



### RESEARCH ARTICLE

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#### Key Points:

- Characterization of wetland hydrological processes by resistivity monitoring
- Highlight seasonally dependent, layered groundwater system
- Quantification of peat shrink-swell by edge detection applied to ERI data

#### Supporting Information:

- Supporting Information S1

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## Integrated time-lapse geoelectrical imaging of wetland hydrological processes

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**Abstract** Wetlands provide crucial habitats, are critical in the global carbon cycle, and act as key biogeochemical and hydrological buffers. The effectiveness of these services is mainly controlled by hydrological processes, which can be highly variable both spatially and temporally due to structural complexity and seasonality. Spatial analysis of 2-D geoelectrical monitoring data integrated into the interpretation of conventional hydrological data has been implemented to provide a detailed understanding of hydrological processes in a riparian wetland. A two-layered hydrological system was observed in the peat. In the lower part of the peat, upwelling of deeper groundwater from underlying deposits was considered the driver for a 30% increase in peat resistivity during Winter/Spring. In Spring/Summer there was a 60% decrease in resistivity in the near-surface peats due to plant transpiration and/or microbial activity. Water exchange between the layers only appeared to be initiated following large drops in the encircling surface water stage. For the first time, we demonstrated that automated interpretation of geoelectrical data can be used to quantify ground movement in the vertical direction. Here, we applied this method to quantify shrink-swell of expandable soils, affecting hydrological parameters, such as, porosity and permeability. This study shows that an integrated interpretation of hydrological and geophysical data can significantly improve the understanding of wetland hydrological processes. Potentially, this approach can provide the basis for the evaluation of ecosystem services and may aid in the optimization of wetland management strategies.

### 1. Introduction

Groundwater-dependent ecosystems (GDEs) are priority habitats under the European Habitats Directive [EEC, 1992]. Wetlands form one of the main GDE typologies and, in addition to their habitat provision, are notable for their critical contribution to the global carbon cycle [Bridgham *et al.*, 2013; Mitsch *et al.*, 2013]. For example, wetlands account for 20–25% of organic carbon in soils [Gorham, 1995], contribute 20–40% of current global methane emissions [Bloom *et al.*, 2010], and explain 70% of annual variations in atmospheric methane concentrations [Bridgham *et al.*, 2013]. The main drivers regulating wetland carbon cycles are related to hydrological factors such as water movement [Limpens *et al.*, 2008]. These factors are also a key control for the establishment, maintenance and restoration of wetland habitats [Mitsch and Gosselink, 2007]; impacting both vegetation [Baldwin *et al.*, 2001; House *et al.*, 2015a] and fauna [Ausden *et al.*, 2001; McMenamin *et al.*, 2008].

In addition to their impact on carbon cycles, riparian wetlands also act as key biogeochemical and hydrological buffers [Mitsch, 1992; Gilliam, 1994; Mitsch and Gosselink, 2007]. They can provide a nitrate and carbon sink, thus improve water quality, and may act as flood defenses [Theriot *et al.*, 2013; Martinez-Martinez *et al.*, 2014]. However, around 55% of total wetland area has been lost in the last century [Davidson, 2014] and there are now national and international initiatives to encourage their restoration.

Hydrological processes in wetlands can be highly spatially and temporally dependent, as a result of heterogeneity in the subsurface [Holden and Burt, 2002; House *et al.*, 2015a]. There is also ongoing uncertainty regarding water movement within peat-dominated wetlands [Reeve *et al.*, 2000, 2006]. Seasonal changes in moisture content are known to result in changes in peat volume with moisture loss causing peat compaction and moisture gain peat expansion. These volumetric changes directly affect the hydraulic properties of peat causing changes in hydraulic conductivity and porosity [Price and Schlotzhauer, 1999; Kennedy and

Price, 2005]. For example, peat shrinkage has been demonstrated to reduce hydraulic conductivity by up to three orders of magnitude and therefore aiding as a self-preservation mechanism against further moisture loss [Chow *et al.*, 1992; Price, 2003]. Thus, an understanding of peat shrink and swell is essential to accurately model hydraulic processes in peat-dominated wetlands.

Investigations of subsurface hydrological properties and processes in wetlands are typically restricted to a limited number of piezometers [Bradley, 1997; Kehew *et al.*, 1998]. This is a result of wetland inaccessibility due to ecological sensitivity, their frequent inundation, as well as the high costs of drilling. Geophysical methods can noninvasively and cost-effectively achieve high spatial coverage to improve our hydrological understanding in these heterogeneous settings [Kettridge *et al.*, 2008; Lowry *et al.*, 2009; Comas *et al.*, 2011; Chambers *et al.*, 2014b].

Electrical resistivity imaging (ERI) concerns the measurement of electrical potentials at discrete locations along the surface or in boreholes caused by the application of electrical currents. It is a well-developed method for the volumetric imaging of temporal hydrological processes [Cardenas and Markowski, 2011; Musgrave and Binley, 2011; Binley *et al.*, 2015], as the resistivities of earth materials are sensitive to changes in moisture content, temperature and pore water chemistry. Geophysical techniques targeting electrical (and dielectrical) subsurface properties, such as ground penetrating radar, and electromagnetic and geoelectrical methods, are also commonly used to investigate subsurface structures in wetlands [Comas *et al.*, 2004; Kettridge *et al.*, 2008]. Monitoring studies aiming to image time varying wetland processes are more rarely applied. ERI monitoring has been used to reveal temperature dynamics in a wetland environment indicating groundwater recharge [Musgrave and Binley, 2011], and to image methanogenesis in peat blocks [Slater *et al.*, 2007] by exploiting the resistivity contrast caused by the seasonal entrapment of non-conductive gas within the originally saturated and conductive pore space. Laboratory studies on the electrical properties of peat have also showed that its bulk resistivity is highly sensitive to changes in pore water resistivity [Comas and Slater, 2004; Ponziani *et al.*, 2012].

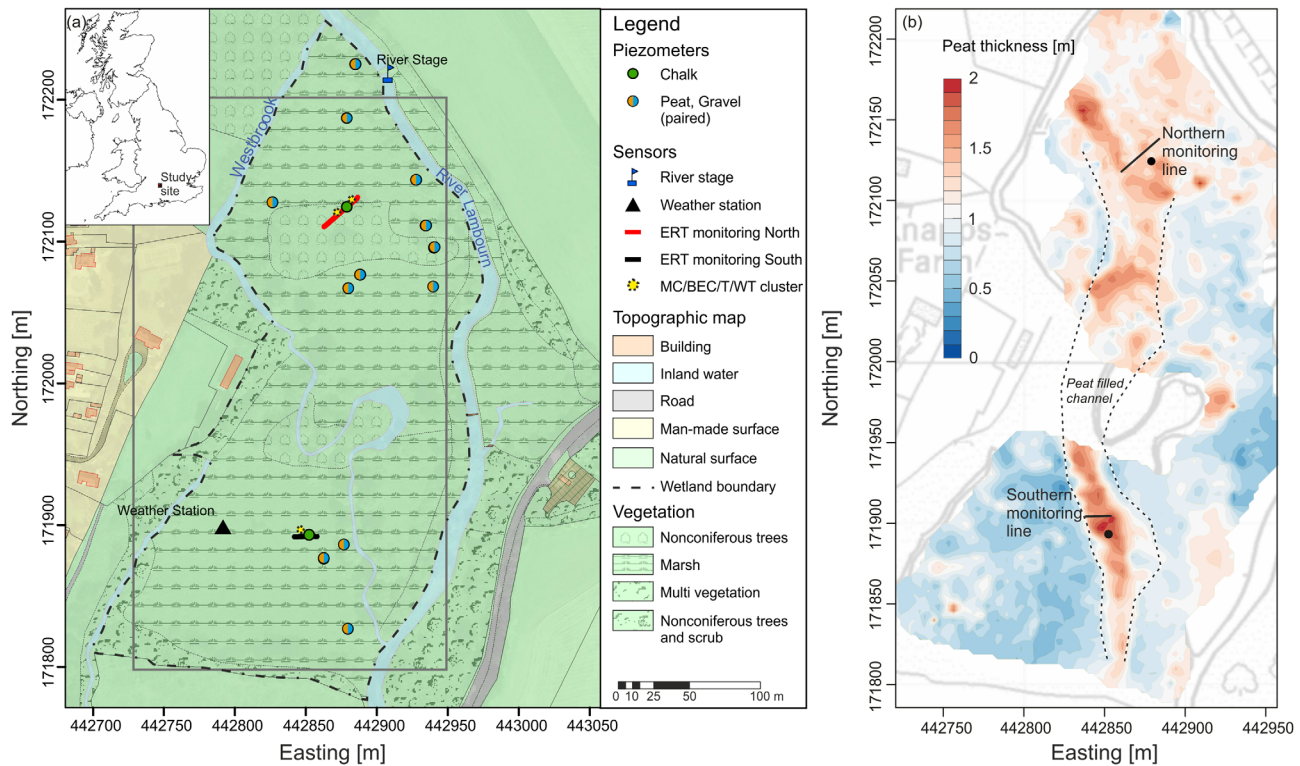
In recent years, automated interpretation of resistivity models has been developed [Chambers *et al.*, 2012, 2014b, 2015; Audebert *et al.*, 2014; Ward *et al.*, 2014] and applied to extract interfaces of lithological boundaries or other target features from 2-D/3-D resistivity images. It has been shown that the accuracy of such methods can be comparable to shallow intrusive investigations, such as sonic or shell-and-auger drilling [Chambers *et al.*, 2012, 2014b], when estimating depth-to-bedrock. Applying these methods to time-lapse imagery could potentially identify movement within any target interface. Expansive soils, such as peat, are subject to large volume changes. Therefore, as distances from the surface to lithological boundaries change due to electrodes moving with the soil, automated ERI interface detection could identify seasonal shrink-swell during annual drying-wetting phases.

The aim of this study is to apply time-lapse geoelectrical imaging integrated with hydrological and environmental sensor data to improve the understanding of wetland functioning as a regulator of both carbon cycling and habitat provision. In particular, this research is focused on the imaging of hydrological processes affecting the resistivity signature of near-surface pore waters, and potential upwelling of nutrient rich, deeper waters into shallow peat layers. Detecting seasonal changes in peat volume from ERI data will also aid in supporting future studies on hydrological process modeling without the need for additional monitoring of surface movements, by providing information on the temporal characteristics of water storage capacities and hydraulic conductivities. Integrating geoelectrical and conventional hydrological data allows to study spatially complex processes, which are not able to be captured employing point sensor data only.

## 2. Methodology

### 2.1. Study Area

The study area is a 10 ha riparian wetland adjacent to the River Lambourn, in Boxford, Berkshire, UK (Figure 1a). It is designated as a Special Site of Scientific Interest (SSSI) owing to the habitat it provides for the near-threatened Desmoulin's whorl snail [Killeen, 2003]. The River Lambourn drains the Chalk of the Berkshire Downs and is strongly groundwater dependent with a base flow index of 0.96 at the nearest gauging station at Shaw [Marsh and Hannaford, 2008]. It has a near-natural flow regime and good water quality, thus providing a reference observatory against which to compare other more anthropogenically impacted sites.



**Figure 1.** (a) Map of the River Lambourn Observatory and the instrumentation network. Inset shows the location of the observatory in the UK. MC—Moisture content, BEC—Bulk Electrical Conductivity, T—Temperature, WT—Water table. (b) Grey box outlines area, which shows the ERT monitoring locations superimposed on a map of intrusively determined peat thicknesses (modified from Chambers *et al.* [2014b]). Figures 1a and 1b are plotted on an Ordnance Survey base layer (© Crown copyright and database rights 2015 Ordnance Survey 100021290).

The wetland is located 13 km downstream of the ephemeral river head. The underlying bedrock is the Cretaceous Seaford Chalk Formation, a fine-grained white limestone with many flint nodules. However, much of the upper part of the profile has been weathered into a low permeability structureless ‘putty’ chalk, which is considered to confine the aquifer where present [House *et al.*, 2015a]. The irregular surface of the Chalk is overlain by river terrace deposits that are primarily coarse-grained gravels, which display a clear braided structure [Chambers *et al.*, 2014b]. In the basal 1–2 m of the gravels, reworked chalk can be present occluding the available porosity and potentially permeability [Allen *et al.*, 2010]. Water movement within the saturated zone is thus predominantly controlled by the numerous vertical and horizontal discontinuities in the chalk and overlying gravels. These gravels are, in turn, overlain by Holocene alluvium (typically < 1 m) comprising predominantly peat, with the thickest deposits occupying a channel-like structure through the site upon a topographic low on the gravels surface (Figure 1b) [Chambers *et al.*, 2014b]. Heterogeneity within the alluvium partly reflects climatic change and anthropogenic impacts, such as large-scale deforestation within the catchment, which are frequently observed in Holocene floodplain sediments [Newell *et al.*, 2015]. Notably, there is an intermittent layer of 0.1–0.2 m of chalky clay within the predominantly peat alluvial profile.

The wetland is divided into a northern and southern meadow by the Westbrook Channel, which diverts water from the Lambourn, and through the site, before rejoining the river (Figure 1a). A spring-fed channel also flows south-westward along the western margins of the southern meadow. Consequently the wetland is almost entirely encircled by surface waters. These exert a strong control on the water table across the wetland. Increases and rapid reductions in channel stage caused by seasonal weed growth and cutting (to increase river flow conveyance), respectively, result in near equivalent changes in the peat water table [Old *et al.*, 2014]. Nevertheless, the site is considered groundwater dependent [House *et al.*, 2015b], with upwelling water from the Chalk thought to be focused within discrete relic channels in the peat, particularly in the northern meadow. These discharges increase the floral biodiversity of the site through their provision of high nitrate, low phosphate waters, in a wetland dominated by poor fen communities. Away from the

localized upwelling, there is strong evidence for nitrate removal through reductive bacterial processes [House *et al.*, 2015a].

## 2.2. ERI Monitoring Installation

In a previous study, Chambers *et al.* [2014b] employed 3-D geoelectrical imaging and interface detection methods to reconstruct the deposit architecture of the Lambourn observatory. They imaged the spatial variability of peat thickness, revealing a channel of thick peat deposits crossing the wetland, and correlated the data with intrusive peat thickness estimates (Figure 1b). Within the underlying gravels, structural features characteristic of braided fluvial systems were found. This study focusses on time-lapse 2-D ERI to characterize dynamic hydrological processes within the wetland.

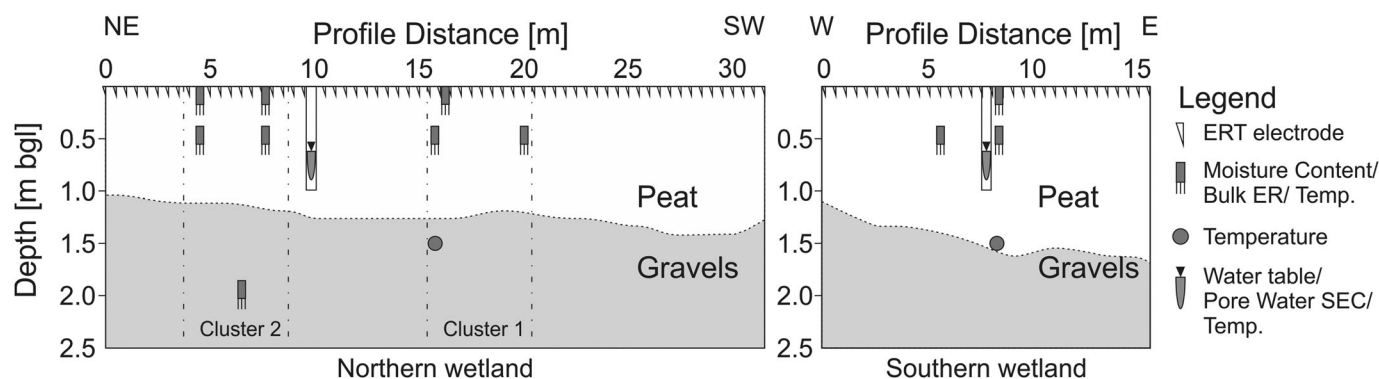
Two geoelectrical monitoring arrays were installed in each meadow of the wetland during December 2012 (Figure 1). The locations targeted thicker peat deposits ( $>1$  m), including a channel-structure identified by intrusive peat thickness measurements (Figure 1b) and 3-D ERI surveying in the southern meadow.

An electrode spacing of 0.5 m was chosen to provide sufficient lateral and vertical resolution to image hydrological processes in the highly heterogeneous wetland material. This spacing enabled imaging of the resistivity distribution (with final model cell areas ranging between  $0.035\text{ m}^2$  and  $0.145\text{ m}^2$ ) to a depth of up to 1.5 m below the peat-gravel interface, which can be found between 1.0 m and 1.5 m below ground level (bgl). The northern and southern arrays comprised 64 stainless-steel electrodes (31.5 m in length) and 32 stainless-steel electrodes (15.5 m in length), respectively, permanently installed just below the surface. The northern array was greater in length to traverse two relic channels within the peat that were considered to be areas of groundwater upwelling [House *et al.*, 2015a], whilst the southern array targeted a single relic channel-structure containing the thickest peat deposits [Chambers *et al.*, 2014b].

A dipole-dipole type measurement configuration was chosen with dipole lengths ( $a$ ) of 0.5 m, 1.0 m, 1.5 m, and 2.0 m, and dipole separations ( $n$ ) of  $1a$  to  $8a$ . The dipole-dipole measurement configuration was chosen as it provides good resolution of both vertical and horizontal resistivity changes [Chambers *et al.*, 2002; Dahlin and Zhou, 2004]. Calculating the depth of investigation (DOI) index [Oldenburg and Li, 1999] shows high sensitivity throughout the imaging section with DOI indices  $R < 0.3$ . Employing a cut-off value of  $R = 0.05$  indicated the effective depths of investigation to be  $\sim 3$  m and  $\sim 2$  m for the northern and southern array, respectively. The peat-gravel interface is at a depth where it can be expected to be well-resolved by the longer northern array, but is likely to be less well resolved by the southern array, partly because the interface is at a greater depth and partly because the southern array is shorter. ERI monitoring commenced in December 2012, and was followed by monthly data acquisition from March 2013, resulting in 21 data sets over a 24 month period.

To obtain a model of the subsurface resistivity distribution, the measured data need to be inverted. ERI inversion is fundamentally a non-unique process (in that an infinite number of models fit the data equally well [e.g., Olayinka and Yaramanci, 2000]), so smoothness constraints are applied to define an optimal solution in terms of a smoothly varying resistivity model. Hence, additional subsurface data are needed to inform the ERI inversion and to compare the resulting models, thereby ensuring their reliability [e.g., Kuras *et al.*, 2009; Wagner *et al.*, 2013; Gunn *et al.*, 2014; Hermans *et al.*, 2015]. For this purpose, environmental sensors measuring moisture content (MC), bulk electrical resistivity (BER), temperature, and water table (WT) were installed at different locations and depths along the monitoring transects (Figure 2).

In the northern meadow, clusters of sensors were installed in two separate phases. The first cluster was installed with the ERI array in December 2012. This comprised MC, BER (corrected for temperature effects; sampling volume  $\sim 2.75 \times 10^{-4}\text{ m}^3$ ), and temperature sensors at 0.1 m and 0.5 m depth, and an additional temperature sensor at 1.5 m depth (Figure 2). The second cluster was established in December 2013, with MC, BER, and temperature sensors installed at 0.1 m, 0.5 m, and 2 m below ground level, where the deepest sensor was within the gravels. These additional sensors were installed after a preliminary analysis of the ERI monitoring data [Uhlemann *et al.*, 2014]. In the southern meadow, the sensor installation was similar to the first phase in the northern part and installed in December 2012. In both the northern and southern meadow a piezometer adjacent to the array was equipped with instrumentation recording water level (assumed equivalent to hydraulic head), temperature, and pore water electrical conductivity in April 2013. All sensors were logged at 30 min intervals.



**Figure 2.** Schematic of the ERI arrays and sensor installations for the northern and southern meadows. Note that the DOI index of the northern and southern installation was  $R = 0.05$  at about 3 m and 2 m depth, respectively, defining the effective depth of investigation.

### 2.3. Resistivity Data Error

For data quality control, every ERI data set was recorded employing a full set of reciprocal measurements, comprising 1528 and 516 measurement pairs for the northern and southern meadows, respectively. The reciprocity theorem states that interchanging current injection and potential dipole (i.e., normal and reciprocal measurement) should result in the same data. Hence, any difference between the two measurements can be used as a robust estimate of the data error [LaBrecque *et al.*, 1996].

These reciprocal errors were calculated for each data set, creating a time-series of error distributions over the monitoring period (Figure 3). Reciprocal errors were also processed and analyzed as outlined in Koestel *et al.* [2008] in order to develop an error model. This model represents the error comprising an absolute measurement error component  $a_{N/R}$  and a relative measurement error component  $b_{N/R}$  applied to  $R_{N/R}$  the mean of the measured normal and reciprocal resistance value. Thus, the reciprocal error can be approximated by:

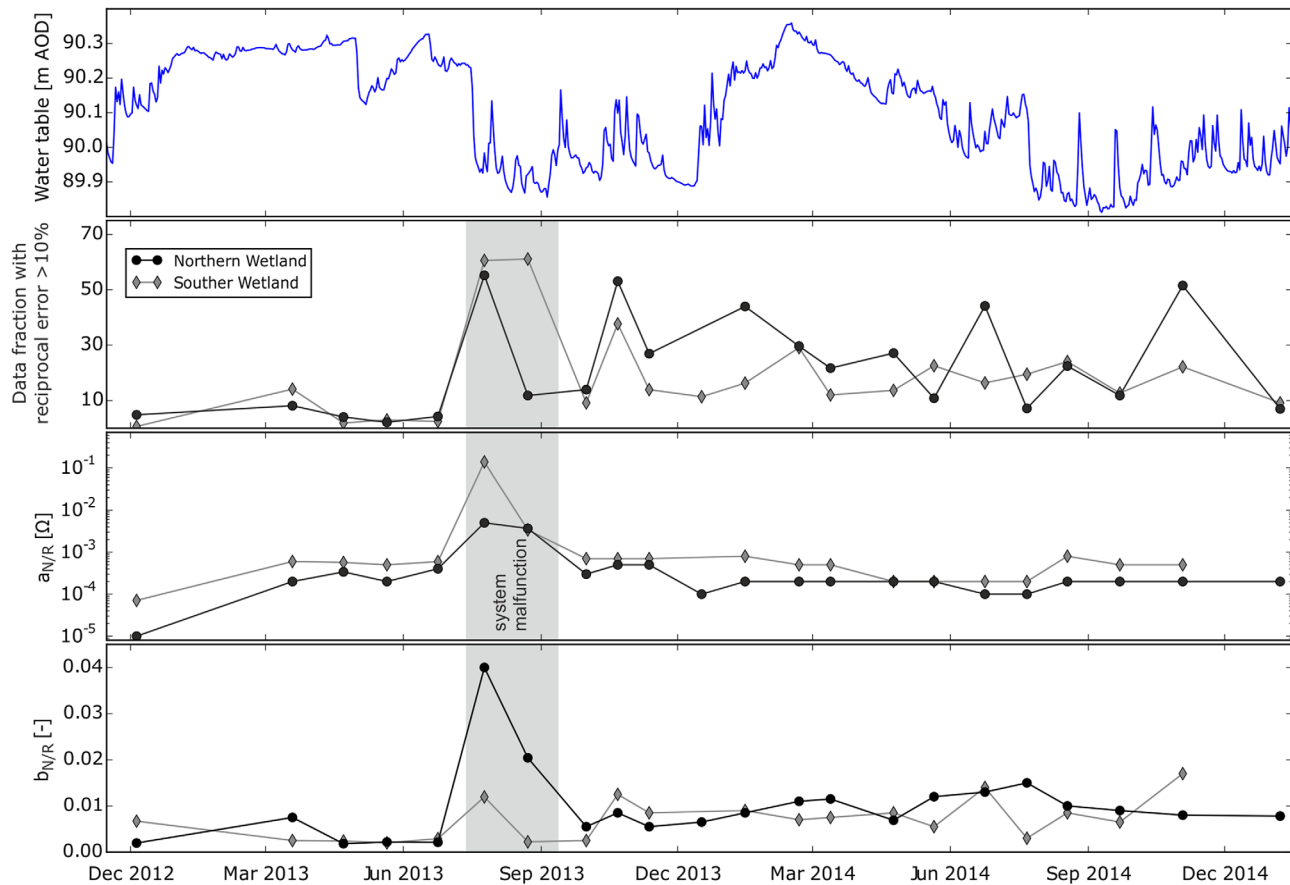
$$\varepsilon_{N/R} = a_{N/R} + b_{N/R} |R_{N/R}| \quad (1)$$

Considering the complete time series, the data of the northern meadow show very good quality with only about 8% having reciprocal errors above 10%. The smallest errors were recorded before a drop in water table in August 2013 (< 5% of data with errors > 10%), after which errors increased (10% – 50% of data with errors > 10%) despite water table recovery and larger variability. A system malfunction in July and August 2013 could be identified by  $a_{N/R}$  exceeding a threshold of  $10^{-3} \Omega$  and comparably high values for  $b_{N/R}$ . Except for these outliers,  $a_{N/R}$  and  $b_{N/R}$  show significantly less variation than the reciprocal error threshold and are comparable at both meadows.

Before inversion, data with reciprocal errors exceeding a threshold of 10% were removed from the data set and replaced by values derived from their respective time-series, using an inverse distance weighting interpolation. This step was necessary, as the employed time-lapse inversion routine requires each measurement set to comprise the same dipole-dipole measurements. Note that data from July and August 2013 were excluded from the time series due to malfunctioning of the ERI measurement system.

### 2.4. Data Inversion

Data were inverted using a smoothness-constrained least-squares inversion method, with the data defined in a space-time domain, allowing for a simultaneous inversion of the complete data set [Kim *et al.*, 2009]. The inversion employed a L1-norm for the data misfit and a L2-norm for the model roughness [Loke and Barker, 1996], favoring smooth resistivity changes in both space and time. To account for the different resolution capabilities, active constraint balancing has been applied to automatically choose the most appropriate Lagrangian multiplier [Yi *et al.*, 2003]. Data, as well as the root-mean-square (RMS) misfit error between modeled and measured data, were weighted by their respective measured reciprocal errors (supporting information equation (S1)), opposed to weighting by the modeled error values, provided the error was above a lower limit of 10% the average percentage noise level, which was calculated from the error values between the 10% and 90% percentiles. This step limits the range of inversion weights calculated from the



**Figure 3.** Data error characteristics and water table of the northern meadow. Given are the measured fraction of the data having a reciprocal error above 10%, as well as the parameters  $a_{N/R}$  and  $b_{N/R}$  of the error model introduced by Koestel *et al.* [2008].

reciprocal error values (M.H. Loke, personal communication, 2015). This was done because, although data weighting by modeled and reciprocal error estimates produced virtually identical resistivity models (supporting information Figure S1) in standalone inversions of two test data sets, employing the modeled error showed artifacts in the time-lapse inversion. This is likely to be caused by an underestimation of the data error for a few measurement configurations [LaBrecque *et al.*, 1996]. For the interpolated values a high error value of about 3 times the standard deviation was chosen to reduce the influence of these data points on the inversion. The inversion converged after 5 iterations as the RMS error between modeled and measured data changed by less than 5% between the last two iterations. The remaining RMS error was 1.79% and 2.88% for the northern and southern meadows respectively, highlighting the good agreement between the modeled and observed data. Note that the average data error was 0.23% and 0.29%, respectively.

Peat expansion and compaction will lead to vertical movements of the electrodes. Note that in the resistivity inversion the electrode locations, and thus the surface topography, are fixed values; this is a limitation of current inversion algorithms. RTK-GPS measurements showed, however, that the topography changes differently along the line over the monitoring period. These lateral variations were found to be less than 5 cm (i.e., 10% of the electrode spacing). While inline displacements of this magnitude are known to cause artifacts of up to 20% in the resistivity models [Szalai *et al.*, 2008; Wilkinson *et al.*, 2010; Uhlemann *et al.*, 2015], Wilkinson *et al.* [2015] showed that sensitivities of dipole-dipole measurements to electrode displacements perpendicular to the surface are significantly smaller. As introduced slope angles are also  $\ll 10^\circ$ , the errors introduced by assuming a constant surface elevation should be negligible [Tsourlos *et al.*, 1999]. Topographic information from February and September 2014 (representative for maximum and minimum annual elevation) was used in a numerical modeling exercise, showing that resistivity variations caused by variations in electrode positions are less than 10% (see supporting information Figure S2).

### 2.5. Temperature Correction

Variations in temperature can have significant effects on geoelectrical monitoring experiments [Rein *et al.*, 2004; Hayley *et al.*, 2007] and resistivity models need to be corrected to a standard temperature to avoid misinterpretation of resistivity data [Chambers *et al.*, 2014a; Chrétien *et al.*, 2014]. Therefore, multilevel temperature data, from 0.1, 0.5, and 1.5 m below ground level, were fitted to the following heat equation employing the methodology described in Chambers *et al.* [2014a]:

$$T_{\text{model}}(z, t) = T_{\text{mean}} + \frac{\Delta T}{2} \exp\left(-\frac{z}{d}\right) \sin\left(\frac{2\pi}{365}t + \varphi - \frac{z}{d}\right), \quad (2)$$

where  $T_{\text{mean}}$  is the average air temperature,  $\Delta T$  the amplitude of the temperature variation,  $z$  the depth below ground level,  $d$  a characteristic depth by which the amplitude of the temperature variation reduced by  $1/e$ ,  $t$  the day of the year, and  $\varphi$  a phase offset to ensure that the surface temperature variation is in phase with the air temperature.

The resistivity models were subsequently corrected ( $\rho_{\text{cor}}$ ) using the ratio model [Hayashi, 2004; Ma *et al.*, 2010], which in terms of resistivity can be written as:

$$\rho_{\text{cor}} = \rho \left[ 1 + \frac{c}{100} (T_{\text{target}} - T_{\text{model}}) \right] \quad (3)$$

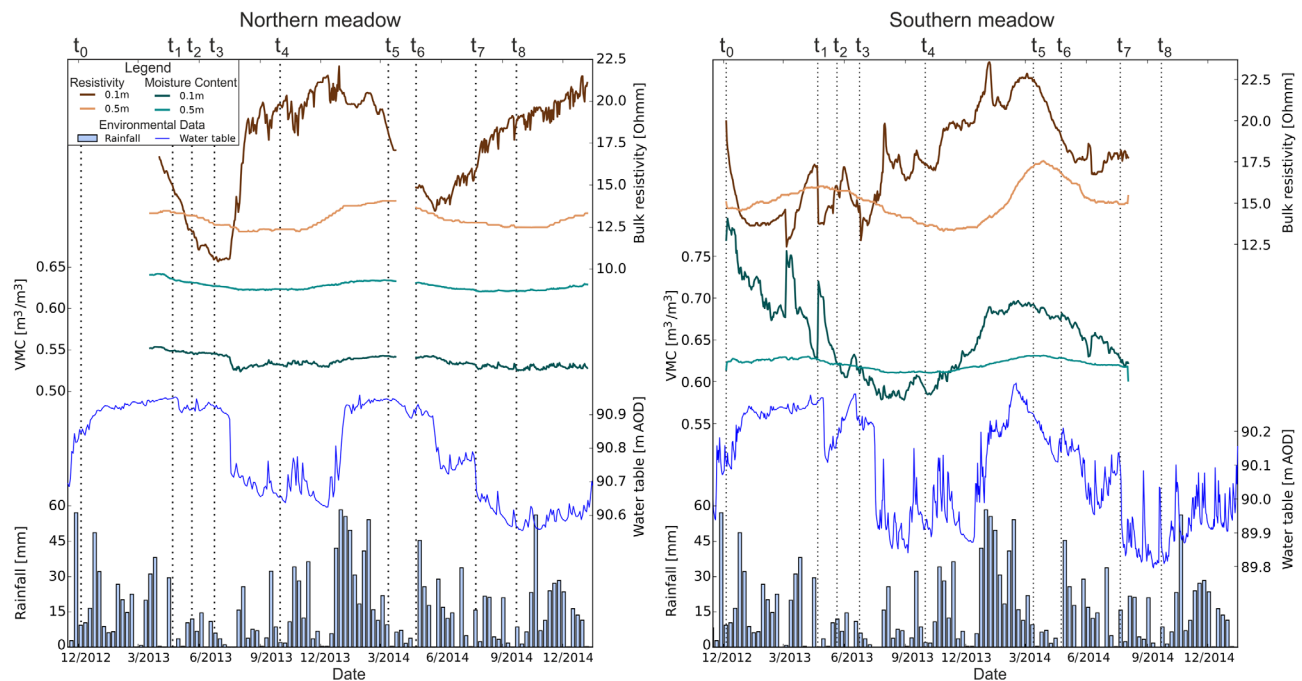
All models were corrected to a target temperature of  $T_{\text{target}} = 9.84^\circ\text{C}$  (the annual mean air temperature at the site over the monitoring period), employing a correction factor  $c$  of  $-2.95^\circ\text{C}^{-1}$ . This value was established in the laboratory by temperature-cycling soil samples under constant moisture content and determining their resistivity. Results thus show the resistivity evolution over time compensated for laterally uniform seasonal temperature variations with depth. Note that non-seasonal temperature effects will still influence the imaged resistivity distribution.

### 2.6. Characterizing Shrink-Swell

The imaging space included a significant change in lithology from low resistivity peat to highly resistive gravels (Figure 2). Assuming that the interface between peat and gravel is fixed in space, only shrinkage or swelling of the peat layer can result in changes of the surface elevation. To test whether it is possible to use ERI to image these changes, the peat thickness of the northern meadow was estimated for every survey using both geoelectrical and real-time kinematic global positioning system (RTK-GPS) measurements. From each resistivity model the boundary between peat and gravel was extracted using a fuzzy  $c$ -means clustering technique [Ward *et al.*, 2014]. This comprises the characterization of the resistivity probability distribution within each model using kernel density estimation to identify distinct resistivity populations that are representative for the peat and gravel. Two approaches to extract the boundary resistivity value have been tested: (1) calculating the probability distribution and clusters for each time step separately and (2) using the probability distribution and clusters for the data from all time steps. For the first approach the boundary is defined by a different resistivity at each time step; for the second the same resistivity value is used for each time step. These two approaches were employed to investigate the effect of the change in resistivity caused by peat expansion and compaction on the estimate. While during expansion pore volume increases and thus resistivity is likely to decrease, during compaction the peat pore volume reduces and resistivity should increase. Considering the likely variations in peat volume and the sensitivity of resistivity to changes in pore volume, the effect on the resistivity falls below the error levels. To quantify shrink-swell of the peat the surface elevation of the monitoring line was resurveyed for a subset of site visits using RTK-GPS, with repeated GPS surveys showing a precision of  $< 3$  cm.

### 2.7. Supporting Hydrological Data

Supplementary hydraulic head data were available from a network of paired piezometers within the peat and gravels (Figure 1). The River Lambourn stage was continually monitored with a pressure transducer, which was corrected to weekly stage board readings to account for sensor drift [Sorensen and Butcher, 2011]. Groundwater samples were regularly collected from peat, gravel, and chalk piezometers following purging and specific electrical conductivity (SEC) determined in the field with a portable meter. Monthly chalk water table data were retrieved away from the site ( $51^\circ 11' 56.37''\text{N}$ ,  $01^\circ 45' 20.00''\text{W}$ ) to represent water



**Figure 4.** Daily rainfall, peat heads, and multilevel moisture content and bulk resistivity (for 0.5 m below ground level the means of the measurements of the two sensors are shown) throughout the ERI monitoring period for the two meadows. MC and resistivity data of the northern meadow were acquired at cluster 1.

table conditions within the aquifer in the Lambourn catchment, which were not impacted by weed cutting and growth in the river [Old *et al.*, 2014].

### 3. Results

#### 3.1. Hydrology

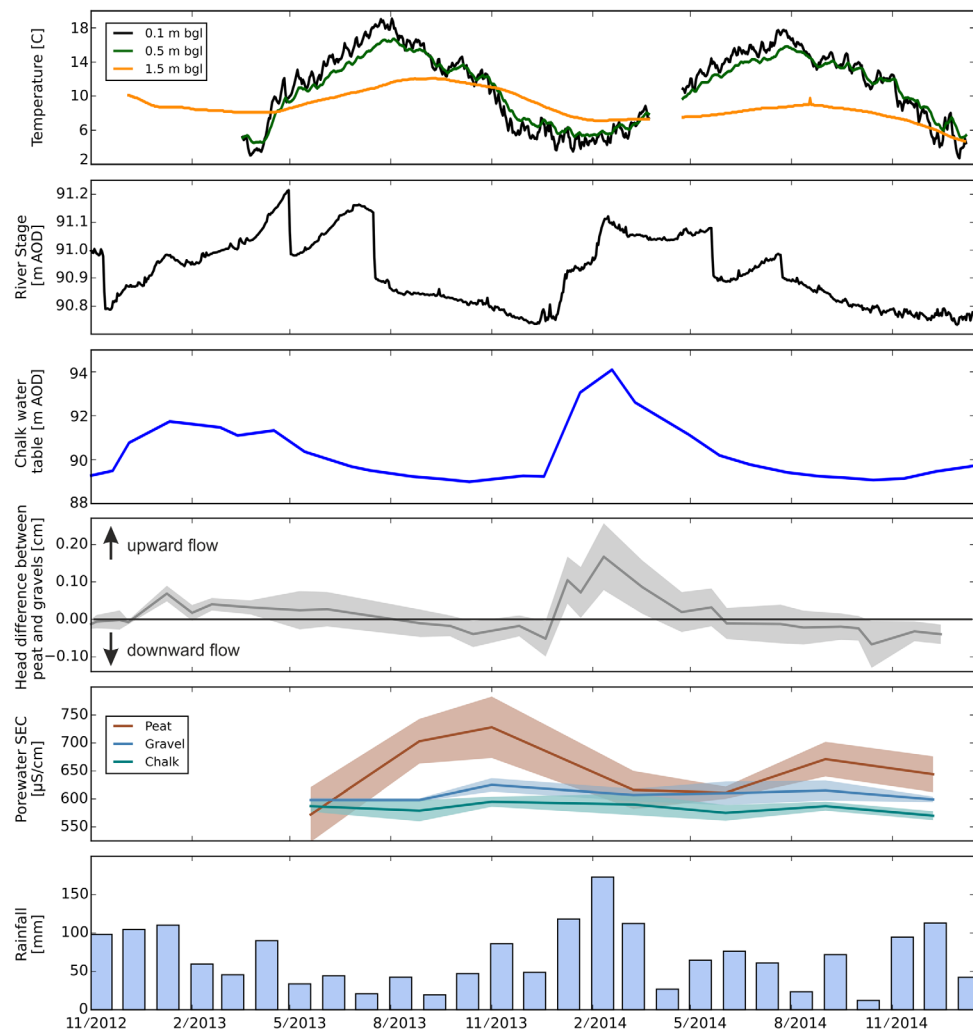
Water levels conformed to the typical seasonal pattern in the UK: peaking between January and April and reaching the annual minimum between August and November (Figure 4). Rising chalk heads appear to initiate seasonal (January–May/June) positive upward vertical head differentials between the gravels and peat within the wetland (Figure 5). These were greatest in early 2014 following exceptional rainfall in Winter 2013/2014 [Muchan *et al.*, 2015], which resulted in unusually high chalk heads. Positive vertical head differentials appear to correlate with a reduction in peat pore water SEC toward that found in the Chalk and gravels. Note that pore water SEC values were measured on water samples taken from piezometers screened across the whole peat profile.

Heads within the peat fluctuated between the ground surface and around 0.3 m bgl. Rapid recessions in peat head in response to river weed cutting and a reduction in stage were evident, e.g., July 2013, May and July 2014, initiating rapid recessions in the peat water table (Figures 4 and 5).

Changes in MC were less than  $0.05 \text{ m}^3/\text{m}^3$  in the northern meadow and were concurrent with the peat head dynamics (Figure 4). The BER data from 0.1 m depth show significant increases in the near-surface resistivity following the falling water table in July 2013 and 2014. Although the water table remained at a constant level between February and June in both 2013 and 2014, near-surface resistivities decreased by more than 80%. The deeper sensors at 0.5 m bgl show a linear correlation between resistivity and MC (Pearson's  $r = 0.81$ ), despite its inverse relationship [Archie, 1942]. Thus, in both the near-surface and deeper monitoring locations, resistivity dynamics are likely to be dominated by changes in pore water resistivity.

In the southern meadow, the more pronounced changes in MC in the near-surface align with the peat heads, but show no response to sudden head changes caused by the weed cuts in the river. The resistivity of the near-surface shows decreasing resistivity during periods of steady, high water table and moisture content (i.e., February–June 2014), while deeper sensors show increasing resistivity with increasing moisture content (Pearson's  $r = 0.78$ ; correlation most pronounced in 2014).





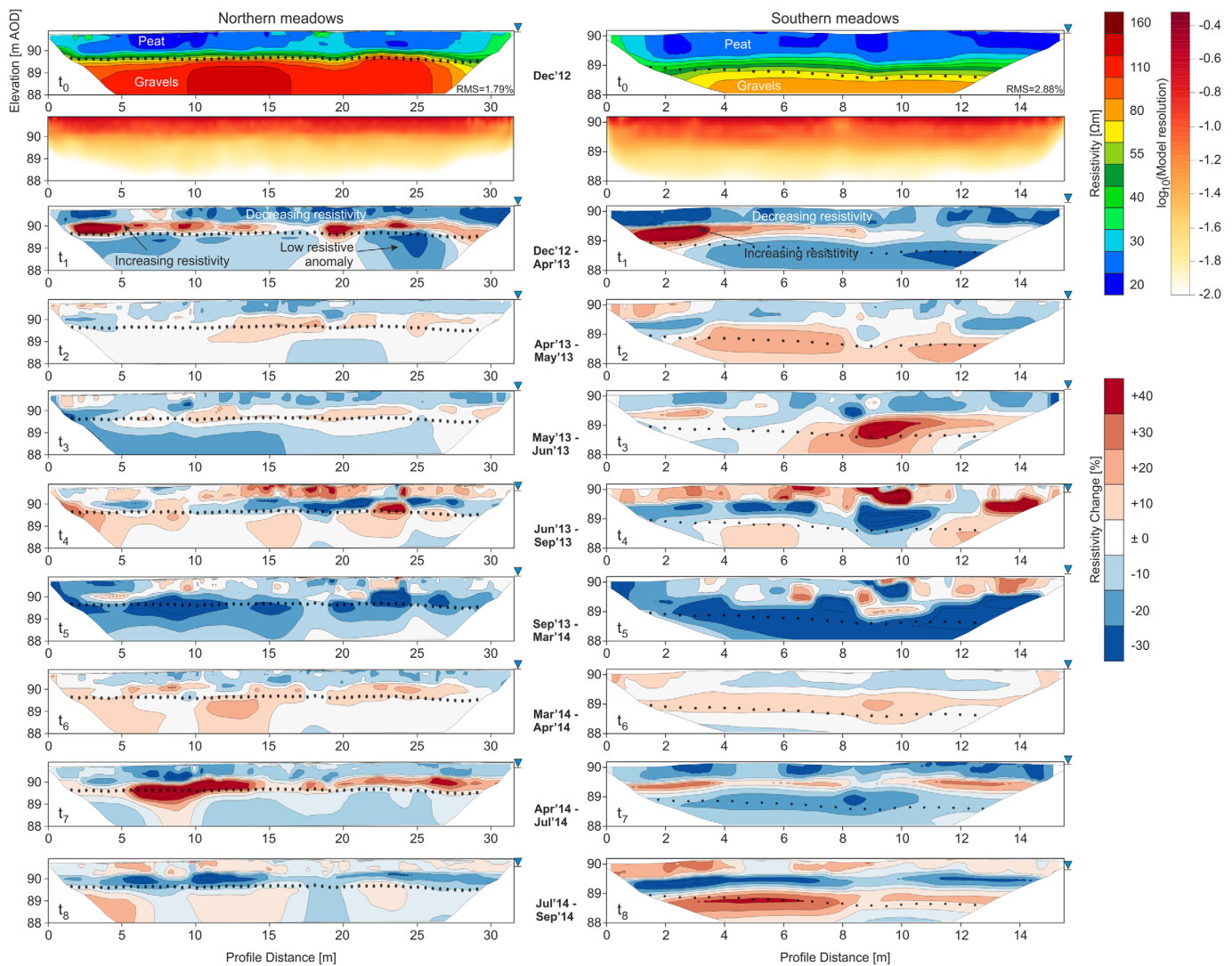
**Figure 5.** Multilevel temperature data from the weather station, river stage, head in the chalk, vertical head differential between the gravel and peat, pore water SEC for peat, gravel, and chalk, and monthly rainfall data for the monitoring period. Vertical head differentials and pore water SEC show mean values across all locations (see Figure 1); shaded areas indicate range. Note that peat pore water SEC values are derived from aggregated water samples and thus represent averaged values.

### 3.2. ERI Monitoring

The resistivity models of the baseline measurements  $t_0$  (December 2012) show low resistivity alluvium (cold colors,  $\rho < 30 \Omega\text{m}$ ), mainly consisting of peat, overlying more resistive clean gravels (warm colors,  $\rho > 55 \Omega\text{m}$ ) for both the northern and southern meadows (Figure 6). The boundary between these two materials is imaged very sharply at the northern meadow and coincides with intrusively determined peat thicknesses [Chambers *et al.*, 2014b]. For the southern meadow, however, this boundary is less pronounced, which is likely to be an effect of reduced sensitivity of the ERI method at depth caused by the use of a shorter electrode array.

Within the peat layer of the northern meadow, two areas of anomalously low resistivity are evident at profile distances  $x$  of 6–12 m and 20–26 m. These can be interpreted as structures having a higher degree of more recent fine material and organic infill, and coincide with channel structures imaged by Chambers *et al.* [2014b]. The low resistivity anomaly at  $x > 12$  m of the southern meadow is a similar channel feature. Within the gravels of the northern meadow between 10 m and 16 m a resistive anomaly can be found, which is likely to be caused by gravels of lower porosity.

To highlight changes in the resistivity models, the proceeding time steps of Figure 6 are shown as resistivity changes between two consecutive time steps (i.e., change between  $t_2$  and  $t_1$ ,  $t_4$  and  $t_3$ , etc.), with blue/red



**Figure 6.** Baseline resistivity measurements ( $t_0$ ), model resolution, and measurements of eight proceeding time steps of the northern and southern meadow. Dates of the measurements are indicated in Figure 4. The time steps proceeding the baseline results are shown as resistivity changes between two consecutive time steps (i.e., between  $t_2$  and  $t_1$ ,  $t_4$  and  $t_3$ , etc.). Blue colors indicate decreasing resistivity, while red colors indicate increasing resistivity. Dotted line shows intrusively determined interface between peat and the underlining gravel (from Chambers et al. [2014b]). The water table is indicated on the right at each time step.

colors indicating decreasing/increasing resistivities respectively. Throughout the monitoring period, resistivity changes showed significant amplitudes of  $-50\%$  to  $+100\%$ , equal to a reduction to half or an increase to twice the initial value, respectively (note that the color scale of Figure 6 has been limited to values ranging from  $-30\%$ – $+40\%$ ). Thus, these values are well above likely noise levels of the ERI changes, which were calculated to be  $\pm 9.8\%$ .

Within the peat, there are two layers that appear to behave differently according to the hydrological conditions. This is particularly clear at time step  $t_1$  from December 2012 to April 2013 (Figure 6) when the peat heads rose toward their annual maximum. In the upper layer (0.0 m – 0.5 m bgl) resistivities were decreasing, whilst in the lower layer ( $> 0.5$  m bgl) resistivities were increasing. This trend continued through time steps  $t_2$  and  $t_3$  (May–June 2013). By time step  $t_4$ , peat heads had dropped by 0.25 m in November 2013 as a result of weed cutting in July 2013, which coincided with an increasing resistivity of the upper layer and decreasing resistivity of the lower layer. There was generally decreasing resistivity within the lower layer through time step  $t_5$ . These ERI observations correlate with the bulk resistivity sensor data over this period (Pearson's  $r = 0.65$ ; Figure 4).

High water table conditions were re-established in the peat by January 2014 and there is evidence for the layered cycle in resistivity observed between  $t_0$  and  $t_5$  recommencing. At  $t_6$  (March–April 2014), resistivities were decreasing and increasing in the upper layer and lower layer, respectively, when the upward vertical head differentials reach their maximum across the site (Figure 5). Although weed cutting occurred in May 2014 and led to a drop in peat heads of about 0.15 m, resistivity changes at  $t_7$  (April–July 2014) showed a continuation and amplification of the trends of  $t_6$ . Following the drop in heads near to the annual minimum at time step  $t_8$ , caused by a weed cut in July 2014, resistivity trends within the layers reversed: increasing in the upper layer and declining in the lower layer, as observed at time step  $t_4$ .

The gravels, mainly imaged at the northern meadow, showed more spatially consistent temporal changes. At  $t_1$  a reduction in resistivity compared to the baseline measurement can be observed, which is highest just below the channel features imaged in the peat ( $t_0$ ). This trend continued through to time step  $t_3$ , where only minor resistivity changes were found in the gravels. Strongly decreasing resistivities can be found at  $t_5$  (September 2013 to March 2014), coinciding with a rise in peat heads and river stage (Figures 4 and 5, respectively). This was followed by only minor changes at  $t_6$  and minor resistivity decreases throughout  $t_7$  and  $t_8$ . Nearly throughout the entire monitoring period an area between  $x = 16$  m–21 m formed a discontinuity of the resistivity trends in the gravels of the northern meadow.

Analyses of pore waters from the peat and gravels provided representative conductivities of  $\sigma_{pw} = 900 \mu\text{S/cm}$  and  $\sigma_{gw} = 600 \mu\text{S/cm}$  for the peat and gravel, respectively. The mean resistivities of the lower peat layer showed a change from 24.8  $\Omega\text{m}$  to 32.6  $\Omega\text{m}$  during the upwelling period from February to July. Assuming that the pore waters and surface conductance  $\sigma_{surf}$  can be represented as resistors in a parallel circuit, the peat conductivity can be defined as (modified from Comas and Slater [2004]):

$$\sigma_{peat} = \left(\frac{1}{F}\right) (a\sigma_{pw} + b\sigma_{gw}) + \sigma_{surf} \quad (4)$$

with  $a$  and  $b$  defining the fraction of the pore space occupied by pore water representative for peat or gravel, respectively, and  $F$  being the formation factor. Note,  $\sigma_{surf}$  was derived from data published by Comas and Slater [2004]. Employing the representative pore water conductivities of peat and gravels, it can be calculated that the pore water in the lower peat consists of 94% gravel and 6% peat pore water at the peak of upwelling. This assumes (1) no ingress of surface waters into the lower peat, which is reasonable given the lack of inundation, as well as both the low permeability of the peat and lateral distance between the arrays and nearest surface waters, and (2) that the pore water in February consisted of peat pore water only.

Changes corresponding to the first part (i.e., reduction and increase of the resistivities of the upper and lower layers, respectively) and second part of the cycle (i.e., increasing and decreasing resistivity of the upper and lower layers, respectively) seem more pronounced in the first monitoring year (Figure 6). This could probably be related to uncommonly dry conditions preceding the start of the ERI monitoring and thus larger changes caused by groundwater recharge to the system.

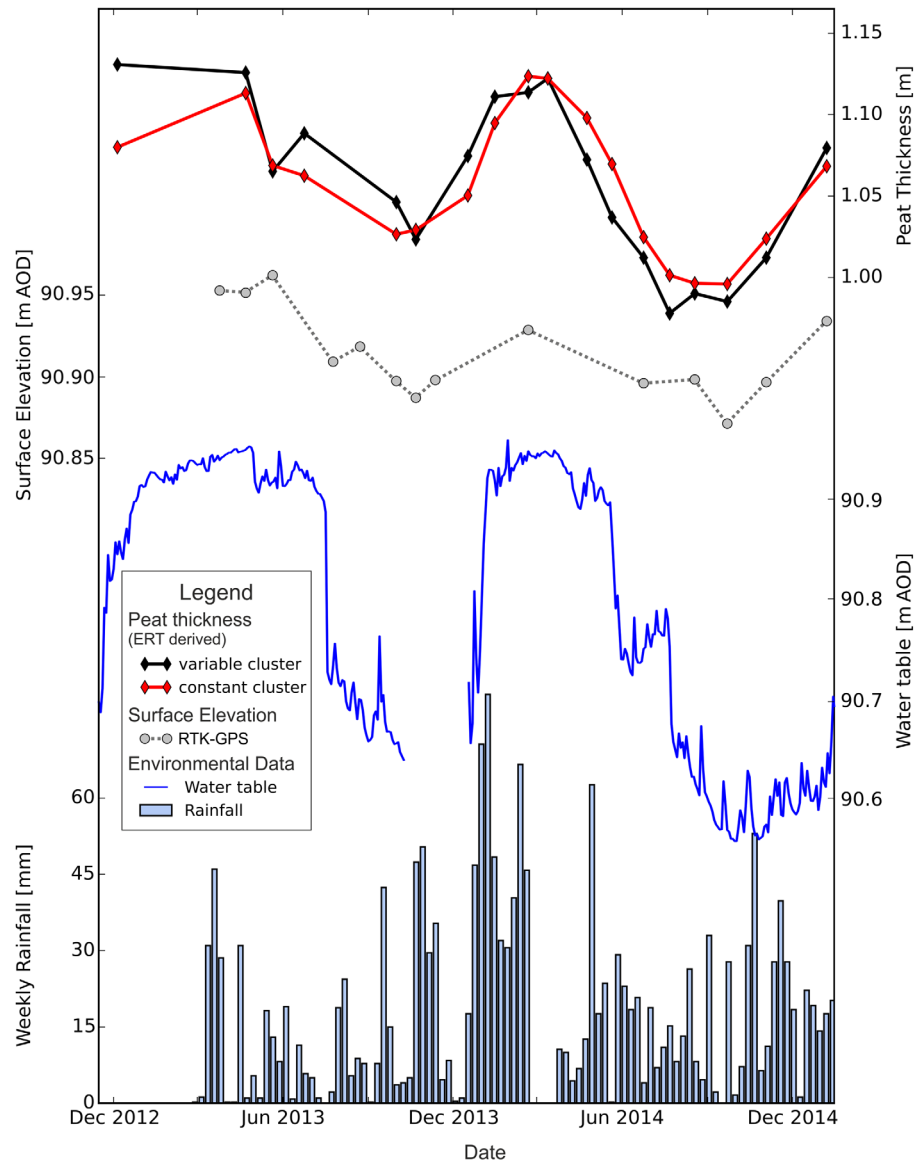
### 3.3. Shrink-Swell Characterization

Episodes of shrink-swell indicated by surface elevation correlate with drying-wetting cycles (Pearson's  $r = 0.52$ ; Figure 7). RTK-GPS measurements demonstrate a change in average topographic elevation (and thus peat thickness) of up to 0.10 m in the northern meadow. Inspection of the boundary extraction of peat thicknesses highlights a similar relationship, with periods of high and low peat thicknesses corresponding to water table highs and lows, respectively. Good correlation is indicated by a Pearson's  $r$  correlation coefficient of 0.80. Results of the two approaches to extract the peat thickness (temporally variable and constant clusters) are comparable. These show the average peat thickness cycling between 0.98 m and 1.12 m, in line with the topographic range.

## 4. Discussion

### 4.1. Imaging Plant Transpiration or Microbial Activity in the Upper Peat Layer?

The decreasing resistivity of the upper peat layer from winter to spring is most likely to be caused by changes in pore water conductivity. This is because (1) the moisture content remained at a constant high during the two imaged seasons, thus indicating fully saturated conditions, and (2) temperature variations,

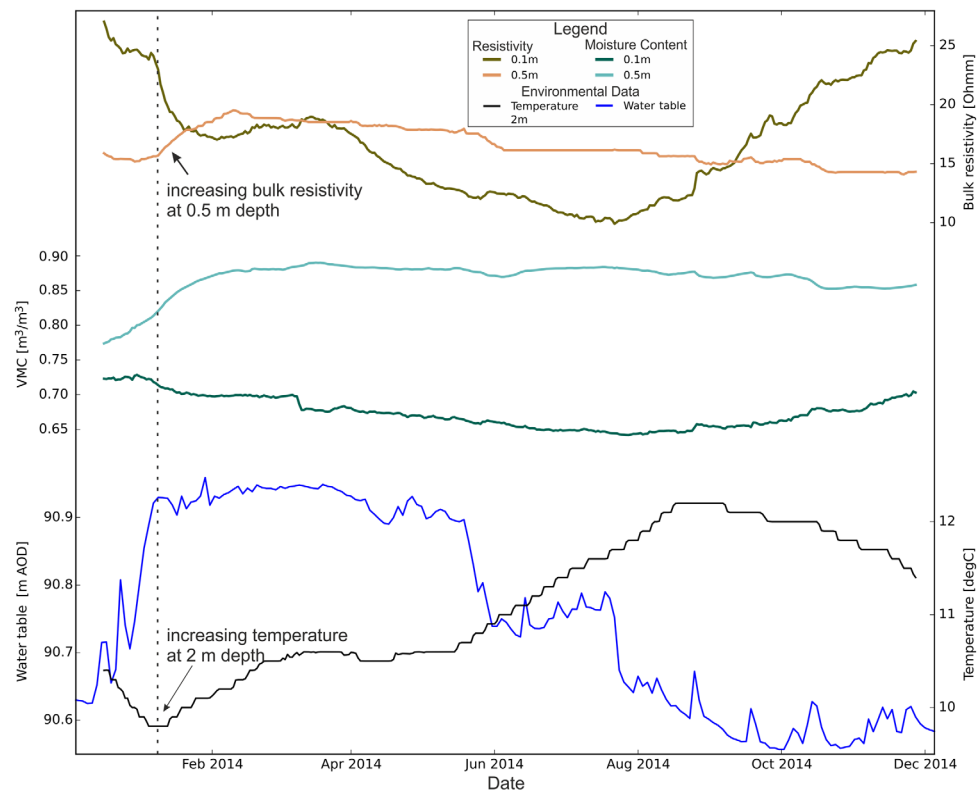


**Figure 7.** Comparison of average peat thickness estimates. Peat thickness was derived from boundary extraction from the ERI models. Surface elevation (determined using RTK-GPS) is representative for changes in peat thickness, assuming that the peat-gravel interface is constant in space. Note the good correlation between peat thickness and water table.

which would have resulted in the inverse trend anyhow, were corrected for. Pore water conductivity dynamics may have been induced by vegetation or microbial activity.

Plant transpiration may result in the accumulation of superfluous minerals, which would not readily be flushed given the relatively low permeability. Vegetation around the arrays is dominated by deciduous reed sweet-grass (*Glyceria maxima*) and lesser pond-sedge (*Carex acutiformis*). These species die back during winter; hence, photosynthesis (and transpiration) will only initiate following the onset of new leaf growth in spring. Both species root sufficiently deeply to uptake water from the majority of the peat profile [Atwell et al., 1980; USDA, 2007], which is inconsistent with the decreases in resistivity in only the upper part of the peat (Figure 8). Nevertheless, the majority of water uptake is likely to be from the near-surface given the greater density of roots near the stem and the perennial availability of water within 0.1 m of the surface negating the need for deep roots at this site.

Microbial activity in the near-surface could contribute to a decrease of the pore water resistivity through the dissolution of minerals from the organic matrix [Atekwana and Slater, 2009; House et al., 2015a]. Higher

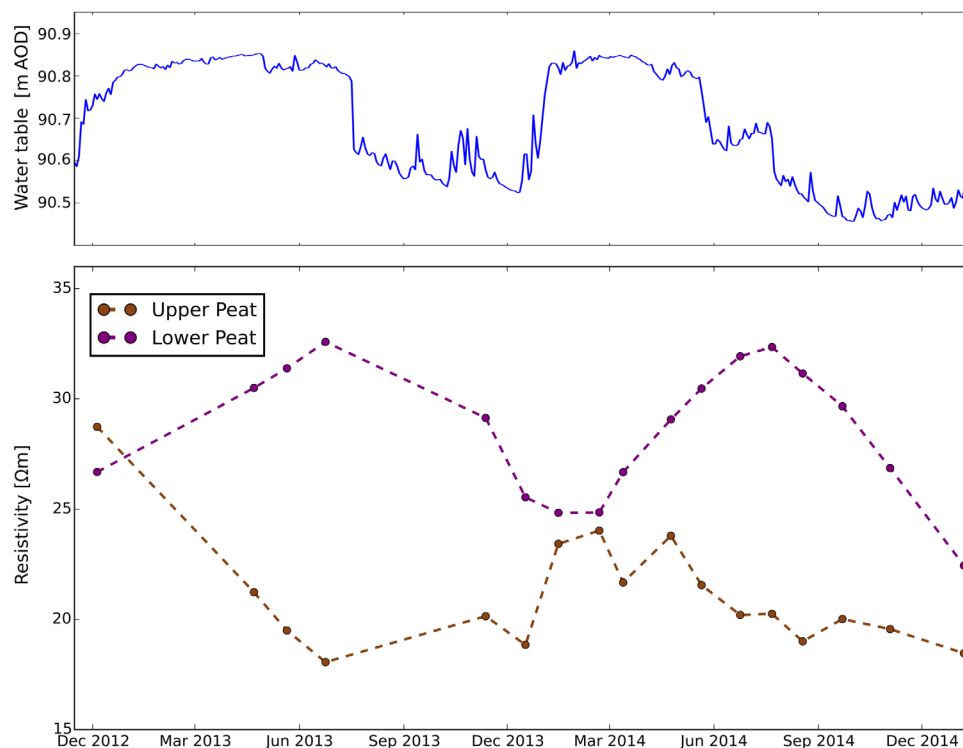


**Figure 8.** Data of sensor cluster 2 of the northern meadow. Note the sudden change in temperature in February 2014, which correlates with increasing resistivities in the deeper peat. Bulk resistivities of upper (0.1 m) and lower (0.5 m) peat show differing temporal trends which are in agreement with ERI monitoring data.

numbers of microbes are frequently observed near the surface or in close proximity to the water table [Sundh *et al.*, 1994; Daulat and Clymo, 1998; Fisk *et al.*, 2003]. This is likely to be the case here, where vegetation dieback around November produces a mass of black litter that represents a readily available carbon food source at the surface. Furthermore, the greater density of roots in this horizon would also enhance oxygen availability, as well as food supply through release of exudates containing significant quantities of carbon [Bertin *et al.*, 2003; Mainiero and Kazda, 2005].

Seasonality in microbial activity is likely to be primarily driven by rising subsurface temperatures in the spring [Picard *et al.*, 2005]. In winter, temperatures stabilize around 5°C, often termed “biological zero” where biological processes in wetlands dramatically slow down or cease [Mitsch and Gosselink, 2007]. Furthermore, exudates and oxygen release from roots in the subsurface are greatest during photosynthesis [Watson *et al.*, 1997; Kuzyakov and Cheng, 2001]. The near-surface horizon remains perennially saturated with moisture contents high, so any water table dynamics are unlikely to be a key driver of microbial activity, unlike other well-drained peatland ecosystems [e.g., Moore *et al.*, 2007].

Root respiration in the soil is also likely to be a contributing factor to near-surface resistivity dynamics, which has been demonstrated to account for around 35–50% of temperate ecosystem respiration in peatlands [Silvola *et al.*, 1996; Lafleur *et al.*, 2005]. The release and subsequent dissolution of CO<sub>2</sub> could lower pore water pH and increase mineral weathering. This would be most likely restricted to the near-surface horizon due to greater root density and relate seasonally to the start of photosynthesis and correlate with air temperatures, thereby peaking in summer [Phillips *et al.*, 2010]. However, subsurface production of CO<sub>2</sub> and other gases such as methane has been shown to increase resistivity by replacing the conductive water within the pore spaces with non-conductive gases [Slater *et al.*, 2007]. This process is most likely to cause the increase in resistivity during the summer season after the fall in water table, when free-phase gases produced in deeper peat layers are released [Bon *et al.*, 2014]. However, this fall in water table also results in significant reduction of moisture content in the upper 0.1 m, which will affect the resistivity signature to a greater degree.



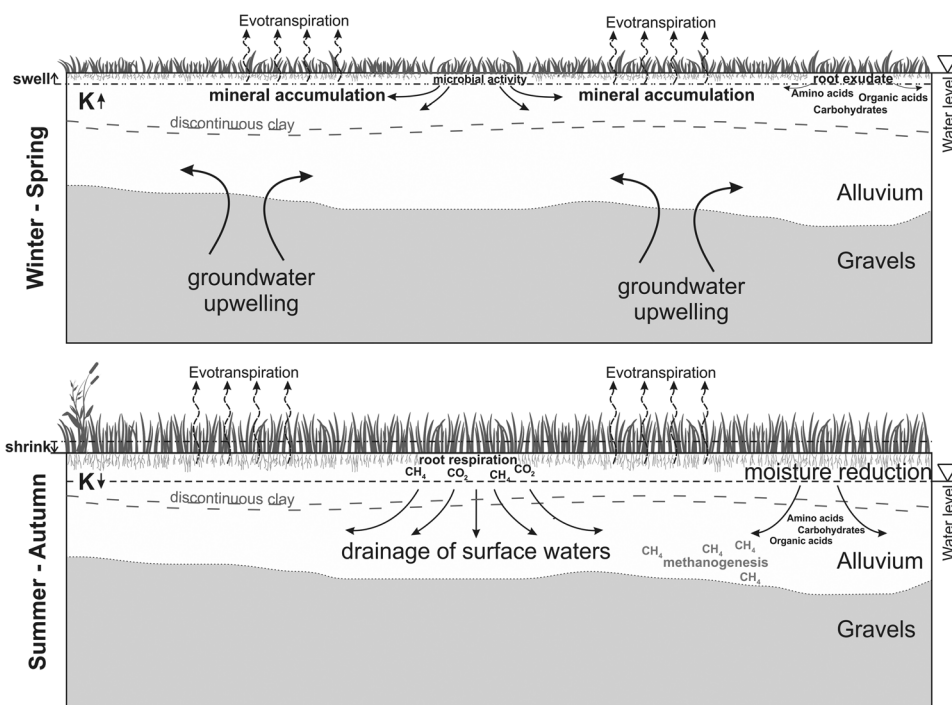
**Figure 9.** Water table and time series of mean resistivities of the upper and lower peat layer, highlighting the seasonal changes.

While all of these processes are likely to cause the imaged changes in pore water conductivity, no biogeochemical data were available to define their specific contributions. This will be investigated in further work, employing peat sampling and biogeochemical analysis, monitoring of gas production, and automated 3-D ERI monitoring of the root zone.

#### 4.2. Imaging Groundwater Exchange or Biogenic Gas Production in the Lower Peat Layer?

The strong increases in resistivity between winter and spring in the lower peat layer are most likely to be a result of the upwelling of more resistive, deeper groundwater. As this period is concomitant with maximum moisture content and the resistivity data are corrected for seasonal temperature contrasts, changes in type of pore fluid are the most likely explanation. This is evidenced by data from cluster 2 in the northern meadow (Figure 8), where a sudden rise in temperature (indicative for upwelling of deeper groundwater) coincides with increasing bulk resistivities. Hydrological observations also demonstrate a transition to positive vertical head differentials, a bulk reduction in peat electrical conductivity, and peat pore water SEC becoming more akin to the gravels and chalk over this period (Figure 5). The fact that these resistivity dynamics are restricted to only the lower peat layer again indicates that the thin chalky clay is significant in minimizing water exchange. However, following riverine weed cutting and the lowering of surface water stage encircling the site, the shallow conductive waters that have developed in the shallow peat layer appear to drain into the deeper peats. This is suggested by an increase in resistivity of the upper and decrease in resistivity of the lower layer (Figure 9). Therefore, groundwater exchange between the layers may be possible when there is a large hydraulic head differential, such as post weed cutting (Figure 10).

Another possible explanation for the deeper peat resistivity variations is biogenic gas production and consequent dynamics of free-phase gases (FPG). While an accumulation of FPG will be indicated by an increase in resistivity, FPG release will show a decreasing signature [Slater *et al.*, 2007]. Accumulation is likely to take place during the spring and summer, with release most likely taking place after the drop in water level in response to the riverine weed cutting. This generally follows the pattern imaged in the lower layer (Figure 9). Accumulation and subsequent release of FPGs could also be a driver for the imaged changes in peat thickness (Figure 7). However, no short-term oscillations in hydraulic head that are indicative for FPG



**Figure 10.** Conceptual diagram showing the processes characterizing periods of groundwater upwelling and drainage of surface waters, as well as peat shrink-swell and its effect on the hydraulic conductivity  $K$ .

ebullition were observed at site. The peatland studied by *Parsekian et al.* [2011] showed the majority of gas production to occur below 2.5 m depth and be dependent on the vegetation type. Fen-dominated peatlands, as at the Lambourn observatory, showed a homogeneous FPG distribution with FPG occupying between 0 and 7% of the pore space. Thus, the effect of FPGs on the imaged resistivity variations is likely to be limited. However, no direct measurements of released gases were obtained, but will be part of a forthcoming study.

Also expansion and compaction of the peat are likely to influence its resistivity signature, due to increasing and decreasing pore space, respectively. While peat thicknesses are shown to correlate well with changes in water table, the resistivity signature of the lower peat is, in comparison, out of phase. Thus, peat shrink-swell is unlikely to have a major influence on the imaged resistivity variations.

The decreasing resistivities imaged in the gravels just below the most pronounced indications of upwelling in the lower peat layer in spring 2014, may have been caused by the elevated temperature of upwelling groundwater. Although the presented data are corrected for seasonal temperature variations, localized temperature anomalies of up to  $5^{\circ}\text{C}$  (temperature contrast between peat and gravel during winter; Figures 5 and 8) would induce resistivity variations of up to 10%. The very low gravel resistivity in March 2014 coincides with the smallest vertical head differentials, and may therefore indicate a continued transfer of conductive surface waters through the lower peat layer into the gravels. The discontinuity imaged within the gravels ( $16\text{ m} < x < 21\text{ m}$ ) is likely to be an area of reduced porosity, where, due to the smaller available pore space, changes in pore water conductivity are less pronounced than in the surrounding higher porosity gravels. However, as the sensitivity of ERI reduces with increasing distance from the electrodes (and thus depth), uncertainties of the processes imaged in the gravels are higher than in the peat layers.

### 4.3. Imaging Shrink-Swell

As shown by *Price and Schlotzhauer* [1999], shrink-swell behavior of the peat shows a clear correlation with the peat water table which is therefore considered the main driver for this process. RTK-GPS measurements showed a cyclic behavior with average topographic heights varying by about 0.10 m, being greater than

the instrument precision that was determined to be 0.03 m. This indicates a variation of 7–10% of the total peat thickness, values that are comparable to studies of peatlands in the Netherlands [Hoogland *et al.*, 2012; Querner *et al.*, 2012] and Canada [Price and Schlotzhauer, 1999]. Very similar ranges were found for ERI derived peat thicknesses. The smooth variation of these in time is likely to be caused by the data inversion methodology, which favors smooth temporal changes. Note that the model cell size at the depth of the peat interface is larger than the imaged variations of peat thickness (0.25 m and 0.14 m, respectively). Therefore, estimation of the small thickness variation was only possible due to the sharp geological interface between peat and gravels that also showed a very strong contrast in resistivity (20  $\Omega$ m to 150  $\Omega$ m). Thus, resistivity values of the model cells close to the interface will show significant variations with changing layer thickness, and therefore a trackable boundary.

For only 1% change in peat thickness Price [2003] found a change in hydraulic conductivity of two orders of magnitude. Hence, the 7–10% change observed at the Lambourn observatory could cause even larger changes, with decreasing hydraulic conductivity in summer due to compaction and increasing conductivity in winter caused by peat expansion. This will not only restrict moisture uptake by plants in summer due to a decreased vertical moisture fluxes and thus affect the vegetation community, but also highlights the need to incorporate seasonally varying hydraulic parameters in hydrologic models of peatlands.

The recent development of automated remote monitoring systems and analysis tools offers the potential to monitor peat thickness variations at high temporal resolution without the need of additional surface elevation monitoring instrumentation. Thus, less instrumentation needs to be installed and study sites can be operated more economically.

## 5. Conclusions

The application of time-lapse geoelectrical imaging in an ecologically sensitive riparian wetland has revealed a two-layer hydrological system within the peat, which is considered a result of a thin intermediate chalky clay layer. Layered groundwater systems are likely to be widespread in riparian wetlands given the varied depositional history in floodplain environments. Resistivity dynamics within the two layers follow separate annual cycles relating to the dominating influence from different processes.

The upper layer is characterized by a resistivity reduction in spring (March–June) when saturated, resulting from processes affecting the pore water conductivity, including microbial activity, plant transpiration, and root respiration. Following riverine weed cutting around June/July, and a drop in the encircling surface water stage, resistivities increase as water drains from the upper peat and the moisture content decreases. In the lower layer, resistivities increase around February due to the influx of more resistive groundwater from the underlying gravels during the seasonal groundwater high. Following the cessation of deeper groundwater input, the conductive water in the upper peat then drains into this lower layer following the weed cutting, thereby reducing the resistivity.

The additional sensor data facilitated a detailed understanding of the geophysical data. A sole, independent interpretation of either data would not have provided the same level of detail to investigate the hydrological processes. Notably, the main benefit of the geoelectrical data is its noninvasive, spatial nature, which was able to visualize the dynamic vertical and lateral extent of multiple hydrological processes and guide the installation of additional sensors to investigate them.

For the first time, we have shown that applying automated interface detection algorithms to time-lapse ERI models can reveal the migration of a geological interface in the vertical direction. This is demonstrated through the observation of shrink-swell within the peat, which is calculated to be similar to that observed through topographic changes at the surface. The conjunction of time-lapse geoelectrical imaging with automated interface detection algorithms could be useful for studying ground movement more broadly, including phenomena such as subsidence, ground heave, or landslides. Whilst ground movement can easily be observed using current surface based techniques such as laser scanning and time-lapse cameras, ERI could provide data with comparable accuracy and information on where in the subsurface movements are induced.



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