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### **Paper:**

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# Bulk meltwater flow and liquid water content of snowpacks mapped using the electrical self-potential (SP) method

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**Abstract.** Our ability to measure, quantify and assimilate hydrological properties and processes of snow in operational models is disproportionately poor compared to the significance of seasonal snowmelt as a global water resource and major risk factor in flood and avalanche forecasting. We show here that strong electrical self-potential fields are generated in melting in situ snowpacks at Rhone Glacier and Jungfrauoch Glacier, Switzerland. In agreement with theory, the diurnal evolution of self-potential magnitudes ( $\sim 60\text{--}250\text{ mV}$ ) relates to those of bulk meltwater fluxes ( $0\text{--}1.2 \times 10^{-6}\text{ m}^3\text{ s}^{-1}$ ) principally through the permeability and the content, electrical conductivity and pH of liquid water. Previous work revealed that when fresh snow melts, ions are eluted in sequence and electrical conductivity, pH and self-potential data change diagnostically. Our snowpacks had experienced earlier stages of melt, and complementary snow pit measurements revealed that electrical conductivity ( $\sim 1\text{--}5 \times 10^{-6}\text{ S m}^{-1}$ ) and pH ( $\sim 6.5\text{--}6.7$ ) as well as permeabilities (respectively  $\sim 9.7 \times 10^{-5}$  and  $\sim 4.3 \times 10^{-5}\text{ m}^2$  at Rhone Glacier and Jungfrauoch Glacier) were invariant. This implies, first, that preferential elution of ions was complete and, second, that our self-potential measurements reflect daily changes in liquid water contents. These were calculated to increase within the pendular regime from  $\sim 1$  to 5 and  $\sim 3$  to 5.5 % respectively at Rhone Glacier and Jungfrauoch Glacier, as confirmed by ground truth measurements. We conclude that the electrical self-potential method is a promising snow and firn hydrology sensor owing to its suitability for (1) sensing lateral and vertical liquid water flows

directly and minimally invasively, (2) complementing established observational programs through multidimensional spatial mapping of meltwater fluxes or liquid water content and (3) monitoring autonomously at a low cost. Future work should focus on the development of self-potential sensor arrays compatible with existing weather and snow monitoring technology and observational programs, and the integration of self-potential data into analytical frameworks.

## 1 Introduction

More than a sixth of the world's population relies on melt from seasonal snow and glaciers for water supply (Barnett et al., 2005). Snow, and runoff from snow, are also major resources for the hydroelectric, tourism and inland fishery industries, and furthermore represent hazards from flooding and avalanches (Mitterer et al., 2011). The availability of snow models constrained by a reliable observational basis, for the forecasting of snow hydrological properties and processes in climate, resource and hazard applications is therefore of considerable socio-economic significance (Wever et al., 2014). However, the parameterisation of fundamental snow hydrological attributes, such as liquid water content and flux, is a well-recognised major source of uncertainty in operational models used in snow and hydrological forecasting (Livneh et al., 2010; Essery et al., 2013). This uncertainty in operational models is rooted principally in the inability of traditional snow hydrological techniques to provide au-

tomated attribute measurements non-invasively and on spatial scales that match those used in operational snow models. Relevant traditional techniques include dielectric (Denoth, 1994) or “hand” tests (Fierz et al., 2009) of snow liquid water contents, lysimeter measurements of discharge, temperature and pH and electrical conductivity of bulk meltwaters (Campbell et al., 2006; Williams et al., 2010) and manual observation or measurement of snow density and grain size (Fierz et al., 2009). Even cutting-edge upward-looking radar measurements of snowpack structure and liquid water content (Heilig et al., 2010; Mitterer et al., 2011; Schmid et al., 2014) compare unfavourably with model predictions of wetting front propagation (Wever et al., 2014), attributed to inherent limitations of the 1-D approach in capturing preferential flow.

By combining field measurements with a theory and model of self-potential signals associated with unsaturated flow in melting snow (Kulesa et al., 2012), we show here that electrical self-potential geophysical data integrated with traditional snow measurements can address these limitations. The self-potential technique is a passive geo-electrical method that exploits the presence of naturally occurring electrical potentials in the subsurface generated as a result of dipolar charge separation when water flows through a porous matrix (“streaming potential”; Darnet et al., 2003; Revil et al., 2006). The self-potential method has a unique ability in delineating, monitoring and quantifying the flow of subsurface water in groundwater aquifers and unsaturated media (e.g. Revil et al., 2006, and references therein), and for application to numerous cold regions (e.g. French et al., 2006; Kulesa, 2007, and references therein). This ability is due to the fact that porewaters generally have an excess of electrical charge due to the electrical double layer at the interface between the solid matrix (in this case snow grains) and porewater. The advective drag of this excess of electrical charge is responsible for a streaming current, whose divergence generates a quasi-static electric field known as the streaming potential (Sill, 1983; Revil et al., 2003). More recently, streaming potential theory has been extended for unsaturated conditions (Linde et al., 2007; Revil et al., 2007; Jougnot et al., 2012). A new theory and numerical model of self-potential signals associated with unsaturated flow in melting snow, along with laboratory tests, strongly promoted the technique as a non-intrusive hydrological sensor of water fluxes (Kulesa et al., 2012) at spatial scales intermediate between snow pits and satellite footprints or, given independent flux measurements, of evolving physical and chemical properties of snow and snowmelt.

We answer two fundamental questions. (1) Can the self-potential method serve as a non-intrusive field sensor of temporally evolving bulk meltwater fluxes and liquid water contents of snow? (2) What are the ambiguities introduced into estimates of liquid water contents from self-potential and bulk discharge data, by uncertainties inherent in the governing snow physical and chemical properties? Lastly we dis-

cuss the implications and possibilities of the technique for future snow measurement and modelling research and practice. Our study thus takes a significant step towards the in situ implementation of the self-potential method for improved characterisation and monitoring of snow liquid water contents and meltwater fluxes.

## 2 Theory, field sites and methods

The Poisson equation relates the electrical field  $\psi$  to the source current density in a partially or fully saturated snowpack:

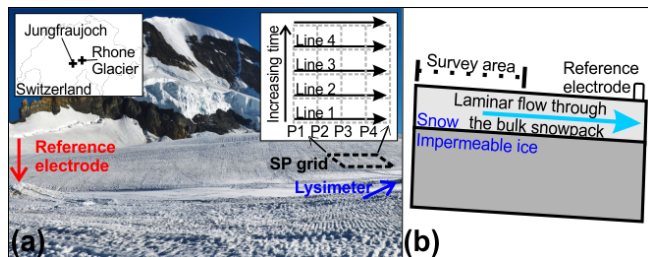
$$\nabla \cdot (\sigma \nabla \psi) = \nabla \cdot j_s, \quad (1)$$

where  $\sigma$  is the bulk electrical conductivity of the porous material (in  $\text{S m}^{-1}$ ), and  $j_s$  is the source current density (in  $\text{A m}^{-2}$ ). Equation (1) applies only in the low-frequency limit of the Maxwell’s equations without external injection or retrieval of charges, or charge storage in the snowpack. Extending the classic Helmholtz–Smoluchowski theory for unsaturated flow in snow, the 1-D solution to Eq. (1) is given by

$$\psi_m - \psi_0 = -\frac{\varepsilon \zeta}{\eta \sigma_w} S_w (H_m - H_0), \quad (2)$$

where  $\psi_m$  and  $H_m$  are respectively the electrical and hydraulic potentials at the measurement electrode,  $\psi_0$  and  $H_0$  are the corresponding potentials at the reference electrode,  $\zeta$  is the zeta potential (V), and  $\varepsilon$ ,  $\eta$ ,  $\sigma_w$  and  $S_w$  are respectively the dielectric permittivity ( $\text{F m}^{-1}$ ), dynamic viscosity (in Pa s), electrical conductivity ( $\text{S m}^{-1}$ ) and relative saturation (dimensionless) of the melt- or rainwaters in the snowpack’s pore space (Kulesa et al., 2012). The zeta potential is the voltage across the electrical double layer at the interface between the ice matrix and the porewaters, as controlled by these constituents’ physical and electrical properties.

To address the specific objectives set out in the introduction through data-driven testing, we developed an experimental survey design to simulate the geometry of the laboratory snow column of Kulesa et al. (2012) (Fig. 1b). It was therefore our aim to characterise bulk meltwater fluxes in inclined snowpacks at two glaciers in Valais, Switzerland, measuring all relevant snowpack attributes for ground truth. Self-potential and traditional snow hydrological measurements were acquired on 13, 14 and 15 June 2013 from the ablation area snowpack at Rhone Glacier, and on 5 September 2013 from the glacial accumulation area at Jungfrauoch Glacier (Fig. 1a). At Rhone Glacier and Jungfrauoch Glacier site elevations were respectively 2340 and 3460 m a.s.l., with surface gradients of  $\sim 8$  and  $17^\circ$ . At the Rhone Glacier all 3 days experienced comparable air temperature, although 15 June was noticeably cloudier with a very low sunshine duration. Because daily average temperatures were between 5 and  $15^\circ\text{C}$  with no fresh snowfall (MeteoSuisse), the snowpacks



**Figure 1.** (a) Example survey set-up, self-potential grid  $25 \times 25$  m. The inset on the left shows the location of both field sites. The inset on the right illustrates the self-potential survey design; to provide each self-potential data value, a profile of 25 data points (P1, P2, etc.) was collected (Line 1, Line 2, etc.), perpendicular to assumed bulk water flow. (b) Schematic of the self-potential experiment developed by Kulesa et al. (2012) for the situ snowpack surveys.

would have experienced significant melting in the weeks before the surveys. We therefore expect them to be physically mature in terms of enhanced grain size and density due to metamorphosis, and chemically mature in terms of invariant meltwater pH and electrical conductivity as preferential elution of solutes has been completed (Kulesa et al., 2012, and references therein).

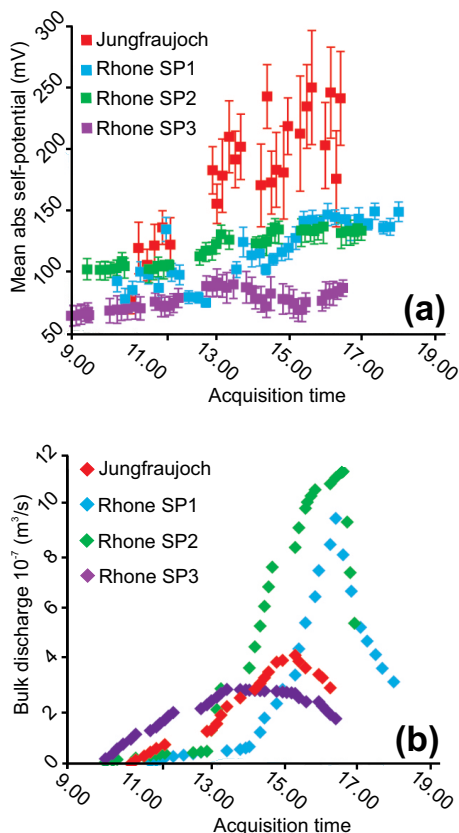
At both sites, more than 100 self-potential measurements were made at the snow surface, and meltwater bulk discharge in a lysimeter, pH and electrical conductivity, and snowpack characteristics including thickness, density, grain size and liquid water content were recorded. Adopting our established acquisition procedures (Thompson et al., 2012), we conducted all self-potential surveys using a pair of lead/lead chloride “Petiau” non-polarising electrodes (Petiau, 2000). The survey was carried out following the potential amplitude method (Corry et al., 1983); this employs a reference electrode in a fixed location and a roving electrode which is moved through the survey area at 0.5 m intervals (Fig. 1a). Self-potential surveys were conducted in profiles of 25 data points perpendicular to the principal direction of water flow, where the latter was assumed to follow the gradient indicated by snow surface topography. All self-potential measurements were taken as differential readings relative to the reference electrode, minimising streaming, electrochemical and thermal potentials at the latter by grounding them outside the survey areas at the top of a local topographic high point (Fig. 1a), submerged in a glass jar, open at the top and filled with water-saturated local media (Kulesa et al., 2003a). The jar was then buried upright  $\sim 1$  m deep to avoid exposure to surface temperature variations. Surveys were carried out with a fixed tie-in point (measured every second line) at the reference electrode, allowing for correction of the effects of electrode polarisation and drift (Doherty et al., 2010; Thompson et al., 2012).

Bulk discharge through a snowpack is preferably measured with a lysimeter (Campbell et al., 2006; Williams et

al., 2010), in this case made up of a series of smaller (guttering) areas joined together to prevent freezing and compaction (after Campbell et al., 2006). The lysimeter was placed at the base of Rhone Glacier’s snowpack, and at the limit of the diurnal melt penetration depth at Jungfrauoch Glacier (determined by daily dye-tracing experiments). Snow density (by balance) and average snow grain size (crystal card and lens) were measured, at the start and end of each self-potential survey to reveal any intermittent snow metamorphosis, using standardised techniques within the top and basal layers of snow pits freshly excavated at the survey sites (Fierz et al., 2009). Liquid water content was estimated using two different techniques, including the hand test (Colbeck et al., 1990; Fierz et al., 2009) in the surface and base layers of Rhone Glacier’s snow pit, and the Denoth capacitance meter (Denoth, 1994) in the surface and base layers of the snow pit at Jungfrauoch Glacier. The latter measurements were acquired across a 2-D grid where the instrument was inserted into the snowpack at a depth of 0.4 m following the same survey spacing as the self-potential measurements.

### 3 Field measurement results

The drift-corrected self-potential magnitudes and meltwater bulk discharges both increase with time through the day until a peak in late afternoon, after which they both begin to decrease (Fig. 2). There is no distinguishable time lag between the measured self-potential magnitude and discharge data (Fig. 2), and the ratio between self-potential and bulk discharge changes consistently over time (Fig. 3). Days 1 and 2 at Rhone Glacier were characterised by higher discharges and self-potential magnitudes compared to day 3, and intriguingly bulk discharge at Jungfrauoch Glacier was akin to day 3 at Rhone Glacier, but self-potential magnitudes at Jungfrauoch Glacier were much higher than days 1 and 2 at Rhone Glacier (Fig. 2). The pH, electrical conductivity and temperature of meltwater, recorded with each bulk discharge measurement, show no consistent temporal or spatial variation across any of the four field surveys. Fluid electrical conductivity values generally ranged between  $1 \times 10^{-6}$  and  $5 \times 10^{-6} \text{ S m}^{-1}$  without spatial or temporal consistency, while pH ranged between 6.5 and 6.9. Snow grain size remained constant at  $\sim 1.5$  mm at Rhone Glacier and  $\sim 1$  mm at Jungfrauoch Glacier, while snow densities ranged between  $555$  and  $573 \text{ kg m}^{-3}$  without spatial or temporal consistency. The very small variability range of the snowpack characteristics measured is consistent with mature snowpacks, as assumed above with reference to prior meteorological conditions. At Rhone Glacier the liquid water content of snow had a wetness index of 3 irrespective of measurement time or location at the surface or base of the snow pit, associated with a liquid water content range of 3–8 % vol. (Colbeck et al., 1990). At Jungfrauoch Glacier liquid water content, measured using the Denoth meter, gave profile-averaged val-

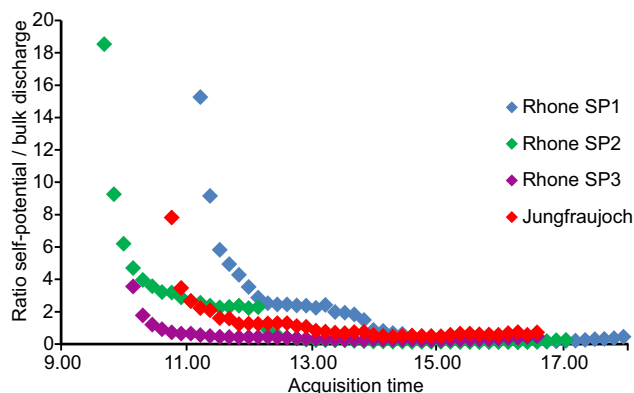


**Figure 2.** Time series of (a) bulk self-potential measurements and (b) bulk discharge measurements for the three Rhone Glacier surveys and the Jungfrauoch Glacier survey. Each self-potential data point represents the mean value of a profile (consisting of 25 data points); the error bars illustrate the variability over each profile. Bulk discharge was measured over each profile by the lysimeter.

ues of 1.5 to  $\sim 5.0$  % vol., increasing consistently throughout the survey period. These measurements and inherent uncertainties are used below for snow liquid water content calculations, uncertainty analysis and sensitivity testing.

#### 4 Objective 1: self-potential as a snow hydrological sensor

Both survey areas were south facing, topographically inclined but otherwise had no visibly distinguished snow surface undulations, and any snow thickness variations were minimal. We therefore expect changes in self-potential magnitudes to be pronounced in the downslope direction, and minimal across-slope along any individual profile (Fig. 1a). Averaging all 25 self-potential data points acquired along any particular profile, a 1-D upslope–downslope series of self-potential magnitudes is produced for a given survey area on a given day, together with uncertainty estimates reflecting natural spatial and temporal variability along the profile (Ta-



**Figure 3.** Ratio between self-potential magnitudes (V) and bulk discharge ( $\text{m}^3 \text{s}^{-1}$ ) for each of the four surveys through time, illustrating that the ratio changes consistently over time.

ble 1). For each profile, the acquisition time of the central data point was assigned to it, and all measurements of snowpack and meltwater properties were averaged over the same time period ( $\sim 20$  min, i.e. the acquisition time of any one self-potential profile). The upslope–downslope series of average self-potential magnitudes thus emulates measurements along a horizontally inclined version of the 1-D snow column used in Kulesa et al. (2012) (Fig. 1b). These authors reformulated the 1-D solution in Eq. (2) to relate measured self-potential magnitudes and bulk discharges through their partially saturated snow column, which we can therefore adapt here to our field experiment.

This adaptation is dependent on four key assumptions, including that (1) water flow within the survey areas' snowpacks is laminar and homogenous in three dimensions, where the snowpack surface and base have constant and equal inclination and thus maintain a spatially constant hydraulic gradient; (2) all contributions to the measured self-potential signal from flow below the base of the snowpack, runoff at the surface of the snowpack and flow outside the lateral boundaries of the survey areas' snowpacks are negligible, and all water contributing to the measured self-potential signals is adequately captured by our bulk discharge measurements; (3) all snow physical and chemical properties controlling the self-potential magnitude do not vary spatially across the survey areas' snowpacks, so that our ground-truth snowpit data apply uniformly across them, and (4) any spatial changes in self-potential magnitudes are dominated by temporal changes in snow or meltwater properties, while static-elevation-driven spatial changes are negligible. We assess the implications of any potential violations to these assumptions in Sect. 6.

At a given time,  $t_n$ , the measured self-potential field,  $\Psi_m(t_n)$ , in our survey area is the difference between the locally produced self-potential field,  $\Psi_1(t_n)$ , and the self-potential field at the reference electrode,  $\Psi_0(t_n)$ . The lat-

ter is unknown in our field feasibility study, although our method of emplacing the reference electrode is elaborate and designed to eliminate, or at least minimise, any streaming potentials at the reference electrode (see Sect. 2). Once the reference electrodes have settled in their environments, we further expect any electrochemical or thermal potentials to be negligible. We can therefore expect  $\Psi_0(t_n)$  to be close to zero, but nonetheless apply caution and take a two-step approach. Initially we eliminate the reference self-potential fields by considering temporal changes in measured self-potential magnitudes before, subsequently, considering absolute self-potential magnitudes.

#### 4.1 Temporal changes in self-potential magnitudes

We can eliminate the reference field by differencing two self-potential measurements acquired at two successive times:

$$\psi_m(t_n) - \psi_m(t_{n-1}) = \Psi_1(t_n) - \Psi_1(t_{n-1}). \quad (3)$$

Equation (3) assumes that  $\Psi_0$  and  $H_0$  are temporally invariant, a reasonable supposition for drift-corrected self-potential data if the reference electrode is correctly emplaced. Recognising that  $\psi_0 = H_0 \approx 0$  for their snow column experiment, Kulesa et al. (2012) reformulated Eq. (2) to show that the self-potential field at a measurement electrode,  $\Psi_1(t_n)$ , can be approximated by

$$\Psi_1(t_n) = \frac{\varepsilon \zeta}{\sigma_w} \frac{S_w(t_n)}{S_e^n(t_n)} \frac{1}{kA} Q(t_n), \quad (4)$$

where  $Q$  ( $\text{m}^3 \text{s}^{-1}$ ) is bulk discharge in the snowpack through the cross-sectional area  $A$  ( $\text{m}^2$ ),  $k$  is permeability,  $S_e$  is effective saturation and  $n \approx 3.3$  is the saturation exponent (after Albert and Krajcicki, 1998; Kulesa et al., 2012). Assuming that any temporal changes in the self-potential field at the reference electrodes in our field experiments are negligible, the difference between successive field self-potential measurements in time can be approximated by

$$\psi_m(t_n) - \psi_m(t_{n-1}) = \frac{\varepsilon \zeta}{\sigma_w} \frac{1}{kA} \left( \frac{S_w(t_n)}{S_e^n(t_n)} Q(t_n) - \frac{S_w(t_{n-1})}{S_e^n(t_{n-1})} Q(t_{n-1}) \right). \quad (5)$$

In the present case we have measured  $\psi_m(t_n)$  and  $\Psi_m(t_{n-1})$  as well as  $Q(t_n)$  and  $Q(t_{n-1})$ . We have also measured, or can estimate from well-established empirical relationships, all other parameters coupling the temporal difference in self-potential fields ( $\Psi_m(t_n)$  and  $\psi_m(t_{n-1})$ ) to that of discharge (expression in the large parentheses on the right-hand side of Eq. 5). To demonstrate the usefulness of self-potential measurements in snow research and practice, we can therefore evaluate Eq. (5) at successive times,  $t_n$  and  $t_{n-1}$ , to calculate temporal changes in the liquid water content,  $S_w$ , of the snowpacks at our field sites. This evaluation is subject

to assumptions (1) to (4) above, and is ground-truthed using snow pit measurements of liquid water contents.

At both Rhone Glacier and Jungfrauoch Glacier, self-potential magnitude ( $\psi_m$ ), bulk discharge ( $Q$ ), electrical conductivity ( $\sigma_w$ ) and cross-sectional area ( $A$ ) (survey area width  $\times$  snow depth) were measured directly. Assuming that water at  $0^\circ\text{C}$  has a dielectric permittivity of  $\varepsilon_r = 88$ , the dielectric permittivity ( $\text{F m}^{-1}$ ) of pore meltwater is  $\varepsilon = \varepsilon_r \varepsilon_0 = 7.8 \times 10^{-9} \text{ F m}^{-1}$ , where  $\varepsilon_0 = 8.85 \times 10^{-12} \text{ F m}^{-1}$  is the dielectric permittivity of vacuum. Permeability ( $k$ ) can be derived from our snow density ( $\rho_s$ ) and grain size ( $d$ ) measurements using Shimizu's (1970) empirical relationship:

$$k = 0.077 d^2 e^{-0.0078 \rho_s}, \quad (6)$$

where  $k$  is in  $\text{m}^2$ ,  $d$  is in  $\text{m}$  and  $\rho_s$  is in  $\text{kg m}^{-3}$ . This commonly used equation was derived from a fit to laboratory data collected with small rounded grains and a starting grain diameter of  $\sim 0.33 \text{ mm}$  (Shimizu, 1970). However, later work ascertained experimentally that Shimizu's (1970) empirical formula does in fact apply to a much larger range of grain diameters, as expected to be encountered in practice (less than  $0.33 \text{ mm}$  to greater than  $2 \text{ mm}$ ) (Jordan et al., 1999). We can therefore expect Eq. (6) to be robust for our purposes. Effective saturation ( $S_e$ ) and  $S_w$  are related through the irreducible water saturation  $S_w^{\text{ir}}$  by

$$S_e = \frac{S_w - S_w^{\text{ir}}}{1 - S_w^{\text{ir}}}. \quad (7)$$

In the absence of direct measurements, we adopt the commonly used values of  $S_w^{\text{ir}} = 0.03$  and  $n \approx 3.3$  (Kulesa et al., 2012), and assume that these values are invariant in space and time at our study sites.

A significant challenge arises however in that there is one remaining parameter, the zeta potential ( $\zeta$ ), which is unknown here and poorly constrained in general. Earlier work on artificial ice samples, of fixed bulk electrical conductivity, ascertained that the zeta potential reverses sign from  $\sim +0.01$  to  $\sim -0.02 \text{ V}$  as equilibrium pH increases from less than 3 to greater than 8 (Drzymala et al., 1999; Kallay et al., 2003). The electrochemical properties of the electrical double layer at the snow grain surfaces, and thus also the magnitude and potentially the sign of the zeta potential, will change over time in a fresh snowpack as the snow is affected by melt, recrystallisation and the preferential elution of ions (Meyer and Wania, 2008; Meyer et al., 2009; Williams et al., 1999). Recent "natural snowmelt" laboratory experiments were consistent with a progressive increase of pH from 4.3 to 6.3 and a simultaneous decrease in electrical conductivity from  $\sim 1 \times 10^{-1}$  to  $\sim 6 \times 10^{-7} \text{ S m}^{-1}$ , as the elution of ions follows a well-known sequence (Kulesa et al., 2012). Upon conclusion of the laboratory experiments of Kulesa et al. (2012), modelled rates of change of pH and electrical conductivity were minimal and the snow column mature. The zeta potential is principally a function of pH and

electrical conductivity, and the combined dependency of the zeta potential on electrical conductivity ( $\sigma_w$ ), meltwater pH ( $\text{pH}_w$ ) and the meltwater pH at the point of zero charge ( $\text{pzc}$ ) can be expressed as

$$\zeta(\sigma_w, \text{pH}) = [\alpha + \beta \log_{10} \sigma_w] \left( \sin \frac{\pi}{12} [\text{pH}_w - \text{pH}_w(\text{pzc})] \right), \quad (8)$$

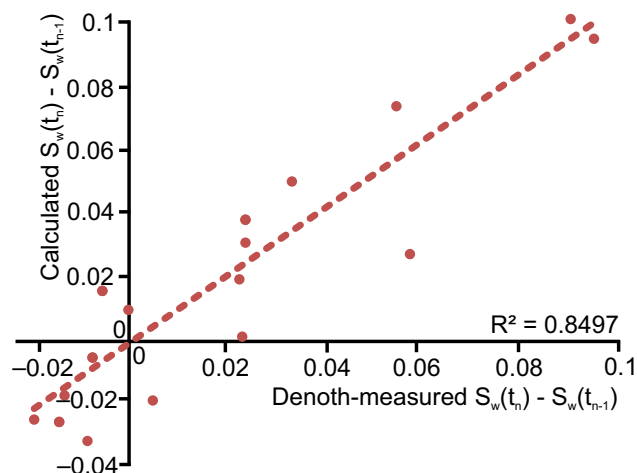
where  $\alpha$  and  $\beta$  depend on the chemical composition of the pore fluid and can be determined empirically (Revil et al., 1999). Kulesa et al. (2012) inferred the zeta potential changed from  $\sim -7.5 \times 10^{-2}$  V at the start of the natural snowmelt experiments to  $+1.5 \times 10^{-2}$  V at the end, when the rate of change of the zeta potential was minimal.

The final values of pH and electrical conductivity that Kulesa et al. (2012) calculated from Eq. 8 were similar to those measured at Rhone Glacier and Jungfrauoch Glacier (respectively  $\sim 6.5$ – $6.9$  and  $\sim 1$ – $5 \times 10^{-6}$  S  $\text{m}^{-1}$ ), suggesting that these in situ snowpacks were likewise mature as expected (Sect. 2). This inference is corroborated by the absence of consistent spatial or temporal changes in either pH or electrical conductivity throughout the survey periods. In the laboratory study of Kulesa et al. (2012), the pH-corrected zeta potential had values around zero for the range of electrical conductivities ( $1$ – $5 \times 10^{-6}$  S  $\text{m}^{-1}$ ) measured at Rhone Glacier and Jungfrauoch Glacier ( $1$ – $5 \times 10^{-6}$  S  $\text{m}^{-1}$ ), and its rate of change became minimal along with those of pH and electrical conductivity. We can therefore expect a small and invariant zeta potential value to apply to the snowpacks at Rhone Glacier and Jungfrauoch Glacier. Indeed, an excellent fit ( $R^2 \approx 0.85$ ) between liquid water contents measured at Jungfrauoch Glacier with the Denoth meter and that calculated based on Eq. (5) is obtained when the zeta potential is assigned a value of  $\sim -1 \times 10^{-5}$  V (Fig. 4). This excellent fit suggests that in situ measurements or empirically derived estimates of the parameters affecting coupling between measured self-potential magnitudes and discharges in Eq. (5) are robust for practical purposes.

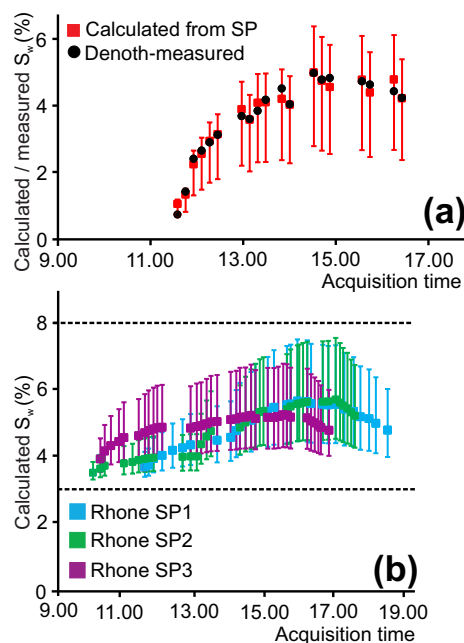
#### 4.2 Absolute changes in self-potential magnitudes

The same parameters affect the coupling between temporal changes in self-potential magnitudes and discharge (Eq. 5), and absolute changes therein as described by Eq. (4) derived by assuming that the reference potential is zero. We are therefore encouraged to calculate absolute liquid water contents from our self-potential data using Eq. (4). We do this initially for Jungfrauoch Glacier because here we have detailed ground-truth measurements of liquid water content made with a Denoth meter. Encouragingly we find that calculated and measured ground-truth data match each other very well (Fig. 5a), attesting to the fact that the reference potentials at Jungfrauoch Glacier may not only be temporally invariant as confirmed earlier, but generally have negligible magnitudes.

We can apply the same expectation of negligible reference self-potential magnitudes to our surveys at Rhone Glacier



**Figure 4.** Temporal differences in  $S_w$  inferred from self-potential data against temporal differences in the Denoth-measured  $S_w$  at Jungfrauoch Glacier, according to Eq. (5).



**Figure 5.** (a) Liquid water content calculated from Eq. (4) for the self-potential survey carried out at Jungfrauoch Glacier, with the corresponding Denoth measurements. The uncertainty range illustrates the minimum and maximum model results for the range of parameters (Table 1). (b) Liquid water content calculated from Eq. (4) for each of the three self-potential surveys carried out at Rhone Glacier. All results are within the range of liquid water content (% vol) estimated by the hand tests (black dashed lines).

on the 3 successive days. We find that absolute liquid water contents inferred from Eq. (4) generally fall well within the range of  $\sim 3$ – $8$  % inferred from our ground-truth hand tests. We can therefore conclude that given careful emplacement of the reference electrode, the simple empirical relationship be-

**Table 1.** Best estimate of each parameter for Rhone Glacier SP (day 2) and relative assumed uncertainty and sensitivity ranges. The sensitivity ranges are based on the measurement accuracy of each measured parameter or the confidence of estimates parameters. The uncertainty ranges are exaggerated from the sensitivity values to highlight the effect of poor measurement or estimation.

Measured/estimated parameters	Best estimate	Sensitivity range	Uncertainty range
Self-potential $\psi_m$ (V)	Variable	$\psi_m \pm 20\%$	$\psi_m \pm 40\%$
Discharge $Q$ ( $\text{m}^3 \text{s}^{-1}$ )	Variable	$Q \pm 20\%$	$Q \pm 40\%$
Electrical conductivity $\sigma_w$ ( $\text{S m}^{-1}$ )	$5 \times 10^{-6}$	$\sigma_w \pm 5 \times 10^{-7}$	$10^{-7}$ – $10^{-4}$
Zeta potential $\zeta$ (V)	$-1 \times 10^{-5}$	$\zeta \pm 50\%$	$10^{-4}$ – $10^{-6}$
Permeability from grain diameter $d$ (m)	0.00175	$d \pm 0.0005$	$d \pm 0.001$
density $\rho$ ( $\text{kg m}^3$ )	555.5	$\rho \pm 70$	$\rho \pm 140$
Cross-sectional area from width $w$ (m)	12.5	$w \pm 5$	$w \pm 10$
depth $dp$ (m)	1.45	$dp \pm 0.2$	$dp \pm 1$

tween self-potential magnitudes, discharge and liquid water content is robust not only in a laboratory setting (Kulesa et al., 2012), but also for application to in situ snowpacks. The self-potential method therefore shows considerable promise as a non-intrusive snow hydrological sensor.

## 5 Objective 2: self-potential sensitivity to uncertainty in snow properties

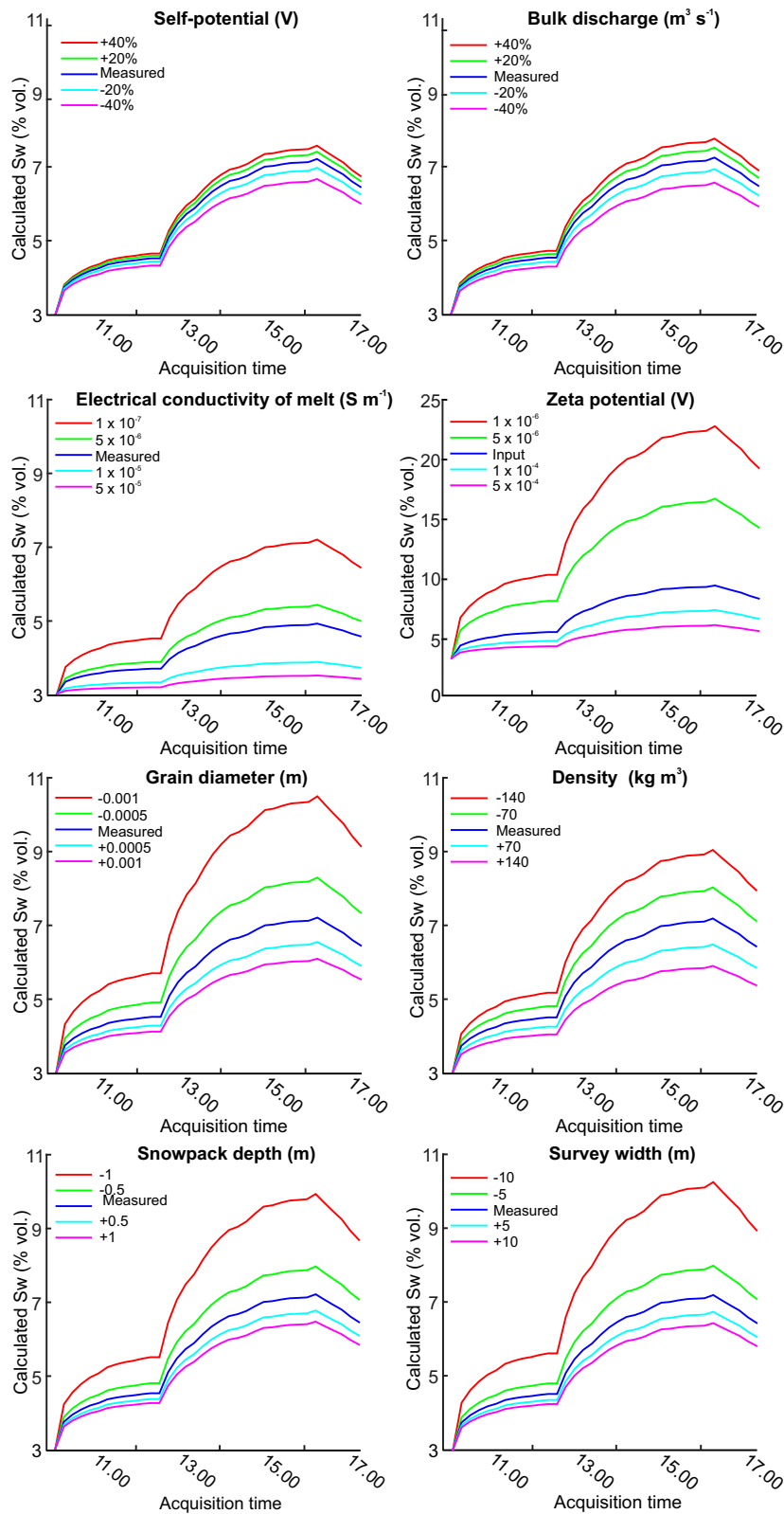
We evaluate the sensitivity of calculated liquid water contents to both individual and combined parameter uncertainties. For each parameter a range of uncertainty values was created, with the respective minima and maxima approximately twice that of the uncertainty (Table 1). Repeat water content calculations were carried out initially by changing each parameter individually for a range of values between the respective minima and maxima. The results cluster broadly in three categories, including the zeta potential (up to  $\sim 20\%$  change in liquid water content within the 50 % uncertainty range), followed by grain diameter, survey area width, electrical conductivity, snow depth and snow density ( $\sim 3$ – $4\%$  change) and bulk discharge and self-potential (2 % change) (Fig. 6). These three categories readily reflect our knowledge of or ability to measure the respective parameters in situ, with surprisingly low sensitivity to cross-sectional area despite our simplistic calculation and significant inherent assumptions (i.e. 1–4 in Sect. 4). Self-potential magnitudes are readily measured in the field with minimum uncertainty (Fig. 6), although the strongly enhanced sensitivity to the zeta potential highlights the need for focused research to tightly constrain possible values of this parameter in in situ snowpacks.

While this gives a good indication of the parameters to which water content calculations are most sensitive, it does not indicate possible feedbacks between parameters. Feedbacks were therefore evaluated by calculating liquid water contents for all possible combinations of the best estimates and minimum and maximum parameter values (Table 1), giv-

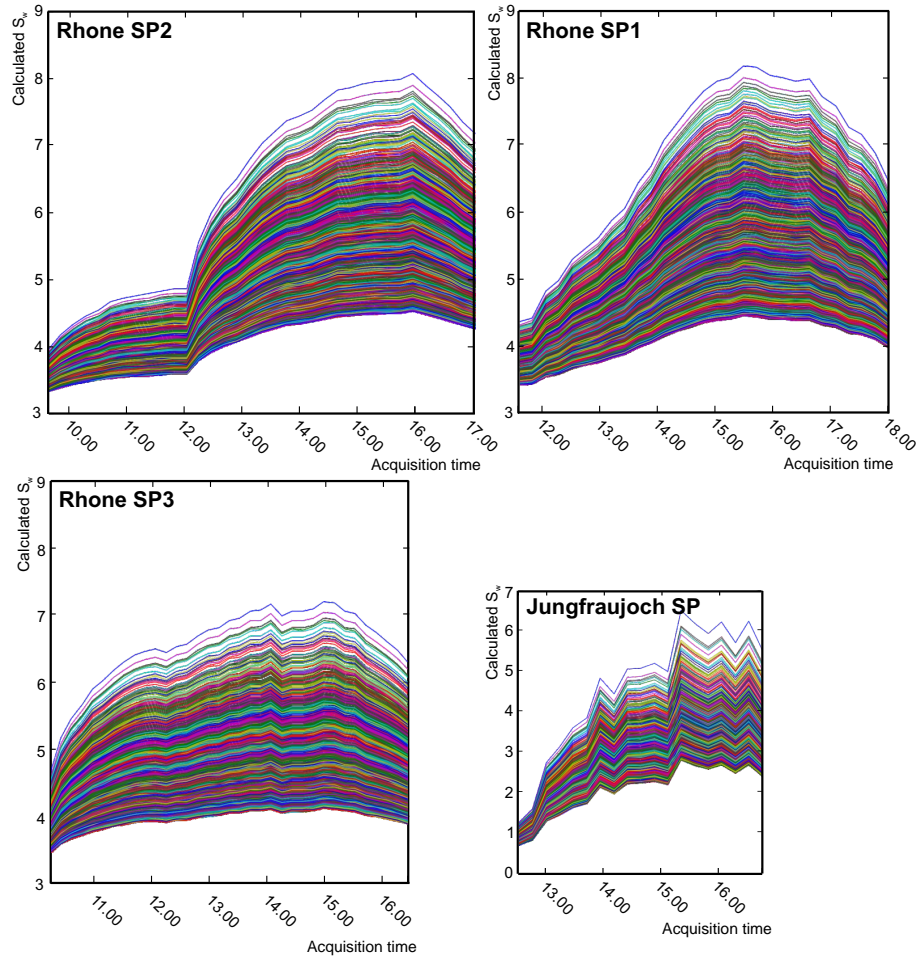
ing over 6500 solutions (Fig. 7). The minimum and maximum outputs were then adopted as the lower and upper uncertainty bounds (Fig. 3). Due to the large potential uncertainty in the zeta potential, the sensitivity range was arbitrarily set to  $\pm 50\%$  for illustrative purposes (Sect. 4).

Despite our consideration of extreme potential error bounds, calculated uncertainties in liquid water contents are restricted to a relatively small range ( $\sim 20\%$  for large assumed uncertainty in the zeta potential, and  $\sim 3$ – $4\%$  otherwise) at both Rhone Glacier and Jungfrauoch Glacier, and absolute values remain within the pendular regime where water bodies in the pore space remain isolated. At the latter site the daily evolution of liquid water contents thus is well captured even if uncertainty is taken into account (Fig. 5b), and likewise at Rhone Glacier, calculated liquid water contents plus uncertainties still fall within the range of field measurements (Fig. 5a). Our inferences thus not only support the notion of Kulesa et al. (2012) that existing snow hydrological relationships are robust for modelling purposes, but also suggest that they may apply to in situ field surveys. These inferences can also provide an explanation for the relatively large self-potential magnitudes generated by relatively low bulk discharge at Jungfrauoch Glacier (Fig. 2). Because we did not observe or infer any consistent or statistically significant differences between Rhone Glacier and Jungfrauoch Glacier in dielectric permittivity ( $\epsilon$ ), zeta potential ( $\zeta$ ), saturation ( $S_w S_e^{-n}$ ), electrical conductivity ( $\sigma_w$ ) or cross-sectional area ( $A$ ), the only remaining parameter that could facilitate the observed relative difference is permeability ( $k$ ). Indeed, using an average snow density of  $564 \text{ kg m}^{-3}$ , the differences in mean snow grain sizes between Rhone Glacier ( $1.5 \times 10^{-3} \text{ m}$ ) and Jungfrauoch Glacier ( $1 \times 10^{-3} \text{ m}$ ) translate into respective permeabilities of  $9.7 \times 10^{-5}$  and  $4.3 \times 10^{-5} \text{ m}^2$ . The relatively reduced permeability of Jungfrauoch Glacier's accumulation-area snowpack therefore likely supported the presence of self-potential magnitudes that were markedly elevated relative to Rhone Glacier's ablation-area snowpack (Eq. 4).





**Figure 6.**  $S_w$  calculations for a range of values for each parameter. In each case the range is an exaggerated uncertainty range (Table 1), highlighting the effect of each individual parameter on the calculated  $S_w$  output, using Rhone Glacier SP day 2 as an example.



**Figure 7.** Full sensitivity analysis for each of the four data sets. Each graph shows the full range of calculated liquid water content ( $S_w$ ) values of every combination of minimum, best estimate and maximum for each of the parameters.

This inference emphasises the sensitivity of the self-potential method to permeability as a fundamental snow hydrological property, along with its observed sensitivity to bulk meltwater discharge and inferred sensitivity to liquid water content.

## 6 Synthesis and conclusions

The ability of the electrical self-potential method to sense meltwater flow in in situ snowpacks is unique, where self-potential magnitudes scale directly with discharge and are zero in the absence of flow. The scaling factor (right side of Eq. 4) depends principally on the liquid water content of the snowpack, its permeability and the water chemistry (Kulesa et al., 2012). We have shown here that diurnal variations in the liquid water content of in situ snowpacks can be derived from electrical self-potential data and bulk discharge measurements with a simple lysimeter. This derivation was subject to four key assumptions (Sect. 4) which we now examine

in turn to identify what, if any, constraints arise for future applications.

The Reynolds number ( $Re$ ) is a common measure of the mode of fluid flow through porous media, as discussed in a relevant cryospheric context by Kulesa et al. (2003a):

$$Re = \frac{\rho_s v L}{\eta}, \quad (9)$$

where  $v$  and  $L$  are respectively characteristic fluid flow velocity (in  $\text{m s}^{-1}$ ) and characteristic length scale of flow (in m), and  $\rho_s$  and  $\eta$  are respectively snow density (in  $\text{kg m}^{-3}$ ) and dynamic viscosity (in Pa s). To a first approximation the transition from laminar to turbulent flow nominally occurs when  $Re \approx 10$ , although laminar flow can persist at much higher values of  $Re$  (for comparison, in open channels transition occurs at  $Re \approx 2300$ ). For our purposes  $v$  can be assumed to correspond to the average linear velocity of flow,  $v = QA^{-1}n^{-1}$ , where  $n$  is effective porosity (ratio of snow and ice densities). In porous media such as snow,  $L$  corresponds to the average pore diameter, and in the absence of

direct evidence is assumed to be equal to grain size. Where snow is denser than  $\sim 490 \text{ kg m}^{-3}$ , such as that at our study sites (average  $\sim 564 \text{ kg m}^{-3}$ ), grain size is expected to be larger than pore diameter (Schneebeli and Sokratov, 2004). This assumption is therefore likely an overestimation of pore diameter. For the respective snow properties and their uncertainties reported in Table 1, values of  $Re$  between  $<1$  and  $51$  are obtained, with a best estimate of  $Re \approx 1.1$ . These values pertain to times of highest measured meltwater discharge when the Reynolds number is likely to be at its greatest. Despite the unrealistically large uncertainty bounds considered in Table 1, and the overestimation of pore diameter ( $L$ ) and associated inflation of the Reynolds number (Eq. 9), we can therefore conclude that meltwater flow in our snowpacks was laminar. The absolute and relative inclinations of the snow surface and base will vary to different degrees within different field areas, thus generating differences in discharge and potentially preferential flow. Indeed, it is an exciting attribute of self-potential measurements that they will, in practice, aid in delineating such differences in meltwater flow.

Persistent meltwater runoff at the snow surface is uncommon, and meltwater flow through underlying soils or ice will normally be negligible or small compared to flow through or at the base of snowpacks. We have also shown that the estimation of snow properties, such as liquid water content, from self-potential data is insensitive to the area of snowpack contributing meltwater flow to the measured signals. Uncertainties in the area of origin of water contributing to measured bulk discharges and thus measured self-potential data are not therefore expected to be a major hindrance to future applications of the self-potential method to snow problems. We have also shown that with the exception of the zeta potential, sensitivity to uncertainties in the snow properties governing the relationship between self-potential data and liquid water contents is small ( $\sim 3\text{--}4\%$  in our feasibility study). Future work must ascertain to what extent longer term monitoring studies are affected by the preferential elution of ions and the associated impacts on meltwater pH, electrical conductivity and thus the zeta potential. Even if such effects were found to be of concern, meltwater electrical conductivity and pH are readily monitored in situ with automated probes and could be measured alongside self-potential data at a calibration location, and subsequently be assimilated in snow models. Being able to characterise liquid water content over significant spatial areas is limited to the spatial distribution and density of possible electrode placement. However, the robustness of the estimation means that in practice, self-potential measurements at several points within the area of interest can in the future make reliable interpolations between measurements in space and time.

The final consideration focused on the assumption that the spatial pattern of self-potential magnitudes, measured during the day across our survey areas, was due to temporal changes in the liquid water content of the snowpack. This assumes that any spatial pattern due to elevation changes between the

bottom and top of our survey areas is comparatively small and indeed negligible. Kulesa et al. (2003a) showed that elevation-driven changes in the self-potential fields measured between upstream ( $\Psi_{\text{up}}$ ) and downstream ( $\Psi_{\text{down}}$ ) locations ( $z_{\text{up}}, z_{\text{down}}$ ) can be approximated by

$$\psi_{\text{up}} - \psi_{\text{down}} = -\frac{\varepsilon \zeta}{\eta \sigma_w} S_w (z_{\text{up}} - z_{\text{down}}), \quad (10)$$

here translated to our notation and adjusted for meltwater saturation according to Eq. (2). Even for the maximum daily values of saturation inferred from our measurements, the elevation-driven spatial pattern has small magnitudes, estimated to be  $\sim -16.0$  and  $-8.4$  mV respectively for Jungfrau-joch Glacier and Rhone Glaciers. These values are an order of magnitude smaller than daily changes measured at the two glaciers (Fig. 2) and are therefore considered to be insignificant for the purpose of the present feasibility study. In similar future applications the relevance of such spatial changes should be assessed on a case by case basis, and would in fact be readily incorporated into quantitative inferences of snow properties from self-potential data where they are of concern.

Overall our findings imply that in principle, self-potential data could be inverted for spatial or temporal variations in any one desired parameter (i.e. discharge, liquid water content, permeability or water chemistry), if independent estimates of the respective remaining parameters are available. Self-potential data are therefore well suited for assimilation in snow models along with meteorological and snowpack observations. We have shown in previous cryospheric applications that self-potential monitoring is readily effected with autonomous arrays of low-cost non-polarising electrodes connected to a high-impedance data logger (Kulesa et al., 2003a, b, 2012). In operational practice for instance, 2-D vertical arrays of electrodes and data loggers could be installed along with meteorological stations and upward-looking radar instrumentation, where the latter is used to monitor snow structure and 1-D liquid water contents. Assimilation of self-potential data along with complementary meteorological and radar data could then facilitate unique insights into daily and longer term variations in 2-D vertical, lateral and preferential meltwater flows, or in liquid water contents. We conclude that the integration of self-potential measurements into existing snow measurement and data assimilation routines shows considerable promise in supporting a reduction of uncertainty in quantifying snow–atmosphere energy exchanges, or in predictive modelling used in operational snow forecasting.

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