed may be more associated
430 with seasonal temperature transience with secondary influence due to solute-based fluctuations.

431 **5.0 Discussion**

432 **5.1 Influence of Water Properties on Formation Resistivity**

433 Riverbed electrical resistivity mapping during low and high stage periods identified a spatiotemporal response
434 within the upper 6 m of rock with the southern cut bank exhibiting a reduction in resistivity while the remaining
435 portions of the river increased in resistivity. This spatiotemporal response, together with observed bedrock surface
436 conditions, indicates that riverbed morphology strongly impacts groundwater dynamics below the riverbed.

437 Long-term resistivity monitoring along the fixed profiles over the pool and riffle portions of the river revealed a
438 transient zone within the upper 2 m and 3 m of bedrock, respectively. In particular, the formation of a low

439 resistivity zone (high electrical conductivity) was observed during the warmer summer period that diminished as the
440 environment cooled. While pore water conductivity depends on electrolytic concentration and temperature, their
441 independent variability could not be decoupled across the entire study area given *in-situ* sensor deployment
442 limitations in a bedrock environment. Although this uncertainty in the driving mechanism of observed electrical
443 changes below the riverbed hindered our ability to definitively define the vertical extent of a potential groundwater-
444 surface water mixing zone, our geophysical data set does suggest that a groundwater-surface water mixing in a
445 bedrock environment may be more limited due to the lower effective porosity of rock and heterogeneous and
446 anisotropic fracture distributions.

447 Aqueous temperature and specific conductance measurements collected in the river stage gauge (RSG1) and shallow
448 bedrock well (SCV6) provided end-member conditions (Fig. 6). These data were used to assess the influence of
449 aqueous conditions on bulk formation resistivity (Fig. 7). While some degree of overlap was observed between
450 groundwater and surface water properties, they were generally differentiable across the study area. That being said,
451 aqueous temperature fluctuations dominated the bulk electrical response over the full annual cycle. Given the
452 impact of temperature on the bulk formation resistivity, observed bedrock resistivity dynamics are attributed to
453 changes in water/rock temperature with secondary effects caused by changes in electrolytic concentration. These
454 findings are consistent with Musgrave and Binley (2011), who concluded that temperature fluctuations over an
455 annual cycle within a temperate wetland environment with groundwater electrical conductivities ranging from 400
456 $\mu\text{S cm}^{-1}$ to 850 $\mu\text{S cm}^{-1}$ dominated formation resistivity transience. Of course, our annual temperature range was
457 more extreme than that of Musgrave and Binley, we examined electrical dynamics within a less-porous, more
458 heterogeneous and anisotropic medium, and captured a broader range of seasonal conditions including ground frost
459 and riverbed ice formation.

460 Measurements collected with a shorter time-step (diurnal) and shorter electrode spacing may have captured more
461 transient rainfall or snowpack melt episodes focused along discrete fracture pathways, possibly leading to the
462 identification of electrolytic-induced transients beneath the riverbed more indicative of a groundwater-surface water
463 mixing zone. Based on the short-period of intraseasonal fluctuations observed in Fig. 11, and the timing and
464 duration of major precipitation or thawing events (5 to 7 day cycles) (Fig.4), it is reasonable to assume that our
465 geophysical time step (days to weeks) was accompanied by some degree of aliasing. Finally, it is possible that
466 shallower sections of rock within the river exposed to direct sunlight during the day, which can vary depending on
467 cloud cover (daily) and the sun's position in the sky (seasonally), may have exhibited a wider range, or more
468 transient temperature fluctuations, than those areas beneath or adjacent to a canopy. A closer inspection of the
469 unfrozen temporal response in zone B reveals a wider range in resistivity relative to the more northern zone C. At
470 this latitude in the northern hemisphere the south shore will receive more direct sunlight; therefore, it is possible that
471 the shallow rock on this side of the river experienced greater fluctuations in temperature (both seasonally and
472 diurnally), thereby contributing to the observed geoelectrical dynamics. Although diurnal fluctuations in surface
473 water temperature can be significant ($>10^{\circ}\text{C}$) (e.g., Constantz et al. 1994; Constantz 1998), the relative effects of

474 such transient temperature fluctuations on our geophysical measurements cannot be assessed given our more
475 seasonally-scaled measurement sampling interval.

476 **5.2 Formation of Ground Frost and Anchor Ice**

477 A dramatic increase in bedrock resistivity was observed with the onset of freezing ground conditions; this can
478 impact a wide range of infrastructure (e.g., dams, hydropower generation), ecologic (e.g., alteration of fish and
479 benthic habitats) and hydraulic functions (e.g., river storage, baseflow) (Beltaos and Burrell 2015). The formation of
480 a highly resistivity zone consistent with a seasonal frost front within the unsaturated portion of the riverbank (Fig.
481 10e; zone A in Fig. 11a), and the accumulation of river ice resulted in marked changes in resistivity. These winter
482 season effects are readily evident in Fig. 11 at the pool and riffle. Here, the magnitude of the resistivity increase
483 may reflect a potential reduction in the hydraulic connectivity between surface water and groundwater during the
484 winter months.

485 Ground frost primarily formed along the riverbank over the southern floodplain at the pool. This topographically
486 higher area was relatively devoid of large shrubs and trees (Fig. 2), and thus, likely experienced more severe weather
487 conditions (e.g., higher winds resulted in less snow pack to insulate the ground). These conditions would promote
488 the formation of a thicker frost zone which propagated to the water table (pool in Fig. 10e). The adjacent northern
489 riverbank and those up-gradient at the riffle were topographically lower (i.e., thinner unsaturated zone) and were
490 sheltered by large trees. The formation of a seasonal frost zone along the riverbank may have implications to
491 baseflow dynamics during the winter months and early-thaw period.

492 A simple sensitivity analysis of the inversion process using different constraints on surface water geometry and
493 aqueous electrical resistivity indicated that model convergence was highly sensitive to surface water geometry and
494 aqueous resistivity. For instance, setting an aqueous resistivity of one-half the true value led to poor model
495 convergence and unrealistic resistivity values for bedrock. The riffle was relatively less sensitive to surface water
496 properties likely because of its overall lower river stage compared to the pool, and hence, relatively lower impact of
497 the surface water body on the apparent resistivity measurement. This sensitivity to surface water properties is a
498 consequence of the high electrical conductivity of the surface water relative to high resistivity bedrock.

499 Anchor ice reduced the electrical connectivity across the riverbed, while the ice crust along the surface of the water
500 altered the effective geometry of the water body, further influencing the inverse solution. While the formation of
501 river ice was accompanied by higher RMS errors at the pool (>6 %) and riffle (>4 %) (Fig. 9) direct measurements
502 of river ice thickness and spatial extent could not be collected, and thus explicitly incorporated into the surface water
503 layer geometry during the winter months.

504 **5.3 Implications to the Conceptualization of Groundwater-Surface Water Exchange in Bedrock Rivers**

505 The fractured dolostone in this study consists of a visible orthogonal joint network approximately orientated at 10°
506 to 20° NNE and 280° to 290° SNW, consistent with the regional joint orientations, with frequencies ranging from
507 centimeter to sub-meter scale where exposed at surface. Streambed resistivity measurements indicate a seasonally

508 dynamic groundwater zone within the upper 2 m to 3 m of riverbed. A less-resistive zone ($<1000 \Omega \text{ m}$) was
509 observed beneath the pool emanating from the north shoreline during warmer low-flow periods (July and August
510 2014). This zone retracted in late-summer but showed signs of reappearance in early-July 2015. A similarly
511 evolving low-resistivity zone ($<100 \Omega \text{ m}$) was observed across the riffle, but was more variable across the river
512 transect. Although dynamic fluctuations in temperature and aqueous conductivity support the potential existence of
513 a groundwater-surface water mixing zone, it is not yet clear how these geoelectrical dynamics were influenced or
514 enhanced by fluid flow in the discrete fractures, exchange between the mobile and immobile pore-phase, and
515 seasonal thermal gradients across the riverbed.

516 Discrete fracture networks and dissolution-enhanced features will result in a more heterogeneous and anisotropic
517 groundwater-surface water mixing zones compared to porous unconsolidated sediment. Swanson and Cardenas
518 (2010) examined the utility of using heat as a tracer of groundwater-surface water exchange across a pool-riffle-pool
519 sequence. Observed thermal patterns and zones of influence (i.e., effective mixing zones) in their study were
520 consistent with conceptual models depicting a pool-riffle-pool sequence. While similar temperature dynamics may
521 be expected across the pool-riffle-pool sequence in a bedrock environment, our coarser temporal sampling interval
522 (days to weeks) combined with our smoothed resistivity models limited our ability to capture subtle diel temperature
523 transience across discrete fractures or flow features. Although the electrical resistivity method was not able to
524 definitively differentiate between groundwater and surface water, our geophysical measurements do provide insight
525 into the magnitude, lateral extent and spatiotemporal scale of geoelectrical transience, which are largely driven by
526 temperature fluctuations within the upper few meters of rock.

527 **6.0 Conclusions**

528 Induced resistivity measurements collected across a 200 m reach of the Eramosa River during low and high-stage
529 periods showed that spatiotemporal variations in resistivity will be dependent upon riverbed morphology, i.e., the
530 exposure of bedding plane and vertical fractures, and competency of the rock surface. Complementary fracture and
531 temperature profiling within the open and lined borehole revealed abundant active groundwater flow zones spanning
532 the upper 8-10 m of bedrock, with strong intra-seasonal and seasonal temperature variations along horizontal
533 fracture sets. While surface resistivity profiles captured geoelectrical dynamics within the upper few meters of rock,
534 these data represent indirect evidence of riverbed transience resulting from changes in groundwater temperature and
535 specific conductance. Geoelectrical transience was primarily governed by seasonal temperature trends with
536 secondary effects arising from porewater conductance; however, spatiotemporal variations in temperature and
537 specific conductance could not be decoupled beyond the fixed monitoring points.

538 Time-lapse electrical resistivity imaging of the pool and riffle portion of the river, sampled on a semi-daily to semi-
539 weekly interval, showed consistently higher resistivity at the pool with more elevated resistivities along the south
540 shoreline. Seasonal cooling was accompanied by the formation of a higher-resistivity zone emanating from the
541 south shore to north shore in both the pool and riffle. This resistivity trend reversed during the seasonal warming
542 cycle, becoming less-resistive toward the south shoreline as seasonal temperatures increased and river stage
543 decreased. The formation of ground frost and basal ice along the riverbed had a strong impact on the seasonal

544 resistivity profiles during the winter months. Intraseasonal geoelectrical transience associated with major
545 precipitation events, which were accompanied by short-period perturbations in surface water temperature and
546 specific conductance had a relatively small impact on sub-riverbed resistivity. This could be explained by a
547 seasonally-sustained groundwater discharge zone across this reach of the river, which would have limited or
548 moderated electrical resistivity changes associated with surface water mixing with groundwater. The formation of a
549 transient 2 m thick low-resistivity zone within the pool in mid-summer appears to be associated with an increase in
550 surface water/bedrock formation temperatures during low river stage. During this time, the riffle was characterized
551 by a 3 m thick low-resistivity zone spanning the entire width of the river, underlain by more resistive material.
552 These lower resistivities at the riffle suggest the presence of more porous bedrock consistent with dissolution-
553 enhanced rock and/or a layer of weathered bedrock zone overlying more competent bedrock.

554 Spatiotemporal resistivity dynamics were governed by riverbed morphology together with seasonal changes in water
555 temperature and electrolytic concentration. During the winter, ground frost and basal ice along the riverbed
556 strongly influenced the electrical resistivity across the riverbed. Observed intraseasonal (i.e., spring, summer, fall)
557 geoelectrical changes beneath the riverbed were strongly dependent upon seasonal temperature trends, with the pool
558 and riffle exhibiting variable horizontal and vertical zones of influence. While the riverbed resistivity was highly
559 susceptible to seasonal atmospheric temperature trends, such transience could have implications to biogeochemical
560 processes, benthic activity, and macro-scale hyporheic zone processes.

561 This study demonstrates that surface electrical resistivity has the capacity to detect and resolve changes in electrical
562 resistivity due to transience in temperature and specific conductance within a shallow bedrock river over a complete
563 annual cycle. Imaging the magnitude and scale of transience within the riverbed will be critical to the advancement
564 of our understanding of mechanisms controlling groundwater-surface water exchange within fractured bedrock
565 rivers. Given the challenges and costs associated with installing sensors within rock and effectively sampling a
566 heterogeneous flow system, minimally invasive surface and borehole geophysical methods offer an ideal alternative,
567 and possibly more effective approach for long-term groundwater-surface water monitoring of bedrock environments
568 by reducing instrumentation costs and impacts to ecosensitive environments.

569 **DATA AVAILABILITY**

570 The data used in this study are presented in the figures. Complete monitoring data sets (Figures 9 and 10) and can be
571 made available upon request from the corresponding author.

572

573 **TEAM LIST**

574 **Steelman, Kennedy, Capes, Parker**

575

576 **AUTHOR CONTRIBUTION**

577 **Steelman** designed the experiment, conducted the surveys, and analysed the geophysical data; **Kennedy** designed
578 the borehole network, logged the core, instrumented the hydrological monitoring network; **Capes** designed and
579 collected the temperature profiles, logged the core and supported hydrological data collection and interpretation;
580 **Parker** contributed to the design of the hydrological geophysical monitoring network, and supported conceptual
581 understanding of groundwater flow through fractured rock.

582

583 **COMPETING INTERESTS**

584 The authors declare that they have no conflict of interest.

585

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593

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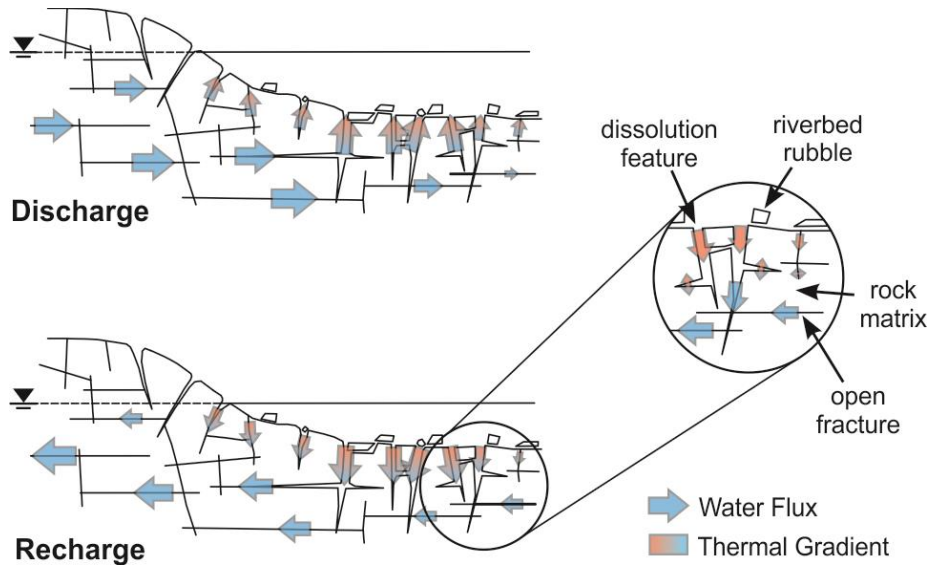
Figures

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Figure 1: General conceptual model of the groundwater flow system beneath a fractured bedrock river.

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Groundwater-surface water mixing is controlled by open fractures and dissolution-enhanced features with secondary

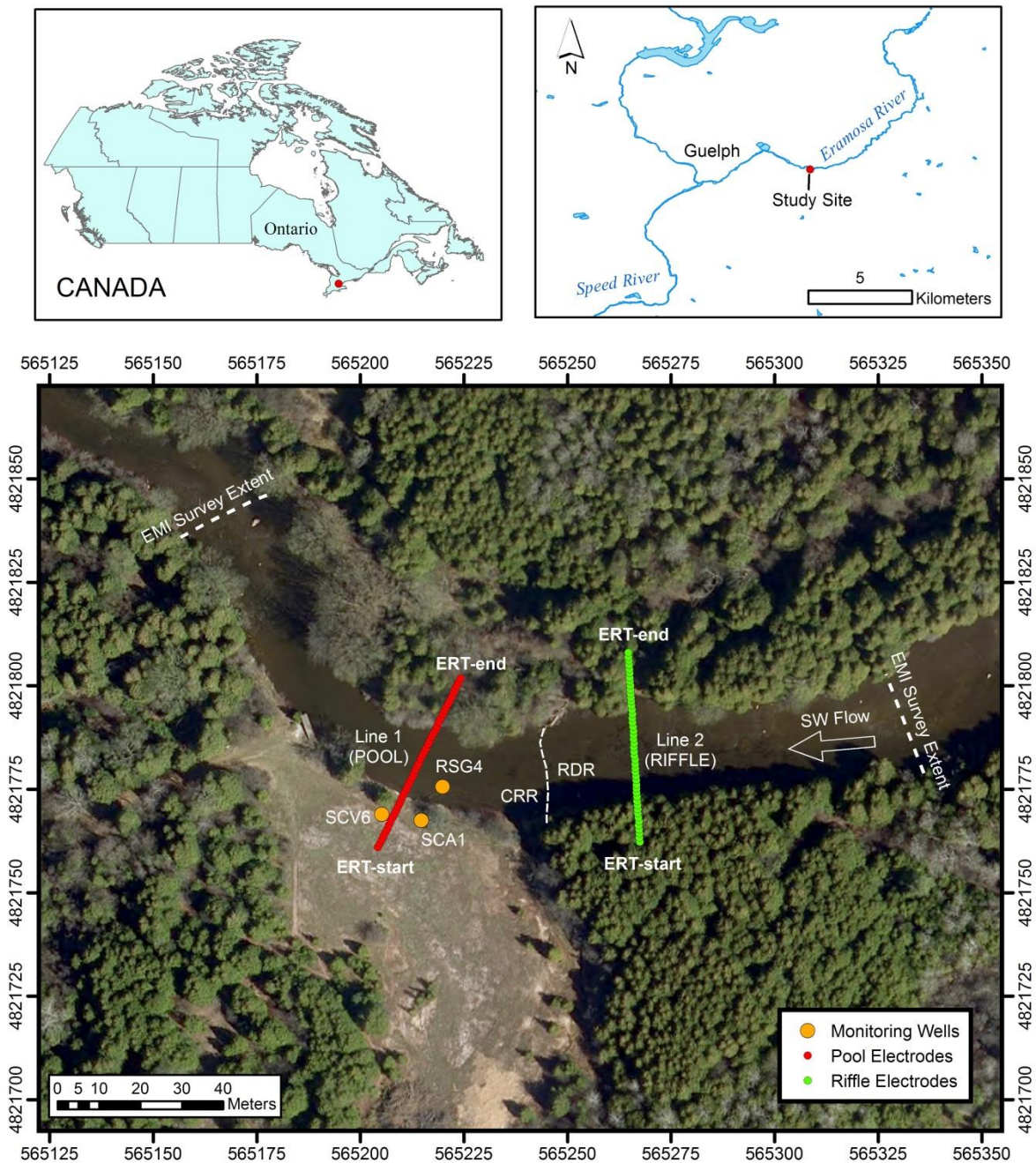
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exchanges (flux or diffusion) occurring between fractures and rock matrix.

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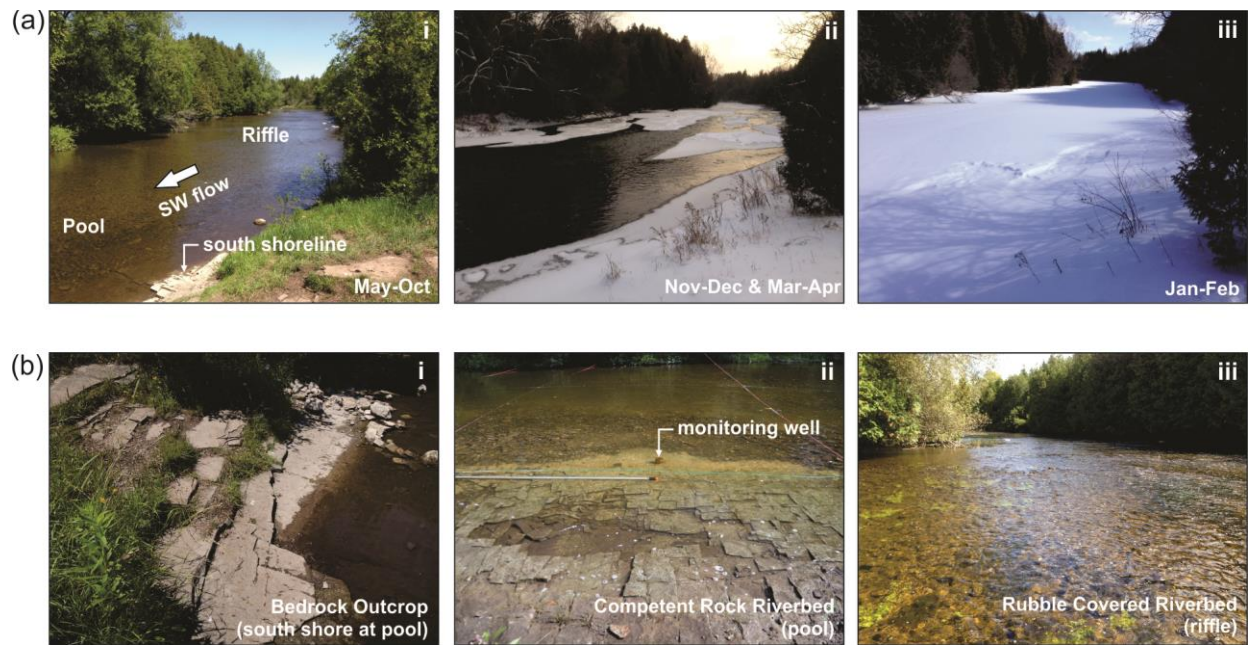


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823 Figure 2: Field site located along the Eramosa River near Guelph, Ontario, Canada. The spatial extent of the
824 electromagnetic induction (EMI) surveys, coreholes and groundwater-surface water monitoring points, and fixed
825 electrical resistivity tomography (ERT) transects are shown relative to the riffle-pool sequence. The riverbed is
826 described as either rubble dominated riverbed (RDR) or competent rock riverbed (CRR) surface.

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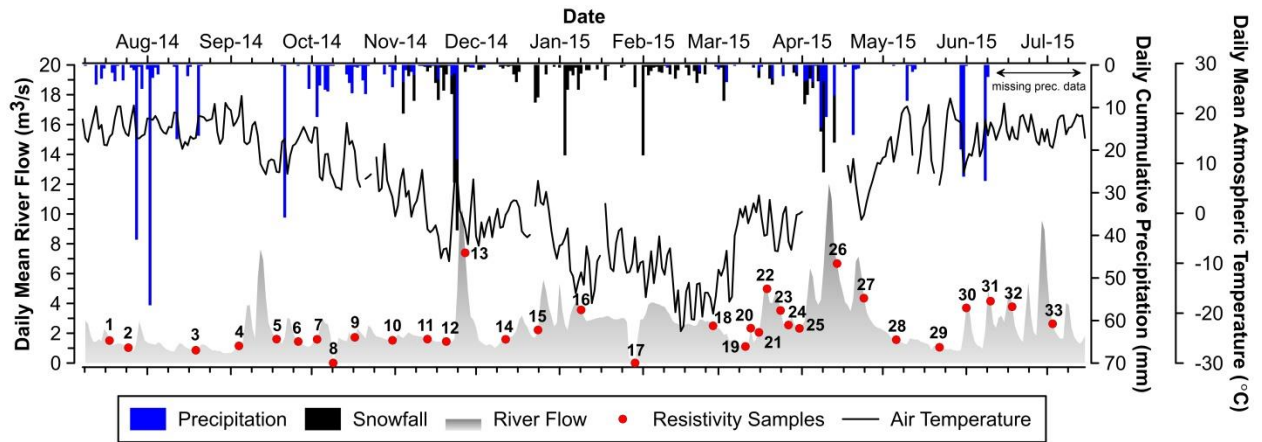


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834 Figure 3: (a) Images of the river during monitored study period. (b) Examples of vertical and horizontal fracturing
835 within pool and rubble covered portions of the riverbed bedrock.

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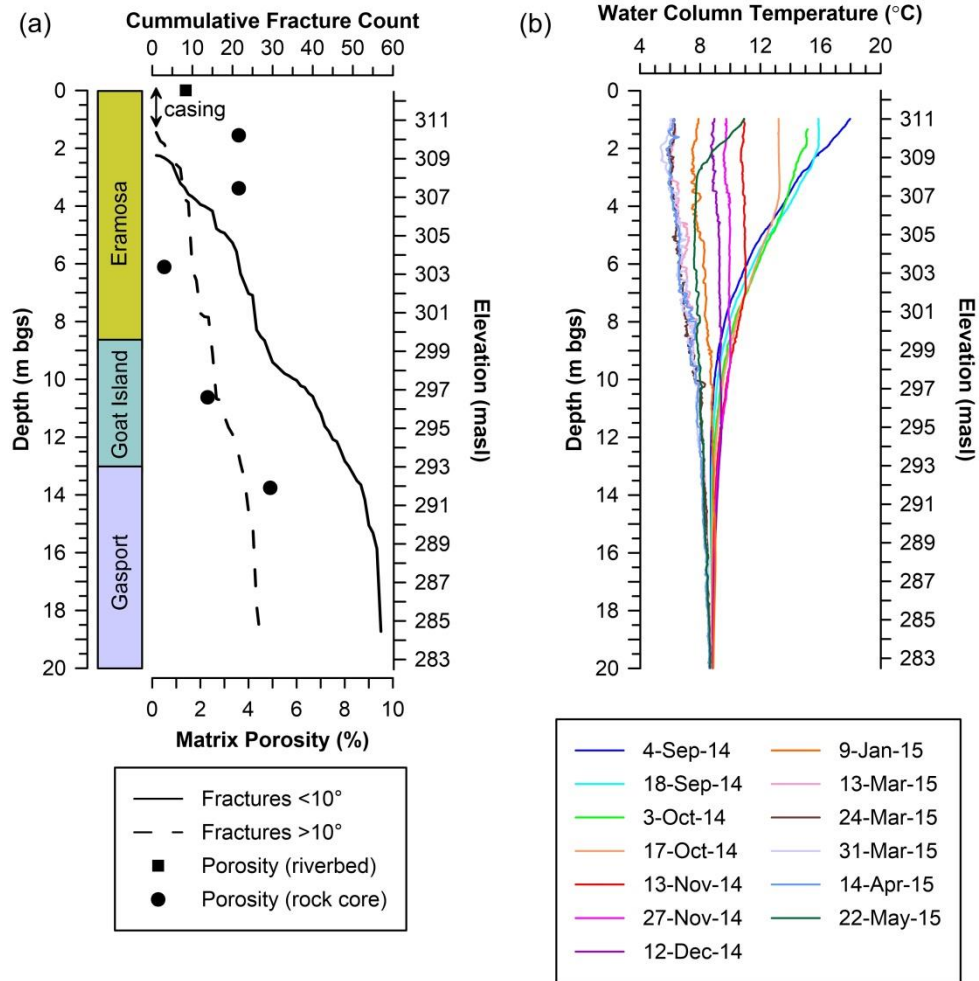
Figure 4: Continuously monitored atmospheric conditions and river flow from Watson Gauge during the study period with superimposed resistivity geophysical measurement events between 18-Jul-2014 and 3-Jul-2015. Note: snowfall is presented as snow water equivalent.

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853 Figure 5: (a) Interpreted rock core from SCA1 (angled corehole plunging at 60° with an azimuth of 340°). Fracture

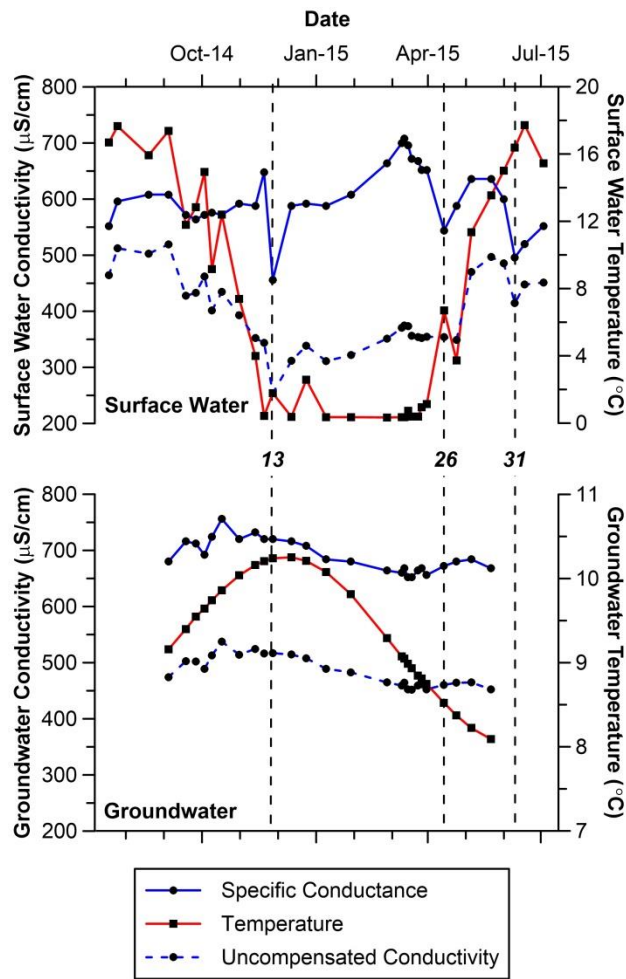
854 frequency and orientations were obtained using an acoustic televiewer log, while matrix porosity measurements

855 were obtained from subsamples of the continuous core. (b) Corehole temperature profiles of the SCA1 Flute™

856 sealed water column.

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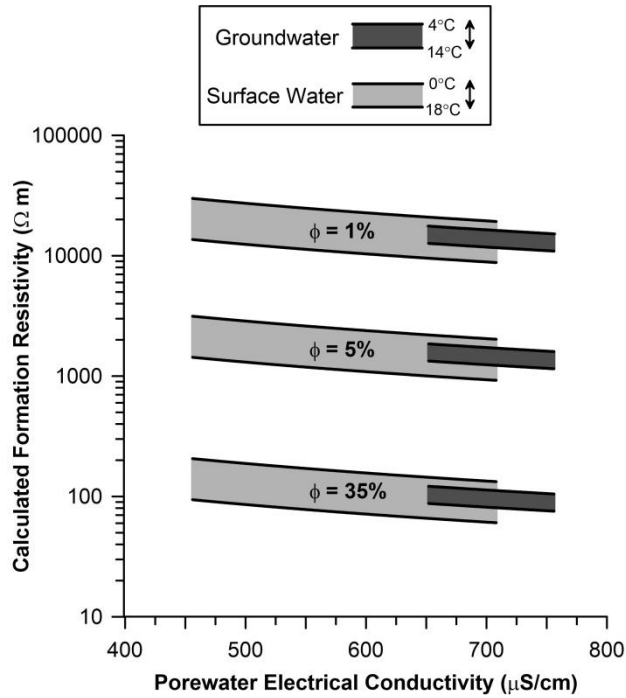


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860 Figure 6: Specific conductance, temperature and uncompensated aqueous conductivity of surface water at RSG4 and
861 groundwater at the bottom of SCV6 (10.5 m bgs). Uncompensated conductivity represents the actual conductivity
862 of the porewater after re-incorporating the effect of temperature using the sensors internal temperature-conductivity
863 correction factor.

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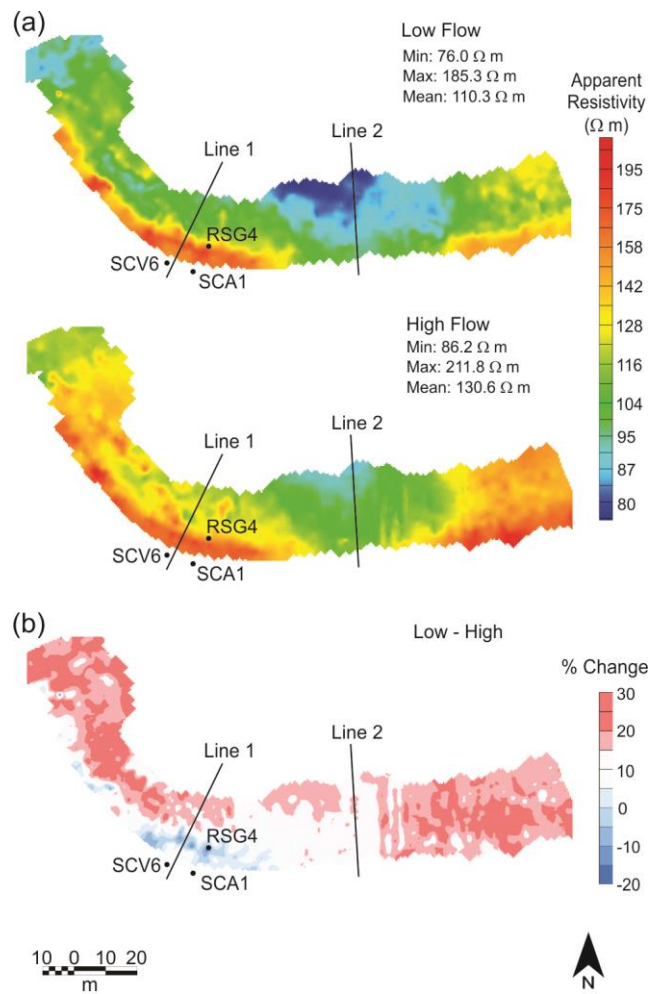


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Figure 7: Calculated formation resistivity based on Eq. (1) and measured variations in surface water and groundwater electrical conductivity including effects of temperature based on Eq. (2). A cementation factor of 1.4 was used to represent the fractured dolostone bedrock. Measured water conductivity and temperature were obtained from CTD-Diver™ sensors deployed in RSG4 (surface water) and SCV6 (groundwater at a depth of 8 m bgs), and the continuous RBR™ temperature profiles shown in Fig. 5b. These data show the potential range in formation resistivity based on the measured range in specific conductance and along with the potential impact of temperature for three different porosity values. Porosities of 1 % and 5 % correspond to the range measured in the core, while a value of 35 % would be representative of a rubble zone.

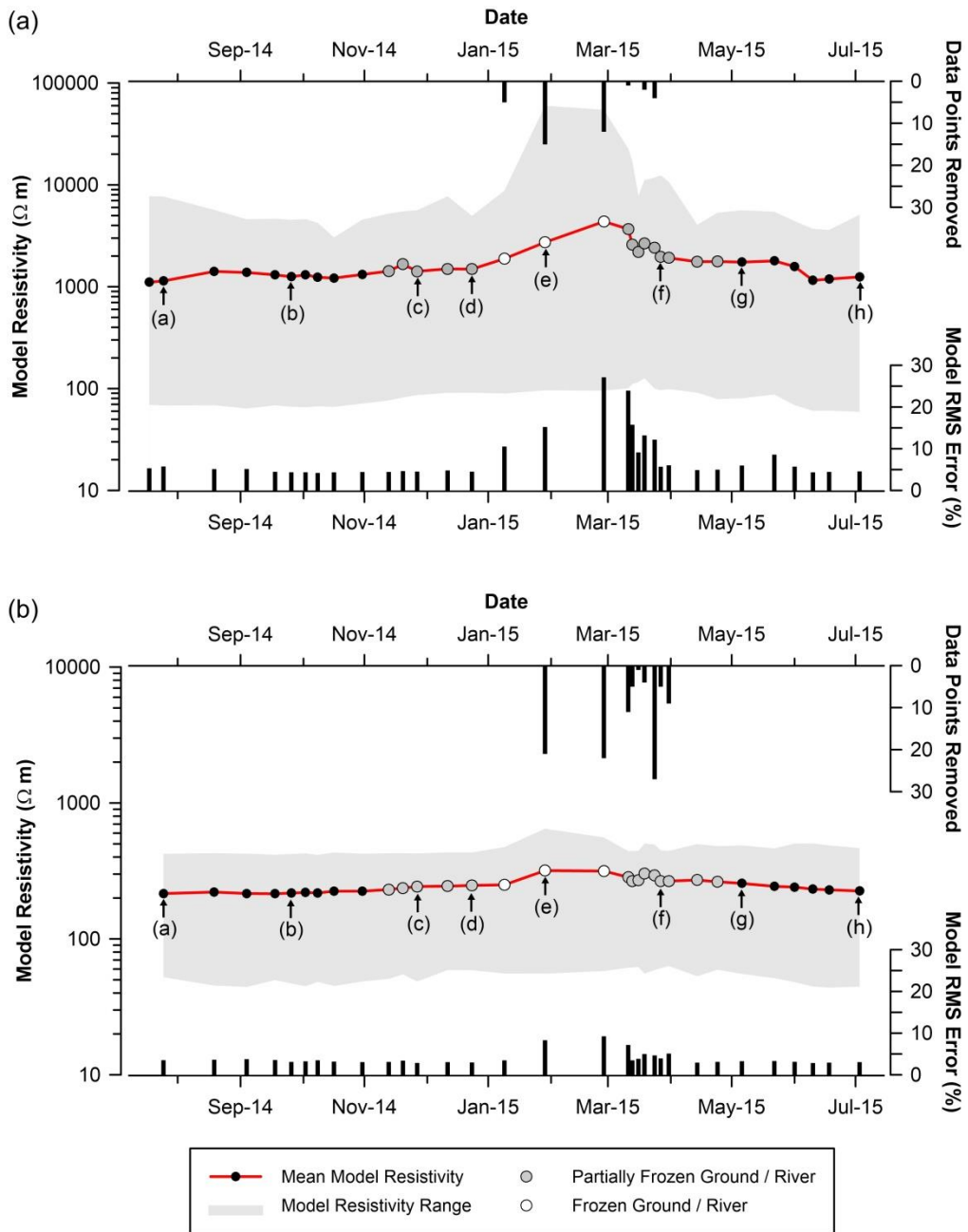
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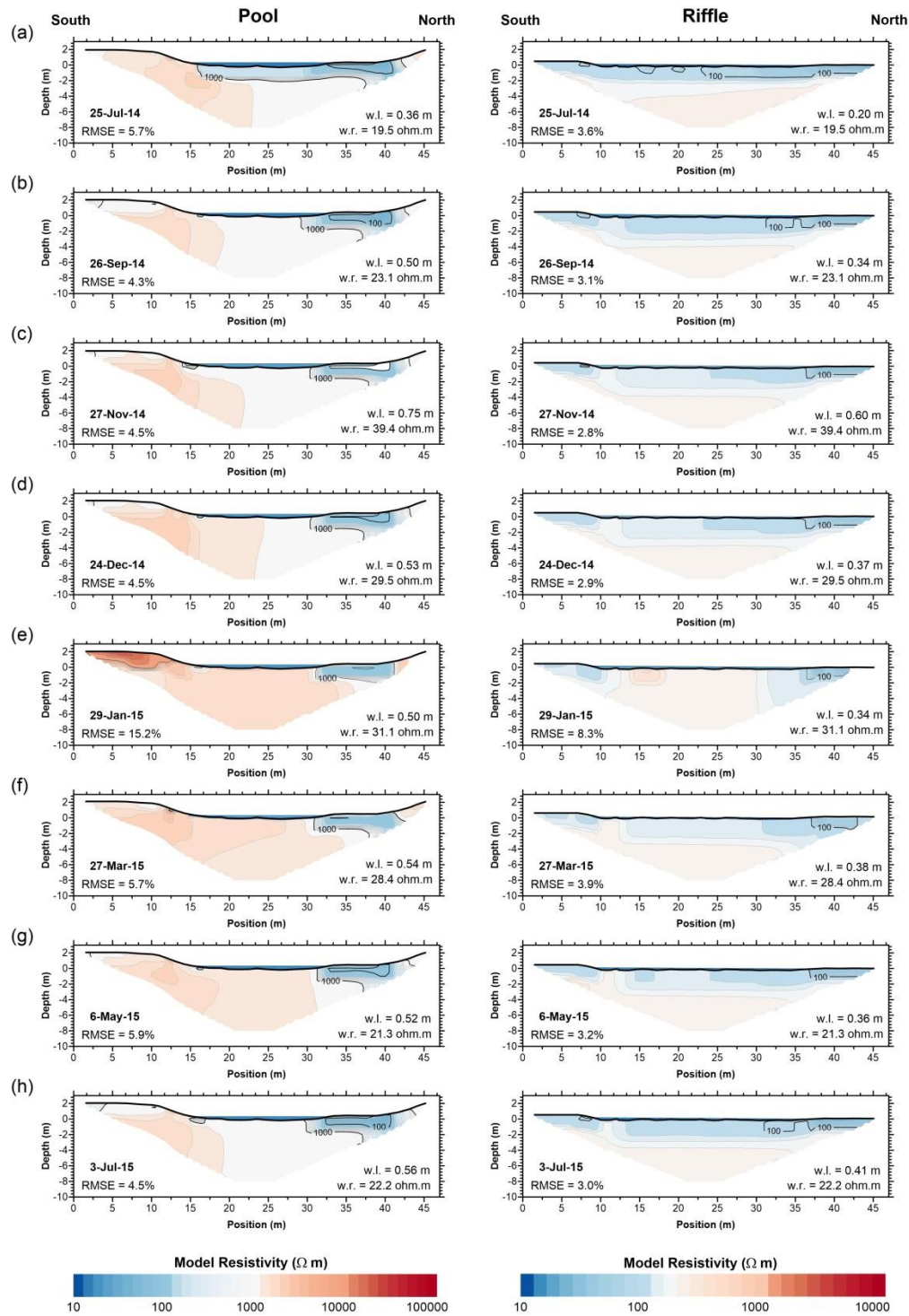
881 Figure 8: Riverbed resistivity obtained using an EM-31 ground conductivity meter during low-flow/low-stage
882 conditions on 7-Jul-2014 and high-flow/high-stage conditions on 3-Apr-2013. (b) Percentage change in apparent
883 resistivity from low to high stage periods.



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885 Figure 9: Temporal variations in inverted resistivity models for (a) pool and (b) riffle. Black dots represent unfrozen
 886 conditions, grey dots indicate partially frozen conditions, while white dots indicate completely frozen river conditions.
 887 Select resistivity models (a-h) along the time series are shown in Fig. 10.

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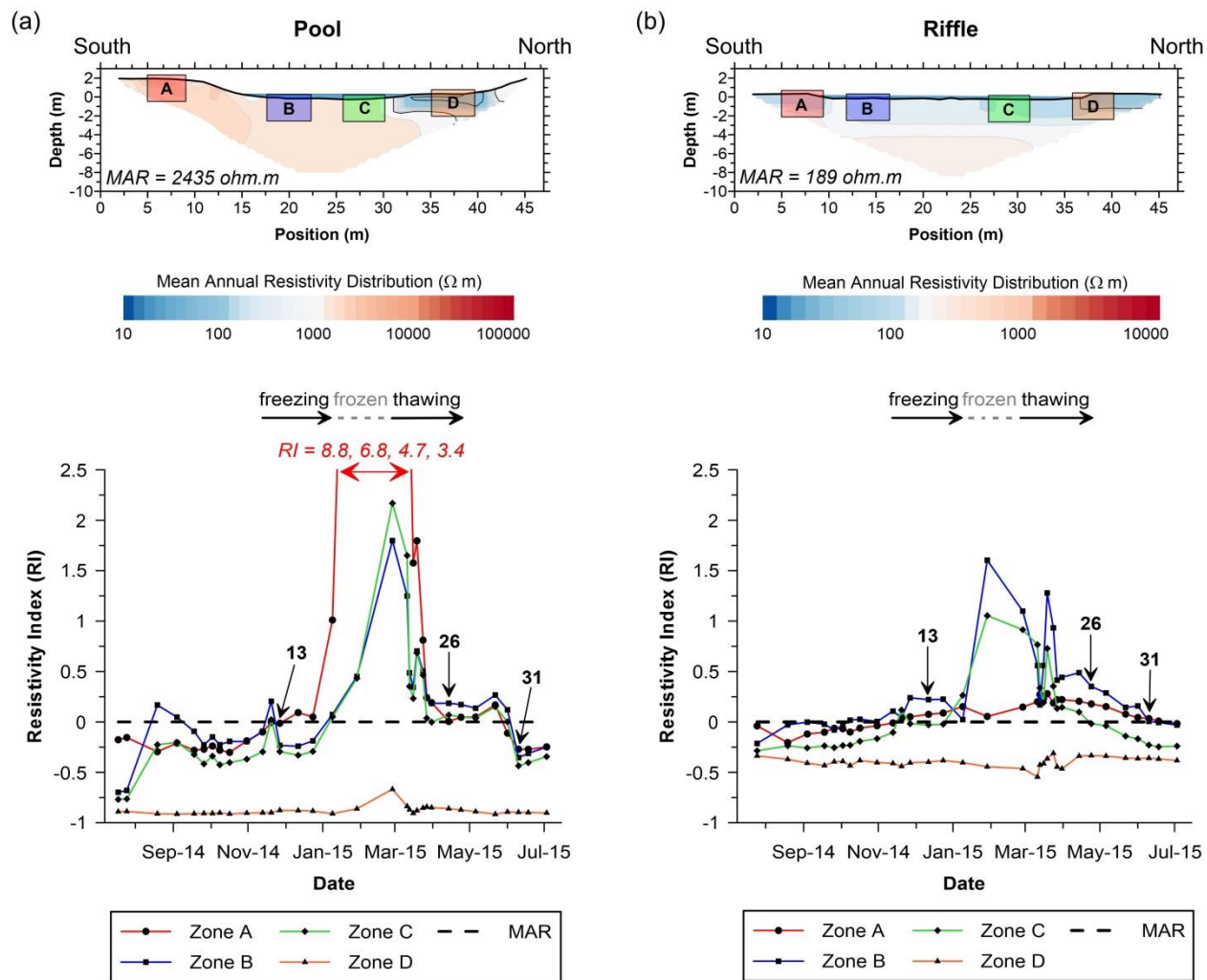


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890 Figure 10: Representative inverse resistivity models across the pool and riffle oriented from south to north.
 891 Datasets (a-h) are identified in Fig. 9. River stage (w.l.) and surface water resistivity (w.r.) values were fixed in the
 892 inverse model. A marked increase in resistivity was observed beneath the river during colder seasonal conditions
 893 (November through March), while lower resistivities were observed during warmer seasonal conditions (July).

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900 Figure 11: Spatiotemporal fluctuations in resistivity within the focused monitoring zones A, B, C and D. The
 901 resistivity index (RI) was calculated using Eq. (3), using the mean zone resistivity (MZR) for a given measurement
 902 date and the mean annual resistivity (MAR) of the whole profile.