

**PHYSICAL CHARACTERIZATION OF TWO FRACTURED
SEDIMENTARY ROCK AQUIFERS IN NEW BRUNSWICK, CANADA
WITH EMPHASIS ON THE DEVELOPMENT OF SELF-POTENTIAL
METHODS**

by

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ABSTRACT

Heterogeneity and anisotropy associated with fractured rock aquifers can make predicting fluid flow pathways difficult using traditional hydraulic testing methods alone. Surface-based self-potentials (SP) measured during pumping and hydraulic testing have been shown in this thesis to provide valuable information used to infer water table drawdown, preferential flow paths and hydraulic properties.

To provide a basis for investigating the SP approach in fractured rock, conceptual models were first developed for two separate aquifers. Fracture patterns, combined with pump test and geophysical logging data allowed for flow directions to be assessed within a confined fluvial sandstone-shale aquifer underlying the Springdale wellfield in the Carboniferous Moncton Subbasin. In the Ordovician to Silurian Matapédia Basin, fracture characteristics were measured within the folded turbidite sequences underlying the Black Brook Watershed, and combined with hydraulic conductivity estimates derived from packer testing to provide a better understanding of heterogeneity responsible for anisotropic groundwater flow conditions.

Both fractured aquifers were subsequently investigated through SP monitoring and numerical modelling. Transient SP signals recorded during pumping in the Springdale wellfield, combined with measurements of the electrokinetic voltage coupling coefficient, allowed spatial and temporal variations in drawdown to be inferred below electrodes positioned around the pumping well. SP-derived drawdown was fitted to the Theis model to obtain transmissivity and storativity estimates at electrode locations. Numerical modelling showed surface SP measurements to be an excellent proxy for hydraulic head at the top of an underlying confined aquifer even in the presence of hydraulic and electrical

heterogeneities. SP signals recorded in the Black Brook watershed during a constant head injection test at a depth of 44 m were used to infer azimuthal anisotropy in fluid flow in directions consistent with dominant fracture set orientations. Numerical modelling showed that fracture transmissivity, length, and frequency all contribute to the magnitude and shape of SP anomalies recorded on surface resulting from flow from the highly transmissive fractures into the surrounding rock matrix.

Results from the SP monitoring approach, combined with the conceptual aquifer models, demonstrated an inexpensive and non-invasive means of assessing water table drawdown pattern, preferential flow directions, and hydraulic properties estimates without the need for additional wells.

DEDICATION

To my wife, Elizabeth, and daughter Mabel

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Chapter 1

Introduction

The presence of fractures in bedrock aquifers can significantly influence local- and regional-scale fluid flow. Fractures can act as fluid flow pathways, or can act as barriers depending on whether the fracture is open or sealed. Characterizing flow through fractured rock is considered one of the most challenging problems that hydrogeologists face (Neuman 2005). Understanding the distribution and hydraulic properties of fractures in bedrock aquifers is important for solving hydrogeological, geotechnical and environmental problems. Fractured rock hydraulic properties can be inferred through hydraulic testing methods in boreholes, such as pump tests and isolated packer tests, and by direct observation of fractures through core logging and in situ televiewer imaging. Unfortunately, detailed inspection of fracture characteristics is seldom completed due to the extensive time required to develop a comprehensive fracture model and to measure their hydraulic properties. As a consequence, fractured aquifer flow and transport models are often developed with sparse data which may oversimplify their heterogeneities or anisotropy leading to erroneous results. To enhance data coverage, non-invasive geophysical methods are increasingly being integrated with geological and hydrogeological data to help characterize and understand the distribution of hydraulic properties and flow pathways through fractured aquifer systems (Binley et al. 2015).

The primary motivation for this research is to integrate surface-based self-potential (SP) measurements with information from borehole geophysical logging, geological mapping and hydrogeological testing to better characterize fracture networks and

preferential flow pathways within fractured bedrock. The associated data can then be incorporated into numerical flow and transport models allowing for more reliable wellfield assessment and improved water well placement. Although this research focuses on aquifers used for drinking water resources, these methods and approaches can also be adapted to characterize fractured rock and flow pathways for contaminant transport studies, and at greater depths, for studying oil and gas hydrocarbon reservoirs, and underground sites for nuclear waste disposal.

The research focuses on two bedrock aquifers in New Brunswick, Canada (Figure 1.1) where information on the distribution of fractures and their hydraulic properties was limited: (1) the Springdale Wellfield, and (2) the Black Brook Watershed. The Springdale Wellfield is situated along the southern margin of the Maritimes Basin, within the fault-bounded Carboniferous Moncton Subbasin. The Maritimes Basin is the largest sedimentary basin within the Appalachian belt in eastern Canada, and underlies central and eastern New Brunswick, all of Prince Edward Island, and portions of Nova Scotia, and Newfoundland. In proximity to the wellfield, the basin is well-known for potash mining and gas development, both of which pose a risk to the shallow groundwater system. The Black Brook Watershed (BBW) is located in northwestern New Brunswick. In this region, municipal and residential drinking water is primarily sourced from the underlying fractured bedrock of the Matapédia Basin which extends throughout most of northwestern New Brunswick, the Gaspé region of Québec, and into northern Maine, USA. Much of the BBW lies within an economically important agricultural belt of the upper St. John River Valley where potatoes are the dominant crop. Both of these fractured aquifers are vulnerable to anthropological influences, which are inferred to have impacted the quality of drinking

water or supply. Specifically, residents near the Springdale wellfield, in the town of Penobsquis (pop. 1,479 in 2006 census), New Brunswick, experienced both lowering of water levels and significant reductions in the yields of their private wells (Jacques Whitford 2005). Although not proven, lowering of the water table was speculated to be related to potash mining in the area. Within the Black Brook Watershed intensive agricultural practices have prompted several studies to assess their effects on surface and groundwater quality (e.g. Milburn 1996; Yang et al. 2010; Chow et al. 2011) and assess aquifer flow pathways and recharge and discharge zones (Zhang et al. 2013; Li et al. 2014). In 2003 the village of Saint-Andre, New Brunswick, located within the BBW, relocated their water supply well due to elevated nitrate levels with concentrations occasionally exceeding the 10 mg/l $\text{NO}_3\text{-N}$ maximum acceptable concentration for drinking water (Health Canada, 2012). Since both of the fractured bedrock aquifers are positioned within regional-scale sedimentary basins, the information obtained and methodologies applied in this thesis at these two locations may be similarly applied to groundwater studies elsewhere in these sedimentary basins where the bedrock exhibits similar regional-scale characteristics.

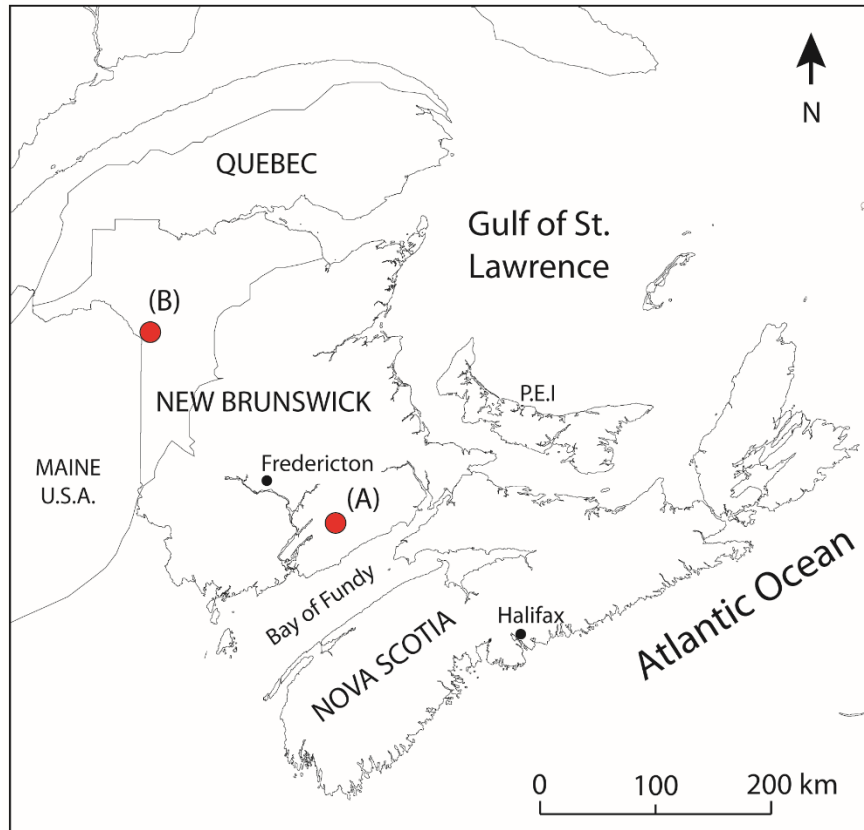


Figure 1.1 Location of the two fractured bedrock aquifers in New Brunswick (A) Springdale Wellfield; and (B) Black Brook Watershed.

1.1 Fracture characterization

Quantifying groundwater flow directions and magnitudes requires an understanding of the geometry and hydraulic properties of the fracture network, in addition to spatial distribution of hydraulic head (Singhal and Gupta 2010). However, the distribution and character of discrete fractures and fracture sets in the bedrock are often poorly understood. To understand the fracture network, it is recommended that direct fracture observations be made from either borehole televiewer images or orientated core, or outcrop mapping and integrated with other borehole geophysical logging and hydraulic testing data. All of this

information can then be incorporated in the development of a hydrogeological numerical model.

In this research, fracture characteristics were routinely documented by measuring orientation (strike and dip), frequency, aperture, and mineral in-filling of fracture surfaces on outcrop exposures and through borehole logging. Both optical and acoustic televiewer logging images allow for observation and measurement of fracture characteristics, such as their character, spacing, apparent aperture, orientation, in-filling, and weathering along a borehole wall that can be statistically analyzed to determine the number and geometry of fracture sets (e.g. Williams and Johnson 2004). Outcrop exposures also provided direct observation of fractures characteristics that were incorporated with the borehole information (c.f. Lemieux et al. 2006; 2009). Fracture spacing was determined as the perpendicular distance between adjacent fractures for a particular fracture set, which provided an indication of the mean frequency of occurrence of this fracture set through the bedrock. Using borehole televiewer images, the apparent fracture apertures were discerned as the perpendicular distance between the walls of an open fracture. Data gleaned from borehole or outcrop fracture mapping were integrated with other standard geophysical logs (i.e., gamma ray, self-potential, single point resistance), core samples, flowmeter and hydraulic testing measurements for a more complete interpretation of the fractured aquifers.

1.2 Flow characterization

In both aquifers, a number of hydraulic testing methods were also implemented for estimating hydraulic conductivity. Constant rate pumping tests were performed in a well at the Springdale wellfield while water levels were monitored in a network of nearby wells.

This approach is considered a standard for measuring spatial and temporal variations in hydraulic head to determine the transmissivity and storativity of an aquifer (e.g. Domenico and Schwartz, 1990). Changes in hydraulic head recorded in a network of monitoring wells can be used to interpret an equivalent transmissivity and storativity over a large volume of rock located between the hydraulic source and the monitoring location, and if enough wells are placed around the pumping well the degree of anisotropy may be determined.

Difficulties may occur in interpreting drawdown curves, as a consequence of heterogeneity and anisotropy in fracture networks, permeable or impermeable boundaries, or wellbore storage effects (e.g. Driscoll 1986; Renard et al. 2009).

In the Black Brook watershed, standard packer test methods were used to provide hydraulic properties of fractured rock within four vertical wells penetrating the aquifer. The zone of influence for packer testing is typically limited to within tens of meters from the well (e.g. Singhal and Gupta 2010). The packer testing approach has been widely used for assessing hydraulic conductivity for hydrogeological, geotechnical, and waste management studies (e.g., Louis 1972; Doe and Remer, 1981; Gale, 1982; Haimson and Doe, 1983; Lapcevic, 1988; Novakowski and Bickerton, 1997).

Although both pumping and packer testing provide similar hydraulic conductivity, transmissivity, and storativity values for fractured rock, values obtained through pumping tend to be higher in comparison (Singhal and Gupta 2010). This difference is related to packer testing being sensitive to hydraulic properties in the vicinity of the test interval, whereas pumping tests are sensitive to the entire zone of influence extending around the pumping well. In order to characterize hydraulic properties of discrete fractures, or fracture zones, information from isolated packer testing is preferred. However, for the purpose of

developing groundwater supply, hydraulic properties from pumping tests are often preferred.

One acknowledged limitation of interpreting pumping or packer tests in fractured rock is that wells are usually poorly distributed due to the high cost of developing an extensive network of wells. A wellfield with a limited distribution is likely to result in errors in determining the shape of water table and the estimation of hydraulic properties through a fractured aquifer. It would be desirable to increase the spatial density of monitoring stations to better recover the spatial distribution of aquifer properties.

1.3 Self-potential (SP) monitoring

One approach of particular interest to the hydrogeological community is the self-potential (SP) method which refers to a passive geophysical technique that measures electric signals naturally generated in the subsurface as a result of coupling between electrical and non-electrical flows (e.g. hydraulic, thermal, chemical; see Sill 1983; Revil et al. 2006; Revil and Jardani 2013; Jouniaux et al. 2009; Jouniaux and Ishido 2012). In the context of non-equilibrium thermodynamics (e.g. Onsager 1931; de Groot and Mazur 2013), coupled flow theory assumes that the fluxes (e.g. electric current density or fluid flux, etc) and the driving forces (e.g. electric potential gradient or hydraulic gradient, etc) are linearly related, resulting in a system of linear equations linked by a matrix of coupling coefficients.

Although there is an abundance of research concerning SP monitoring of groundwater flow in the subsurface, relatively few studies have focused on flow through fractured bedrock aquifers. With that in mind, a secondary motivation of the research was to demonstrate how monitoring and modelling of SP signals can provide useful information to help interpret groundwater flow in fractured rock and to estimate hydraulic properties.

The most attractive aspect of the SP method for the purpose of monitoring groundwater flow is that equipment requirements and installation are inexpensive and non-invasive as compared to drilling and installing a network of monitoring wells. A large network of SP electrodes can be installed on the surface to enhance spatial coverage of the monitoring system. Data recorded by the SP method are measured as the voltage difference between two non-polarizing electrodes connected to a voltmeter. By connecting the multi-electrode array to a multiplexing high-impedance voltmeter, transient self-potential data can be recorded relative to a reference electrode over the duration of a hydraulic test. To ensure high accuracy in the SP measurements voltmeters should have an internal impedance at least 10 times higher than the impedance of the ground between the measurement and reference electrode (e.g. Revil et al. 2012). Non-polarizing electrodes, such as Pb/PcCl₂ electrodes, are preferred because they exhibit long term stability and have a lower dependence on temperature variations between electrodes in a multi-channel electrode array (Petiau, 2000).

1.3.1 Electrical double layer

In general, the main contributors to SP signals in aquifers are electrokinetic processes driven by hydraulic gradients (e.g. Fitterman 1978; Sill 1983; Fournier 1989). These SP signals are associated with electrical convection currents generated as a consequence of coupling between flows of groundwater and electrical charge. In a water-saturated porous or fractured medium, an electrical double layer forms at the rock-water interface consisting of a first layer of ions tightly adsorbed to the rock, known as the Stern Layer, and a second layer of ions, known as the diffuse layer, attracted to the surface charge by the Coulomb force. A conceptual model illustrating the electrical double layer formed at the rock-water

interface is presented in Figure 1.2. The solid surface of the rock interface exhibits a net electrical charge that can be either positive or negative, dependent on surface chemistry and pH conditions. For the pH and chemical conditions present in most freshwater aquifers, the solid surfaces typically exhibit negative charge associated with loss of a proton from the surface chemical groups (e.g. silanol group $\text{SiOH} \leftrightarrow \text{SiO}^- + \text{H}^+$). The negative charge of the solid surface attracts positive ions and repels negative ions from the pore fluid resulting in an adsorbed layer (Stern layer) and a more loosely bound diffuse layer of positive ions (e.g. Ishido and Mizutani 1981; Revil et al. 1999; Revil and Jardani 2013). This leads to an excess of positive ions (or depletion of negative ions) concentrated adjacent to the pore wall relative to the bulk pore fluid. The excess of positive charge decreases exponentially away from the solid surface towards the center of the pore where the fluid is assumed electrically neutral. The boundary between the Stern and the diffuse layer is defined as the shear plane (or slip plane). The potential difference between the solid surface and the shear plane is termed the zeta potential (ξ) and is important in determining the magnitude of electrokinetic effect. Under fluid flow conditions, positively charged ions within the diffuse layer are free to move in the direction of the pore fluid resulting in the establishment of convection currents J_k . The movement of positively charged ions in turn leads to the development of an electrical potential gradient driving conduction currents J_c in the opposite direction (Revil and Leroy, 2004). Since convection currents are the result of fluid flow they are restricted to the saturated portion of the aquifer. However, the resulting conduction currents are free to flow through the entire subsurface along a path of least resistance (incorporating bedrock resistivity structure) and therefore may be measured on the ground surface at a distance above the saturated aquifer.

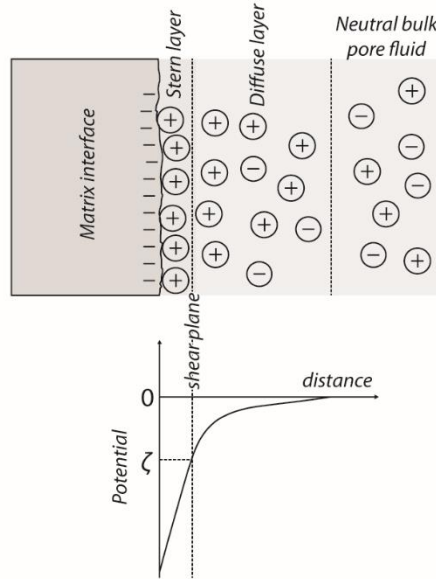


Figure 1.2 Conceptual model of the electrical double layer in the vicinity of a rock-water interface, which includes the Stern layer, diffuse layer, and neutral bulk pore fluid. The electric potential decays exponentially as a function of distance from the rock-water interface.

1.3.2 Application of SP method to hydrogeological problems

The application of the SP method in this thesis relates to measurements of SP responses to hydraulic gradients and groundwater flow induced by pumping or hydraulic testing in a well. The coupling mechanism responsible for the SP signal is electrokinetic in nature, where the primary flow is movement of water driven by hydraulic gradients associated with pumping or injection of water in an aquifer. In a case where water is being removed by pumping in an unconfined aquifer with uniform electrical conductivity, the fluid sink and induced hydraulic gradients around the well are responsible for generating an SP response at surface that is expected to be proportional to the drawdown of the water table (e.g. Rizzo et al. 2004). Figure 1.3a shows an example of a hypothetical heterogeneous aquifer where the drawdown cone is preferentially elongated parallel to a transmissive set of

fractures. Following the previous discussion of the electric double layer, water flowing towards the pumping well results in a positive charge accumulation near the well and a positive SP response measured directly above the well (Figure 1.3b). In an aquifer where convection currents are exactly balanced by conduction currents in the opposite direction, the SP measured on surface is linearly related to the drawdown of the water table (hydraulic head) through the voltage coupling coefficient (V/m) (e.g. Rizzo et al. 2004; Figure 1.3c). This parameter can be readily measured in laboratory experiments, or derived empirically through field measurements using a monitoring well and electrode pair.

It is also important to understand how a heterogeneous distribution of aquifer properties influences the development of SP sources. The theoretical basis for rigorous forward modelling of SP in terms of convection current induced by a primary hydraulic gradient, resulting in the generation of conduction currents was first presented by Sill (1983). Apart from primary sources associated with fluid sources and sinks (e.g. pumping or injection wells), this theory shows that secondary SP sources are generated by groundwater flow perpendicular to boundaries where there is a contrast in (a) hydraulic conductivity, (b) electrical conductivity, or (c) electrokinetic coupling coefficient. Because the electrokinetic coupling coefficient is also dependent on several other parameters, the latter secondary source may also be expanded to include dependent parameters of the coupling coefficient, such as fluid dielectric constant, zeta potential, viscosity, or formation factor (Wurmstich 1995). The magnitude of the secondary SP source depends on the magnitude of the parameter contrasts and hydraulic gradients driving flow across the boundary.

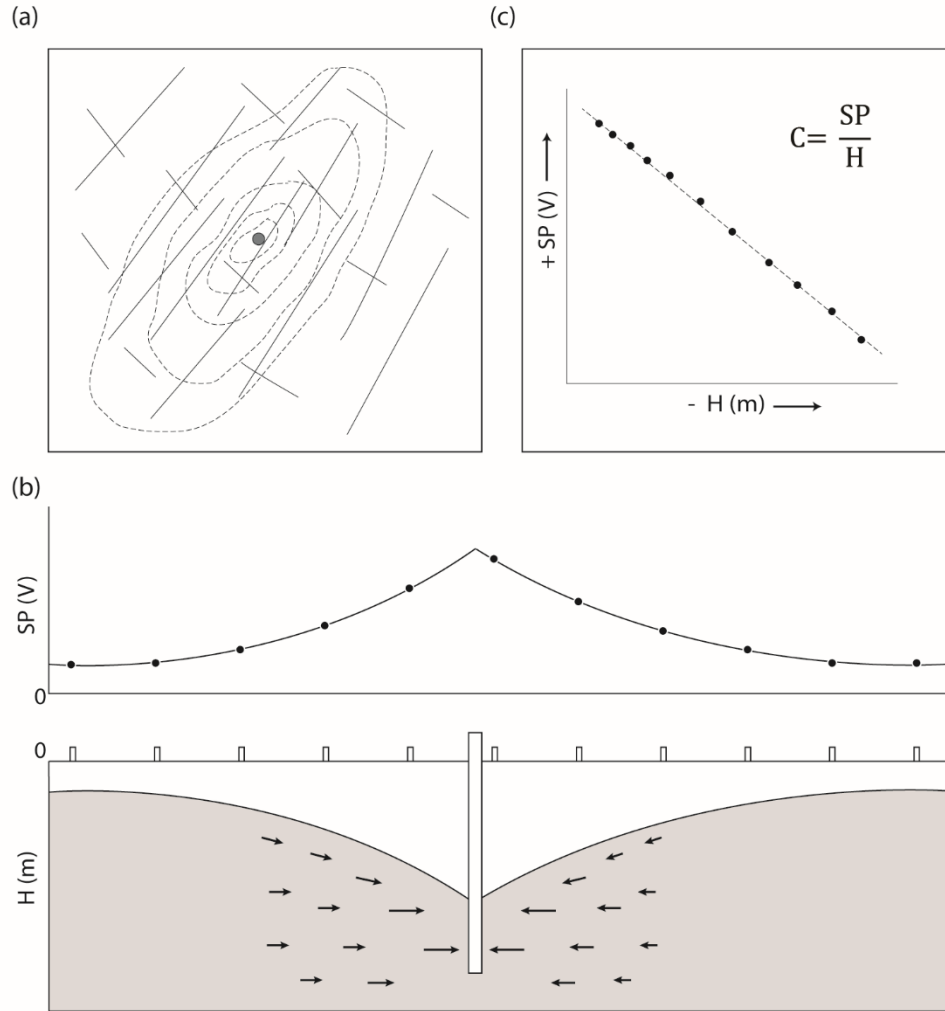


Figure 1.3 Conceptualized aquifer presents the relationship between hydraulic head drawdown in a heterogeneous unconfined aquifer with uniform electrical conductivity and the expected SP response measured on the surface during a pumping test. (a) Plan view of a heterogeneous fractured aquifer with drawdown cone extending parallel to strike of hydraulically conductive fractures, (b) vertical section through aquifer showing SP response associated with drawdown, where arrows represent fluid flow, and (c) SP expected on surface plotted against hydraulic head during the pumping period where the slope of the linear correlation represents an apparent voltage coupling coefficient.

The SP method has been studied for a variety of hydrogeological applications including approximating the shape of the water table, inferring hydraulic aquifer properties, and determining preferential flow directions, both in saturated aquifers (e.g. Bogoslovsky and Ogilvy, 1973; Fournier 1989; Birch 1998; Darnet et al. 2003; Revil et al. 2003; 2004;

Suski et al, 2004; Rizzo et al. 2004; Jardani et al. 2006; Linde et al. 2007; Maineult et al. 2008; Malama et al. 2009; Straface et al. 2007; 2010; Revil and Jardani 2013) and in unsaturated media (e.g. Aubert and Atangana 1996; Thony et al. 1997; Aubert et al. 2000; Perrier and Morat, 2000; Doussan et al., 2002; Darnet and Marquis, 2004; Chidichimo et al. 2015).

Early experimental SP data recorded by Bogoslovsky and Ogilvy (1973) during pumping at steady state in an unconfined aquifer showed that under normal conditions where the bedrock has a negative electrokinetic coupling coefficient, a significant positive SP anomaly could be generated near a pumping well. They further showed that SP anomalies could be used to estimate the radial extent of drawdown in the vicinity of a pumping well with a limited number of monitoring wells installed. Using recent advances in interpretation procedures, Revil et al. (2003), based on the earlier work of Fournier (1989), analyzed the SP response measured by Bogoslovsky and Ogilvy (1973) and obtained estimates for the depth and shape of the water table in reasonable agreement with the reported piezometer measurements. Darnet et al. (2003) performed an SP inversion of the same dataset to deduce the hydraulic conductivity, electrokinetic coupling coefficient and thickness of the homogeneous unconfined aquifer. By also incorporating resistivity structure into their inversion, they showed clearly that the SP anomaly on surface would be influenced by the flow of electric currents from depth to surface along the conductive well casing. Rizzo et al. (2004) presented a linear relationship between the SP signal and hydraulic head around a pumping well in a confined silty sand aquifer during the recovery phase following a pumping test. Transient SP data recorded using an array of electrodes on surface above the aquifer were converted to a hydraulic head distribution and combined with direct measurement of water levels in a small network of piezometer wells. By assuming a

homogeneous and unconfined aquifer under steady-state, they interpreted the SP data to determine an average hydraulic conductivity (Rizzo et al. 2004). Straface et al. (2007) used the transient SP data from Rizzo et al. (2004) to first invert the transmissivity and storativity parameters of the Theis equation at monitoring well locations, then apply a sequential linear estimator (i.e. SLE) (Yeh et al. 1996) to determine the spatial distribution of transmissivity using the SP data. Malama et al. (2009a; 2009b) developed semi-analytical model solutions for calculating hydraulic conductivity and specific storages for confined and unconfined aquifers. In the case of the unconfined aquifer, their solutions were successfully used to calculate hydraulic conductivity, anisotropy ratio, specific storage, and specific yield using transient SP data while drawing an aquifer down at the Boise Hydrogeophysical Research Site (BHRS). Mainault et al. (2008) also showed transient SP data can be used to estimate hydraulic properties while performing periodic pump tests in a confined aquifer. Straface et al. (2010) were able to obtain average hydraulic conductivity from an unconfined large-scale laboratory model. They first determined the distribution of hydraulic head as a function of the spatial distribution of SP by kriging with external drift (i.e. KED), then inverted for hydraulic conductivity through a nonlinear least squares method. Their hydraulic conductivity estimates were similar to those determined by plotting well data on a Neuman-type curve, and were also within half-order of magnitude of values obtained using the same methodology as Rizzo et al. (2004).

Following the rigorous modelling approach of Sill (1983), several studies have modelled SP signals during hydraulic testing in both homogeneous and heterogeneous isotropic aquifer scenarios (e.g. Wurmstich and Morgan 1994; Titov et al. 2005; Minsley et al. 2006; Sheffer and Oldenburg 2007; MacAllister 2016). Wurmstich and Morgan (1994)

modeled a deep oil-saturated reservoir bounded by units with lower permeability and higher electrical conductivity in order to predict SP signal on the surface during pumping. They show that the electrical conductivity assigned to the confining units significantly influence the magnitude of SP on surface. Titov et al. (2005) evaluated the relationship between SP measured at surface and hydraulic head measured by Rizzo et al. (2004) in an underlying confined aquifer undergoing pumping, where over the course of pumping the hydraulic head level in the aquifer changed from confined to unconfined. They demonstrated that despite the aquifer being heterogeneous in hydraulic conductivity, modelling predicted a laterally consistent linear relationship between drawdown and surface SP that was proportional to the electrokinetic coupling coefficient. Minsley (2007) modelled an SP response accompanying the injection phase of an aquifer storage and recovery (ASR) process in the confined Dammam sandstone aquifer in Kuwait. The model was radially symmetric with fluid injection simulated in a 20 m interval screened at a depth of ~250m, approximately 100 m below the top of the aquifer. Modelling results predicted that the primary contribution to the SP response is from fluid injection with a negative anomaly up to ~3 mV observed above the well location. Assuming a constant injection rate over a 12 month simulation period, additional temporal and spatial changes to the SP response were caused by heterogeneities in resistivity of the aquifer as the fluid front progressed, and by a salinity contrast between the injected freshwater and the aquifer water. Sheffer and Oldenburg (2007) modeled SP in a heterogeneous aquifer while pumping. Their model assumed a homogeneous distribution of electrical properties, where only hydraulic conductivity changed between units. MacAllister (2016) presented results from field borehole monitoring and forward modelling of transient SP responses related to the

progression of the saline water front in highly fractured chalk bedrock. The fracture zones were suggested to act as the primary transmissive pathways for fluid flow, and were modeled following an equivalent porous media approach. Results modeled at an observation borehole 1.7 km from the coast showed that SP signals were related to progression of a saline fluid front driven by tidally-induced hydraulic gradients. Recent 2D discrete-dual-porosity modelling by Roubinet et al. (2016) simulated SP responses within discrete fractures and complex fractured networks to evaluate the effects of flow within fracture planes and across the boundaries into the less permeable matrix. Their results show that, under a constant pumping rate, the strongest SP magnitudes were generated along the transmissive fractures hydraulically connected to the pumping well. They also concluded that exchange of fluid between the rock matrix and fractures is necessary to generate significant SP signals associated with pumping; modelling fluid flow through fractures alone, ignoring matrix fluid flow, will lead to SP that are orders of magnitude too small.

Despite the number of SP studies with a focus on monitoring groundwater flow and assessing hydraulic properties of the shallow subsurface, the majority of these studies experimented on homogeneous unconsolidated sediments or porous bedrock conditions. Only a few interpretations considered the influence of lithology or fracture induced heterogeneity or anisotropy on the development of electrokinetic SP signals (e.g. Bogoslovsky and Ogilvy 1972; Hotzl and Merkler 1989; Fagerlund and Heinson, 2003; Ishido and Pritchett 2003; Wishart et al. 2006; Suski et al. 2008; Robert et al. 2011; MacAllister, 2016). Bogoslovsky and Ogilvy (1972) conducted early SP laboratory studies on fractured media by using a series of quartz glass plates with different infill materials. Their results showed a dependence between fracture aperture and SP signals by adjusting

the spacing between parallel plates. Hotzl and Merkle (1989) measured SP around a well while performing an injection/grouting experiment in the upper 10 m of a shallow bedrock aquifer. By comparing their results with joint orientations, they were able to show that SP measurements can be used to detect preferential flow within more permeable fracture sets. Fagerlund and Heinson (2003) used SP signals, combined with borehole water levels, to approximate the hydraulic head distribution generated by pumping in a fractured rock aquifer. Their results show clear patterns in the head distribution contours which are used to infer flow anisotropy related to the geometry of a fracture network. Wishart et al. (2006) showed that flow through a network of hydraulically transmissive fractures in a laboratory fractured block model was responsible for generating azimuthal variations in SP signals related to preferential flow directions. Their analysis revealed that anisotropy in their SP response was coincident with strike direction of fracture sets, and that an increase in the SP magnitude was proportional to an increase in flow rate by higher hydraulic gradients. Using the same methodology, Wishart et al. (2008) investigated four fractured rock aquifers to demonstrate the linear relationship between SP magnitude and frequency of fractures along the same trend. Suski et al. (2008) used borehole SP logging to detect and characterize hydraulically active fractures along a deep borehole in a fractured carbonate aquifer. Their analysis incorporated geochemical and borehole geophysical data to relate electrokinetic SP signals to flowing zones within the fractured network. Robert et al. (2011) acquired electrical resistivity tomograms and SP data to assess natural groundwater flow pathways through carbonate fracture zones. They reported that negative SP anomalies and low resistivities were later confirmed to be associated with high yielding fracture zones by drilling and pumping. In fractured rock, SP effects related to electrical and hydraulic

heterogeneity are often investigated by treating variable fracture densities as equivalent porous media with spatially varying hydraulic and electrical conductivities (e.g. Ishido and Pritchett 2003; MacAllister 2016; DesRoches and Butler 2016). Alternatively, recently researchers have started to model cumulative SP fields caused by the exchange of water between fractures and matrix in discrete fracture networks (e.g. Roubinet et al., 2016; and Chapter 5 of this thesis).

1.3.3 SP instrumentation

The SP measurements reported in this dissertation were acquired using a network of non-polarizing Pb-PbCl₂ electrodes constructed based on the design of Petiau (2000). Figure 1.4 shows an example of the non-polarizing SP electrode consisting of a lead wire in contact with a lead chloride salt in a kaolinite mud. These electrodes exhibit low noise levels, low temperature influences, and minimal polarization making them stable over extended periods of time compared to other metal-salt electrode designs (Petiau 1980; 2000).

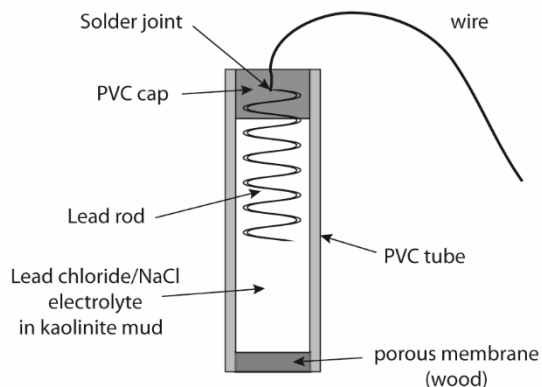


Figure 1.4 Schematic of the non-polarizing SP electrode construction based on design from Petiau (2000)

The network of electrodes was interfaced with a Campbell Scientific AM32B multiplexer connected to a CR1000 data logger with an input impedance ($20\text{ G}\Omega$) that was far greater than the typical impedance of a few $\text{k}\Omega$ observed between any electrode in the monitoring array and its reference electrode. This high input impedance is necessary in order to avoid current leakage in the voltmeter. The remote electrode was located at a large enough distance from the well where influence was expected to be negligible from pumping or hydraulic testing. Two additional electrodes were deployed at perpendicular directions from the reference electrode in order to provide two remote dipoles for monitoring telluric fields and any other ambient electrical noise.

One of the challenges to interpreting SP data is the difficulty in separating electrical noise that is recorded with the self-potential data. Many studies have recognized that SP data can be challenging to interpret due to multiple sources of signal and noise (e.g. Corwin, 1990; Perrier et al., 1997). Sources of noise include powerline harmonics, telluric currents, long-term thermal changes in soil temperature, and electrode drift (e.g. Revil et al. 2012). The magnitude of noise may be comparable to, or even much greater than SP arising from electrokinetic coupling. In most monitoring studies, SP data are post-processed to deal with high-frequency noise using a variety of lowpass filters (e.g. Rizzo et al. 2004), and wavelet based filters (e.g. Moreau et al. 1996). However, noise occurring at frequencies exceeding half the sample rate can be aliased to appear at lower frequencies in the same bandwidth as expected electrokinetic effects. This highlights the importance of removing noise sources in the field before the raw data are digitized.

The SP measurements reported in this thesis were acquired by operating the CR1000 data logger in a mode that integrated voltage over 16.66 ms (one cycle of 60 Hz) to cancel stable powerline noise occurring at 60 Hz and its harmonics. To further reduce noise, the leads from each SP dipole were terminated, at the inputs to the multiplexer, with a passive custom designed resistor-capacitor (RC) filter to attenuate noise at frequencies higher than ~ 1 Hz, including powerline harmonic and telluric fields. Figure 1.5 shows both the Campbell Scientific AM32B multiplexer and the CR1000 data logger interfaced with the RC filter box. Details of the RC filter design and its performance are presented in Chapter 4.

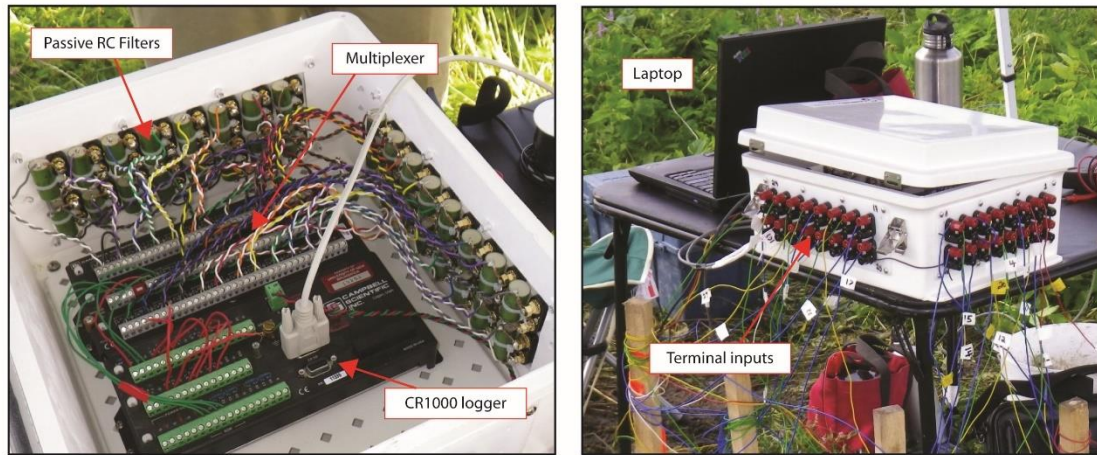


Figure 1.5. Custom made passive RC filter box and data acquisition system.

1.4 Thesis objectives

Although there is an abundance of research concerning SP monitoring of groundwater flow in the subsurface, relatively few studies have focused on flow through heterogeneous fractured bedrock aquifers. Therefore, the specific objectives throughout this work are as follows:

- (1) integrate borehole and outcrop fracture measurements with structural, hydrogeological and borehole geophysical data to develop conceptual aquifer models for two fractured bedrock aquifers in New Brunswick;
- (2) develop a low-noise multi-channel SP recording system to enable efficient acquisition of high quality data with broad spatial coverage;
- (3) using information from the conceptual aquifer models, monitor and interpret SP fields observed at surface in response to pumping or hydraulic testing with the objective of inferring water table drawdown and any fracture-related azimuthal anisotropy in fluid flow;
- (4) perform rigorous numerical modelling to determine the types of aquifer heterogeneity and anisotropy most amenable to detection by surface SP monitoring; and
- (5) demonstrate a simplified approach to interpret transient SP responses recorded over a network of electrodes to approximate the shape of the water table and estimate equivalent aquifer hydraulic properties.

1.5 Thesis outline and contribution

This introductory Chapter 1 describes the motivation for the study, presents a summary of the overall research approach, and describes the overall objectives and contributions of the thesis. Chapters 2 – 5 are presented in the form of four stand-alone journal articles, followed by conclusions in Chapter 6. The articles in Chapters 2 and 3 have been published in *Hydrogeology Journal*, while the article in Chapter 4 has been published in *Geophysical Journal International*, both of which are international peer-reviewed journals. Chapter 5 has been written as a journal-article and has been submitted to *Geophysical Journal International* for review. These articles represent significant contributions towards understanding how fracture networks influence the predominant

flow paths in two specific fractured aquifers and how non-invasive SP measurements can be applied to hydrogeological investigations in fractured rock.

The first article (DesRoches et al. 2012), reprinted in Chapter 2 presents an analysis of pump test and geophysical well log data acquired in a fluvial sandstone aquifer within the resource-rich Moncton Subbasin near the community of Springdale, New Brunswick, Canada. The fracture network was characterized statistically in terms of a set of sub-horizontal bedding plane partings and 3 steeply dipping fracture sets with different orientations, spacings, and apparent aperture distributions, and dominant flow directions during pumping were assessed for purposes of developing a water wellfield protection plan. Overall, the multidisciplinary approach furthered our understanding of heterogeneity and anisotropy conditions observed at the Springdale Wellfield. Heterogeneity in hydraulic conductivity was determined to be related to a lack of lateral continuity of the channelized sandstones units enveloped in shale, coupled with differences in fracture characteristics between the two lithologies. The integrated results contributed to the development of a conceptual hydrogeological model. In addition, observed fracture network characteristics and pump test drawdown data are interpreted in terms of the regional geological stratigraphy and structure, as inferred from geological reports, and maps. This provides a basis for extrapolation of the conceptual hydrogeological model at Springdale to other similar areas along the southern margin of the Moncton Subbasin.

The second article (DesRoches et al. 2014), reprinted in Chapter 3, provides a comprehensive fracture network analysis in a folded calcareous to argillaceous shale aquifer underlying the Black Brook Watershed (BBW) in northeastern New Brunswick, Canada, which forms part of an important agricultural region along the St. Johns River valley. This

study was motivated by elevated nitrate levels identified in private and municipal wells within the BBW, which have been related to intense application of agricultural fertilizers (Chow and Rees 2006). The study aimed to determine how fracture network characteristics and hydraulic conductivity distributions influence groundwater flow, and to provide a first-order approximation for anisotropy used to improve an equivalent porous media model results. The fracture network was evaluated in boreholes and large outcrop exposures using geophysical and geological methods, while hydraulic conductivity distributions were determined through borehole hydraulic (packer) testing. A subwatershed-scale conceptual and numerical groundwater flow model was also developed to further validate the hydraulic properties derived from field investigations by parameter estimation and calibration based on stream flow and hydraulic head observations. This article also describes relationships between the fold geometry, the contemporary stress field, and observed fracture patterns, and draws attention to their influence on groundwater flow in the study area. Wide-aperture and oxide-stained bedding plane fractures within large-scale open folds provide the preferential groundwater flow pathways, whereas a predictable distribution of high-angle fracture sets provide vertical connectivity. Given the consistency of regional-scale fold axis orientations and contemporary stress field throughout the BBW and surrounding area, the observed fracture patterns can be broadly extrapolated to other nearby areas provided that variations in bedding dip can be determined.

The third article (DesRoches and Butler 2016), reprinted in Chapter 4, confirmed the viability of using SP to infer water level drawdown and estimate transmissivity through a confined and heterogeneous fractured aquifer during pumping. Information gathered and presented in Chapter 2 was used to develop the initial conceptual hydrogeological model,

which was employed in the interpretation of the SP data. Clear transient SP anomalies with magnitudes between 2 to 9 mV were monitored at 31 different electrode locations using a simple 0.8 Hz lowpass filter on each channel to improve on the noise rejection characteristics of a commercial data logger. The SP anomalies generated by pumping were used in combination with a measured electrokinetic voltage coupling coefficient to estimate spatial variations in drawdown around the well. Apparent transmissivity and storativity values were then obtained by fitting the SP-derived drawdown data to the Theis model. Transmissivity results compare favorably with values derived by monitoring-well water levels during long-period pump tests. The conditions under which the simplified semi-empirical interpretation of drawdown can be expected to yield reliable results were investigated by rigorously modelling the relationship between SP and hydraulic head for a confined aquifer inspired by the Springdale site.

The fourth article (DesRoches and Butler 2017, in review), in review with the *Geophysical Journal International*, and reprinted in Chapter 5, presents both monitored and modeled SP signals recorded on the surface during a constant head injection test where fluid was injected into a 2.2 m interval isolated between two inflatable packers centered approximately 44 m below surface in a vertical well. The fractured bedrock aquifer for this study is located in the Black Brook Watershed (BBW). A strong negative SP response was observed at the start of the injection period with strongest magnitudes recorded by electrodes nearest the well and the SP response decreasing at more distant electrodes. Deflation of the packer system resulted in a strong reversal of the SP signal. The SP anomaly pattern observed on surface at steady state showed a strong correlation with the strike of two moderate to steeply dipping fracture sets known to intersect the test interval.

Numerical modelling of water flow and resulting SP fields expected within simplified fracture network models showed that azimuthal patterns in SP are expected to be mainly controlled by transmissive fracture orientations. The strongest SP sources are associated with water flowing out of the transmissive fractures into the tighter matrix and other less permeable cross-cutting fractures. Additional sensitivity modelling showed that increasing fracture frequency near the well, while holding the injection rate constant, increases the SP magnitude and enhances the SP anomaly parallel to the transmissive set. Decreasing the length of the transmissive fractures leads to higher fluid fluxes into the matrix and into cross-cutting fractures close to the well, resulting in a broader and higher magnitude SP anomaly. Results from this field and modelling study, reinforced by recent 2D modelling of SP in fractured rock by Roubinet et al. (2016), provide evidence that surface-based SP monitoring during constant head injection tests has the ability to identify groundwater flow pathways within a fractured bedrock aquifer.

1.6 Statement on co-authorship

In the following research I was responsible for designing field experiments, acquiring data, post-processing and interpreting the geological, hydrogeological and geophysical data with few exceptions. The hydraulic packer test data reported in Chapter 3 was acquired jointly by Environment Canada (EC) and Agriculture and Agri-Food Canada (AAFC), and the groundwater flow model incorporated into Chapter 3 was developed by co-author Dr. Serban Danielescu (AAFC). The Pb-PbCl₂ electrodes (based on a design by Petiau 2000) and the custom-made passive resistor-capacitor (RC) filter box (designed jointly with my thesis supervisor (Butler et al. 2011)) for removing high-frequency noise from the electrical field measurements were specifically constructed for this thesis. For all articles, I drafted

the manuscripts and figures, implemented editorial changes and submitted the papers for publication. Scientific and editorial guidance were offered by my co-authors and journal peer-reviewers.

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Chapter 2

Influence of fracture anisotropy and lithological heterogeneity on wellfield response in a fluvial sandstone aquifer of the Carboniferous Moncton Subbasin, Canada¹

2.1 Abstract

Pump tests and geophysical logs acquired in a fluvial sandstone aquifer within the resource-rich Moncton Subbasin of New Brunswick, Canada, have been used to characterize fracture patterns and flow directions for purposes of developing a water wellfield protection plan. Fracture patterns consist of three high-angle sets, and a low-angle set parallel to bedding. NW trending high-angle fractures, predominantly in fluvial sandstone units, appear to be most important in controlling groundwater flow directions. This contrasts with an earlier regional hydrogeological study that attributed most flow to sub-horizontal bedding plane fractures.

Water levels monitored during a 72-hour pump test revealed drawdown extension parallel to the NW trending fracture set. Drawdown curves indicate that the aquifer is laterally constrained – likely reflecting differences in fracturing observed between the channelized sandstone and surrounding shale units. As a result, groundwater flow induced

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by pumping is influenced by both fracture anisotropy and by the heterogeneity of the fluvial depositional environment. Relationships observed between fracture patterns, regional geological structure and lithology provide a basis for extrapolating the conceptual model to other nearby areas in the region, where potential impacts of geological resource development on groundwater are attracting public concern.

2.2 Introduction

Groundwater extracted from the Carboniferous Maritimes Basin provides an important water supply for a large rural population in the provinces of New Brunswick, Prince Edward Island, and Nova Scotia in eastern Canada. In New Brunswick the basin underlies roughly one third (24,000 km²) of the land mass and the aquifers consist predominantly of fractured, fluvial sedimentary rock, which is difficult to characterize due to geological heterogeneity and a lack of knowledge about the distribution of fracture patterns and their influence on groundwater flow rates and directions (Rivard et al. 2008a).

Motivation for the study originated from significant groundwater supply issues in the town of Penobscis (pop. 1,479 in 2006 census), New Brunswick. Beginning in late 2004, local residents experienced both lowering of water levels and significant reductions in the yields of their private domestic wells (Jacques Whitford 2005). Water reductions have been postulated to be related to potash mining in the area, although this has not been proven. The puzzling water supply problem, coupled with public unease about both potash mining and oil and gas development in the region, led the Province of New Brunswick to request a detailed borehole geophysical and hydraulic investigation to improve understanding of the aquifer system and develop a protection plan for a new municipal wellfield. The new wellfield, situated in the community of Springdale, approximately 1 km southeast of

Penobsquis, is currently supplying water to local residents and may be used as a regional water source in the future.

The study area is located along the southern margin of the Maritimes Basin, within the Moncton Subbasin (3700 km²), adjacent to the Caledonia Uplift terrane. The Maritimes Basin (148,000 km²) is the largest sedimentary basin in the Appalachian belt, underlying central and eastern New Brunswick, all of Prince Edward Island, and portions of Nova Scotia, and Newfoundland (Figure 2.1). A number of major ENE trending faults and fold axes cross the study area leading to complex geological and hydrogeological conditions that have complicated local wellfield development (Gemtec 2010) as well as water management in a nearby underground potash mine (Jacques Whitford 2005). Few studies have focused on characterizing fracture orientations and lithological heterogeneities within the Maritimes Basin at the wellfield scale. Carr (1964) provided a regional geological and hydrogeological investigation focused on pump test results from different bedrock formations within aquifers of the Moncton Subbasin. Most recently, investigations by Rivard et al. (2008a; 2008b) carried out over a large portion of the Maritimes Basin suggested that the fractured sedimentary rocks are highly stratified and lenticular in nature. The study further suggested that the fractured bedrock contains an abundance of water which predominantly circulates through the more transmissive bedding parallel fractures. The latter observation differs from the results in our study area near the boundary of the Caledonia Uplift terrane in the Moncton Subbasin (Figure 2.2). Rivard et al. (2008a) did however acknowledge that fracture orientations at different locations within the basin could express a completely different pattern associated with various degrees of tectonic uplift and faulting along the basin and subbasin margins.

The present study uses a multidisciplinary approach to further our understanding of heterogeneity in the bedrock aquifer responsible for anisotropic groundwater flow conditions observed at the Springdale Wellfield, and to analyze hydraulic properties of the aquifer through borehole pump tests and geophysical logging methods. Heterogeneity in permeability is found to be related to differences in fracture characteristics of the sandstone and shale units coupled with a lack of lateral continuity in the fluvial depositional environment that resulted in an aquifer/aquitard sequence. Integrated results contribute to the development of a conceptual hydrogeological model which has been used as the basis for design of a wellfield protection plan. In addition, observed fracture network characteristics and pump test drawdown data are interpreted in terms of the regional geological stratigraphy and structure, as inferred from geological reports, and maps. This provides a basis for extrapolation of the conceptual hydrogeological model at Springdale to other similar areas along the southern margin of the Moncton Subbasin.

2.3 Geology

The Springdale Wellfield is located 18 km east of the town of Sussex, New Brunswick along the South Brook valley, and is situated within the fault-bounded Carboniferous Moncton Subbasin (Figure 2.2). The Moncton Subbasin represents a substructure of the larger Carboniferous Maritimes Basin, composed of clastic nonmarine fluvial and lacustrine deposits, comprising mainly sandstone and shale, and gypsum, potash and rock salt evaporites, which were folded, faulted and thrust during the Carboniferous period (St. Peter and Johnson 2009). The subbasin is dominated by ENE trending reverse fault systems (Berry Mills, Penobsquis, Springdale, Gordon Falls, and Clover Hill faults) and, to a lesser extent, by more minor north trending extensional faults (Wilson 2005; Hinds 2008,

St. Peter and Johnson 2009). It is bound to the south by the Caledonia Uplift and to the north by Kingston and Mascarene Uplifts (St. Peter 1993, see Figure 2.2). The complex fault system is a consequence of transpressional tectonics generating thrust faults with a combination of dextral strike-slip and dip-slip movement (St. Peter 1993; Wilson and White 2006). Wilson (2005) and Wilson and White (2006) recognized several faults which experienced periods of reactivation throughout the basin development, including late contractional reverse displacements along the Berry Mills, Gordon Falls, Urney and Clover Hill faults. Later development of the Penobsquis and Springdale Faults is interpreted to be linked to the development of the Windsor Group salt structure (Wilson and White 2006). These faults are located approximately 1 to 2 km north of the wellfield and dip towards the north as thrust faults (Figure 2.3) (Hinds, 2008). The north trending Urney and Cedar Camp faults are located approximately 1.5 km to the east and west of the site and terminate roughly along the Springdale fault. The Urney fault is a low angle thrust fault that appears to be associated with the Gordon Falls fault at depth (Wilson 2005).

Groundwater in the study area is derived from the fractured sedimentary rock in the Carboniferous Boss Point Formation of the Cumberland Group (311 to 318 Ma). Depositional units include massive to trough cross-bedded sandstones, interbedded with fine-grained siltstones and shale deposited within large braided channel systems with associated lacustrine and fluvial environments (Browne and Plint 1994). Carr (1964) considered these deposits to be an excellent quality aquifer within the Moncton Subbasin, with high-angle fracture sets controlling the conductivity of the aquifer. Carr (1964) also noted springs associated with the Boss Point Formation were most commonly found along the sub-horizontal boundary between its sandstone units and the impermeable fine-grained

facies of the underlying Mabou Group sediments. In the area around the wellfield, fractured bedrock outcrops are typically associated with aquifer discharge zones resulting in natural springs that are supplying the nearby river.

Quaternary sediments blanketing the area typically consist of 10 to 20 meters of alluvial sand, and a thin till veneer (Rampton et al. 1984). In some locations drilling logs report gravel, silt and minor clay overlying a thin glacial till layer of variable thickness (2.7 to 6.1 m based on geological logs) and relatively low hydraulic conductivity at the base of the deposit. At high elevations, overburden sediments are observed to be relatively thin (~ 4 meters thick), and appear to thicken into the valley towards the South Brook (~ 10-20 meters thick). Areas of thin overburden are expected to yield recharge to the aquifer, particularly where the glacial till is absent.

2.4 Geophysical logging

Geophysical logging methods were employed in five vertical wells to characterize the fracture network in the Springdale Wellfield (Figure 2.4). Fracture characteristics were determined using acoustic and optical borehole televiewer imaging methods (Figure 2.5) to provide a continuous and oriented view of the borehole wall from which fracture orientation, aperture, and depth were determined. The orientations were calculated taking variations in borehole inclination into account, although inclinations measured by the televewers deviated only marginally (<2 degrees) from vertical. Televiewer resolution was sufficient to identify fractures with apertures as small as 0.1 mm, although aperture estimates below 1 mm were unreliable. Fracture frequencies, in 1 m intervals along the length of borehole, were determined by inspection of the televiewer images and plotted as

shown in Figure 2.6. These plots typically show a decrease in fracture frequency with depth in the five surveyed wells.

Other continuous geophysical logs recorded in the five boreholes to assist in lithological and hydraulic interpretations included three-arm mechanical caliper, natural gamma, single point resistance, and self-potential (Figure 2.6). The caliper log was used together with acoustic televiewer travel-times to identify borehole diameters as well as intervals characterized by a high density of fracturing. The natural gamma and single point resistance tools were particularly useful in distinguishing between sandstone and shale dominant units.

A heat pulse flowmeter was used in three boreholes to identify vertical flow along the borehole axis produced by transmissive fractures or fractured zones. Flow measurements were made at 2 m intervals along the entire saturated length of each hole under both ambient conditions and with a pump operating in the borehole just below the water table. This approach was intended to be used to calculate the transmissivity of each 2 m interval according to the method presented by Paillet et al. (1987). However, transmissivity calculations proved untenable due to the overwhelming influence of an intermittent high yield pump operating in the municipal production well (PW07-1) 80 to 220 m away from the three tested boreholes. Instead, the flow data was used qualitatively to identify hydraulically conductive fracture intervals, and determine vertical flow directions throughout the boreholes where negative flow is downward and positive flow is upward in the well.

2.5 Fracture network characterization

Two fracture types were identified consisting of 179 high-angle fractures and 84 bedding plane fractures. A histogram plot of observed fracture dips (Figure 2.7a) shows a bimodal distribution, skewed to higher dip angles. Application of a Terzaghi correction factor $1/\cos\theta$ to each bin (where θ is the dip angle at the center of each histogram bin) to account for the sampling bias associated with using vertical boreholes (Terzaghi 1965), yields the corrected histogram displayed in Figure 2.7b, which shows a fracture distribution that is strongly dominated by high angle dips. Fracture orientations were plotted on an equal-area stereographic projection and the pole densities were weighted to account for the low probability of intersecting high-angle fractures with vertical boreholes (Goodman 1976; Robin and Jowett 1986). Figure 2.8 presents the weighted fracture density contours of the high-angle and bedding plane fractures. While the high-angle and bedding plane fracture poles are broadly distributed about the stereonet they are dominated by fracture planes dipping towards the west and northwest. Analyses of the weighted fracture orientations provide evidence for four discrete fracture sets, consisting of three high-angle sets and a bedding plane parallel set.

2.5.1 High-angle fractures

Although the use of vertical boreholes favors intersections with shallowly dipping bedding fractures, high-angle fractures were observed to be more abundant, comprising 68% of the documented fracture population. Statistical evaluation of the high-angle fracture orientations revealed three discrete fracture sets (F1, F2, and F3). The mean F1 fracture plane strikes ENE and dips steeply to the WNW which is roughly orthogonal to the mean F2 fracture plane that strikes NW and dips steeply to the SW. Although the

formation of orthogonal fracture sets are not fully understood, Price and Cosgrove (1990) suggested that such fracture systems may result from regional uplift and stress relief (e.g. salt doming or post-glacial unloading).

Figure 2.9 displays the strike orientations of the three weighted high – angle fracture sets as rose diagrams. The fracture orientations are in good agreement with the trend of the regional- and local-scale fault systems and geological structure in the area mapped by Hinds (2008) (faults shown on geological map in Figure 2.3). The ENE trending F1 fractures are extensively developed and are parallel to the ENE trending major fault system that transects most of the Moncton Subbasin, including the more local trend of the Penobsquis and Springdale Faults located 1 to 2 km north of the study area, and the Clover Hill Fault located 6 km south. Although the Penobsquis and Springdale faults are the dominant tectonic features at the study site, only 32 % of the fractures are aligned with their trend. Regional fault trends formed as a result of transpressional strain following deposition of the Mabou and Cumberland Group sediment in the area, where the dominant fault type is strike-slip with a compressional component (Wilson and White 2006). These faults are interpreted to be genetically linked to the development of F1 fractures identified in the study area. A small number of F1 fractures show signs of movement which is evidenced by bedding planes and bedding plane fractures with minor amounts of offset in televiewer images (Figure 2.5b). The orientation of these fractures parallels the trend of the larger strike-slip fault structures documented by Wilson (2005) and St. Peter and Johnson (2009). Rivard et al. (2008a) working elsewhere in the Moncton Basin also identified a set of fractures (similar to the F1 set) aligned parallel with the major ENE trending fault system. The high-angle dip direction of the F1 fracture set towards the northwest is the same as the

dip direction of the Penobsquis and Springdale Faults located north of the site (Wilson and White 2006). The ENE trending fracture set also roughly parallels the fold axis of the Springdale and South Branch syncline and Cedar Camp anticline located between 500 and 3000 m southeast of the study area.

The F2 fracture set has a mean strike to the NW which is perpendicular to the nearby Springdale and Penobsquis faults and at a 30° angle to the smaller Urney and Cedar Camp Faults that trend N-S and dip towards the west (Hinds 2008). The NW oriented fracture patterns are thought to be a result of either the strike-slip component of the transpressional stresses, or a stress relief mechanism resulting from the compressional forces that generated thrusting and salt doming 2 km to the north. Fractures oriented along this trend are inferred to have a higher hydraulic conductivity based on anisotropy observed in the water table drawdown during pump tests, and on analyses of fracture aperture and spacing (discussed further in Sections 2.5.4 and 2.6). Increased hydraulic conductivity may result from fault and fracture opening and reactivation caused by strike-slip movements along the ENE trending fault systems (Wilson 2005) whereby fractures oriented NW may exhibit an extensional component. Similar arguments have been proposed by others who have observed that fractures oriented parallel to the direction of principal compressive stress form extension fractures that tend to be open, whereas those fractures perpendicular to this direction tend to be closed (e.g. National Research Council 1996, Singhal and Gupta 2010).

Fractures comprising the F3 set have a mean strike NNE and dip steeply to the west. This fracture set has a mean trend 50° different from the F2 fracture set; the two sets are interpreted to be a conjugate pair developed in response to the same stress field.

The three high-angle fracture sets were further assessed to determine the relationship between fracture orientations and the predominant shale and sandstone lithologies identified in borehole logs. Based on televiewer logs, a number of high-angle fractures identified in sandstone units were observed to terminate along sandstone/shale contacts or at bedding plane fractures. Rose diagrams of the high-angle fracture orientations for the fractured sandstone and shale lithologies are displayed in Figure 2.10. Fractures in the sandstone units include an abundance of F1 and F2 fracture orientations. In contrast, fractures in the shale units are dominantly of types F1 and F3, with very few F2 fractures present.

2.5.2 Bedding plane fractures

Bedding plane fractures (S1), readily apparent in the televiewer logs, are relatively shallowly dipping ($< 20^\circ$) and extremely variable in orientation as would be expected in a fluvial environment. Many fractures are stained with oxide minerals suggesting they may be susceptible to groundwater flow. These fractures are typically formed as a result of stress relief during tectonic uplift related to extensive weathering and erosion, salt doming and perhaps deglaciation, or possibly as flex-slip planes during folding (Price and Cosgrove 1990). Wilson and White (2006) identified multiple periods of subsidence, contraction, and uplift during the Carboniferous which could be a responsible mechanism. Similar stress relief mechanisms were postulated for other layered aquifers where bedding plane fractures play an important role in providing groundwater flow pathways in bedrock aquifers (e.g. Michalski and Britton 1997, Nastev et al. 2008, Rivard et al. 2008a; 2008b). A number of bedding plane partings with large apertures were observed in televiewer logs at sandstone/shale contacts, typically at the base of the sandstone sequence (e.g., Figure 2.5a).

Analysis of the fracture apertures and spacing, as detailed below, indicates that although they do not dominate the population, they still contribute significantly to groundwater flow.

Significant, heavily fractured zones consisting of multiple fracture intersections over 15 to 50 cm intervals were also identified. These zones were commonly associated with drilling-induced enlargements (wash-outs) revealed by the caliper log. Ten such highly fractured zones, all exhibiting weathering and oxidation staining along the fracture surfaces, which are indicative of groundwater movement, were identified in the five boreholes.

2.5.3 Fracture spacing distribution

Fracture spacing contributes significantly to hydraulic conductivity and groundwater flow in fractured rock aquifers (Singhal and Gupta, 2010). Fracture spacings were measured as distances between fractures within the same set observed along the borehole walls. This analysis method requires a correction to be applied to calculate the fracture spacing perpendicular to the fracture plane. Perpendicular fracture spacing S_n was calculated from (Priest, 1993):

$$S_n = S \cos \theta \quad (2.1)$$

where S is the spacing observed along the borehole wall, and θ is the angle between the borehole and mean dip of the fracture set.

Goodness-of-fit tests were evaluated using the exponential, log-normal, and Weibull distribution curves for each fracture set (significance level of 0.1) in order to calculate the optimal distribution curve to fit the data (D'Agostino and Stephens 1986). High-angle fracture sets are best fit by a Weibull distribution, and the bedding plane fractures are best fit by the log-normal distribution. Although, the fracture spacing data for the three high-

angle fracture sets are best characterized by the Weibull distribution, the goodness-of-fit test suggests that the exponential distribution also fit the F2, F3, and S1 fracture sets. Mean fracture spacings of the 3 high-angle fracture sets and bedding plane fractures, as modeled by the three different distribution functions, are summarized in Table 2.1.

Table 2.1 Fracture spacing parameters and goodness of fit statistics for the four fracture sets and three possible distribution functions. High-angle fracture sets F1, F2 and F3 are best represented by the Weibull distribution. A log-normal distribution provides the best fit to the bedding plane fractures (S1).

Distribution Function	Fracture Set	Fracture Spacing Parameter			Goodness-of-Fit Test	
		Mean S_n (m)	Shape	Scale	Anderson-Darling	P-value
Exponential	F1	0.37	-	-	1.38	0.04
	F2	0.28	-	-	0.68	0.29
	F3	0.64	-	-	0.33	0.76
	S1	1.23	-	-	0.77	0.22
Weibull	F1	0.37	0.80	0.32	0.32	>0.25
	F2	0.28	0.92	0.27	0.32	>0.25
	F3	0.64	1.12	0.67	0.32	>0.25
	S1	1.23	0.98	1.22	0.70	0.07
Log-Normal	F1	0.49	-1.83	1.49	0.73	0.05
	F2	0.34	-1.94	1.31	0.45	0.27
	F3	0.74	-0.91	1.10	0.57	0.12
	S1	1.33	-0.35	1.13	0.25	0.74

2.5.4 Fracture aperture and conductivity

Measured apertures (corrected for fracture dip relative to the vertical borehole wall) ranged from < 1 mm to >100 mm. It is recognized that fracture apertures recorded along borehole walls by televiewer logging may display somewhat larger openings compared to true fracture apertures due to enlargements from the drilling process. Forty-eight fractures were identified with aperture widths less than 1 mm, which represents the lower measurement limit of the televiewer referred to as a truncation bias. For the purpose of this hydrogeological study, these fractures are thought to be insignificant with respect to the

overall transmissivity of the well. Likewise, a few highly fractured zones and eroded fracture intervals were identified with aperture widths greater than 30 mm resulting from multiple fractures intersecting the same borehole interval as well as effects of drilling and widening of the borehole wall. Fractures with apertures larger than 30 mm were difficult to quantify with confidence, and because they were uncommonly observed in the data, they were not included in the distribution analysis. The resulting fracture apertures exhibit a log-normal distribution, as shown in Figure 2.11. Mean and median fracture apertures calculated for the three high-angle fracture sets range between 5.5 to 6.3 mm and 4.2 to 5.6 mm, respectively. Although the bedding plane fractures consist of fewer fracture planes, they tend to have larger apertures with a mean aperture value of 7.5 mm. As a result of overlying pressure, apertures of bedding fractures tend to decrease slightly with increasing depth.

Equivalent hydraulic conductivities of the individual fracture sets were estimated, assuming that they possess a constant aperture, using the cubic-law relationship (Snow 1968)

$$K = \frac{\rho_w g N b^3}{12\mu} \quad (2.2)$$

where ρ_w is water density, g is gravitational acceleration, μ is the dynamic viscosity of water, b is the mean fracture aperture for the fracture set, and N is the mean fracture density of the fracture set expressed in fractures per meter. Mean fracture densities for the four fracture sets in this case were determined as the reciprocal of the mean fracture spacing calculated by the distribution curves (Priest and Hudson 1981). Calculated hydraulic conductivities for the individual fracture sets range from 2.4×10^{-4} to 7.4×10^{-4} m/s, where the F2 fracture set

with a mean strike of $145^{\circ} - 325^{\circ}$ is characterized by the highest conductivity (Table 2.2). Although one might expect these estimates to be high, due to possible over-estimation of fracture apertures as discussed above, we note that they are the same order of magnitude as the hydraulic conductivities of 6.3×10^{-4} to 5.1×10^{-4} m/s determined from pump test calculations (described below), and are comparable to hydraulic conductivities determined by Rivard et al. (2008a) for the Boss Point Formation aquifer in other parts of the Moncton Subbasin.

Table 2.2 Equivalent hydraulic conductivity values calculated using mean apertures and fracture densities for the four fracture sets identified in borehole logging

Fracture Set	Strike Orientation (degrees)	Mean Aperture (mm)	Mean Fracture Spacing (m)	Fracture Density (Fractures/m)	Equivalent Hydraulic Conductivity (m/s)
F1	$070^{\circ} - 250^{\circ}$	5.46	0.37	2.74	3.7E-04
F2	$145^{\circ} - 325^{\circ}$	6.30	0.28	3.60	7.4E-04
F3	$017^{\circ} - 197^{\circ}$	5.74	0.64	1.55	2.4E-04
S1	Horizontal	7.50	1.33	0.81	2.6E-04
Average		6.25	0.63	2.18	4.0E-04

2.6 Hydrogeological characterization

Groundwater drawdown monitored during a 72-hour pump test showed that the Springdale wellfield aquifer is hydraulically connected along a transect of bedrock wells extending to the south of the well that was tested (08-1). The pumping well (8 inch diameter) was open from 18.9 to 58.8 m depth intersecting roughly 10 m of fractured sandstone, in a depth interval between 27.3 and 39.8 m (elevation 34.7 to 47.2 m above sea level) that is assumed to act as the primary aquifer. The well was pumped at 11.21 L/sec (148 igpm) and water levels were monitored continuously in the pumping well and 11 bedrock observation wells, located 12 m to 1183 m away. The pumping well recorded a maximum drawdown of 5.20 m, and seven of the eleven monitoring wells showed

discernible drawdown during the pumping period ranging from 1.99 m to 5.01 m below static water levels. Two monitoring wells completed in the overburden, at distances of 7.5 m and 210 m from the pumping well, showed zero drawdown, suggesting that the fractured bedrock behaves as a confined aquifer (wells not identified on map). This is consistent with geological logs which identify a clay rich glacial till layer near the base of the overburden deposit that could act as a confining layer. Well 07-2 was under artesian pressure with static water measurements located slightly above surface. Drawdown observations suggest that groundwater flow between the pumping well and monitoring wells to the NW is restricted based on observations of zero drawdown in three commercial wells (MW1, MW2, and MW3) located 948 m to 1117 m away, and a residential well located 834 m (MW4) away from the pump. Although driller's logs have identified sandstone intervals in the wells, the lack of drawdown may be the result of poor fracture connectivity to the NW or lateral heterogeneity associated with the intersection of different channelized sandstone intervals. On the other hand, a residential well (MW8) situated 1183 m SE of the pumping well was drawn down 1.9 m below the initial water level, indicating that the drawdown cone has a preferential extension in that direction. It is acknowledged that the spatial distribution of monitoring wells to the NE and SW of the pump was limited, making delineation of the drawdown cone in those directions difficult. Nonetheless, zero drawdown in well 09-2 revealed that there is reduced connectivity in the NE direction. Extensive drawdown along the NW-SE trend coupled with the lack of drawdown observed in 09-2 to the NE is suggestive of an anisotropic flow regime, presumably controlled by the NW-SE trending F2 fracture set. However, the observed results might also be a consequence of geological heterogeneity and poor hydraulic connectivity, given that the most transmissive interval in

the pumping well was likely a fractured sandstone layer that does not appear to extend to well 09-02.

Drawdown curves from the pump test were analyzed using the Cooper-Jacob straight line method (Cooper and Jacob 1946) for a confined aquifer in order to determine mean aquifer transmissivity and storativity. The effect of wellbore storage was noted to diminish after roughly five minutes of pumping. In most wells, the drawdown curves are characterized as two straight line segments on a semi-log plot (Figure 2.12). The initial straight line segments were interpreted to give transmissivity and storativity estimates by the Cooper-Jacob method, corresponding to the withdrawal of water from the fractured aquifer. The drawdown curves reach an inflection point after roughly 60 minutes where their slopes on the semi-log plot suddenly increase by approximately three-fold. Such an inflection is commonly attributed to the arrival of the drawdown cone at an impervious boundary (Driscoll 1986), which might in this case correspond to a zone of poorly fractured rock that is inadequately connected hydraulically to the dominant fracture network. In boreholes farther than 250m from the pumping well the inflection point was not observed suggesting that the drawdown cone may have already been affected by the impermeable boundary condition. Due to the effect of the impermeable boundary, only the early-time data was analyzed in the calculation of aquifer transmissivity and storativity, and the late-time data was disregarded.

The drawdown response in the aquifer is interpreted to be a function of the transmissivity and storativity of the aquifer, as well as the lateral heterogeneity of the highly transmissive fracture network. Similar drawdown responses exhibiting increases in the drawdown slope beyond an inflection point have been documented in other parts of the

Moncton Basin, and attributed to lateral variability in transmissivity, resulting from the lenticular nature of fluvial channel deposits observed throughout the Maritimes Basin (Rivard et al. 2008a). In those investigations, increases in the slope of the semi-log drawdown curve by more than a factor of two were attributed to the influence of more than one impermeable boundary.

Aquifer transmissivities and storativities calculated from the drawdown data are listed in Table 2.3. The drawdown data during the pump test indicate transmissivity values obtained for the aquifer ranged from 5.1×10^{-3} to $6.3 \times 10^{-3} \text{ m}^2/\text{s}$ (436 to 544 m^2/day) based on calculations from early time-drawdown behavior. Aquifer storativity data range from 2.9×10^{-3} to 3.5×10^{-5} , indicative of a confined aquifer typically in the range of 10^{-3} to 10^{-6} (Singhal and Gupta 2010). Aquifer hydraulic conductivities are estimated to be 2.5×10^{-4} to $3.2 \times 10^{-4} \text{ m/s}$, assuming a saturated aquifer thickness of 20 m, similar to the equivalent conductivities determined using apertures and fracture spacings. The thickness of 20 m corresponds to the average thickness of highly fractured sandstone unit observed in all the wells, which are believed to represent the main water bearing zones. The higher storativity value for well 08-02 is considered to be an outlier among the monitored wells with it being the only estimate greater than 5×10^{-5} . This may be a consequence of the monitoring well being located only 12 m away from the pumping well.

Table 2.3 Summary of aquifer properties determined from pumping test analyses for monitoring well drawdowns.

Well	Depth to Bedrock (m)	Total Depth (m)	Well Elevation (m)	Distance to Pumping Well (m)	Maximum Drawdown (m)	Transmissivity (m ² /s)	Storativity
07-1	10.7	42.7	71.7	245	4.69	5.2x10 ⁻³	3.6x10 ⁻⁵
07-2	23.5	36.6	68.5	229	4.64	5.1x10 ⁻³	4.9x10 ⁻⁵
PW07-1*	12.2	42.7	72.9	218	4.74	5.7x10 ⁻³	3.5x10 ⁻⁵
08-1**	12.2	58.8	74.6	-	5.20	6.3x10 ⁻³	-
08-2	18.3	60.9	74.3	11.5	5.01	6.1x10 ⁻³	2.9x10 ⁻³
09-2	4.6	91.4	115.5	270	0	-	-
MW1	15.5	99.1	64.1	953	0	-	-
MW2	24.7	123.5	64.0	1117	0	-	-
MW3	-	76.2	65.3	948	0	-	-
MW4	-	18.6	65.1	834	0	-	-
MW5	-	32	62.9	226	0	-	-
MW6	12.2	30.5	80.7	423	4.26	-	-
MW7	-	30.4	77.2	592	4.09	-	-
MW8	-	30.4	87.5	1183	1.99	-	-

* current wellfield production well

** well used for the pump test analysis (back-up production well)

2.7 Discussion

A conceptual understanding of the aquifer underlying Springdale Wellfield has been developed based on the synthesis of borehole geophysical logging and hydraulic data from this investigation, as well as a review of the geological setting, including regional geological structure, and of the limited hydrogeological data available from elsewhere in the Moncton Subbasin. The fracture network consists of three high-angle fracture sets and a set of sub-horizontal bedding plane fractures. Flowmeter measurements in selected boreholes indicate that the high-angle and bedding plane fractures in the sandstone units are significantly more conductive than the fractures in the surrounding shale units. Although interpreted only qualitatively, the flowmeter logs indicate that the aquifer can be subdivided into three hydraulic intervals, corresponding to a 15 to 25 m thick sandstone unit enclosed between an upper and lower bounding shale unit, which can be identified across multiple wells. In the sandstones, fracture frequency and aperture widths tend to be slightly higher, and oxidation staining of fracture surfaces tends to be more apparent compared to the shale units.

Oxidation staining on fracture planes provides evidence of significant groundwater flow (Nelson 2001). In particular, this evidence points towards the high-angle fractures as the primary pathway for groundwater flow. A combination of F1 and F2 fracture orientations dominate the sandstone units, resulting in a near orthogonal fracture network, which is well-connected providing an efficient aquifer system. In contrast, fractures within the shale units are dominated by the F1 and F3 orientations, with very few F2 fractures present. Although both lithologies contain a large number of ENE fractures (F1), these fractures are oriented parallel to the regional fault systems which exhibit substantial compressional and shear movement from transpressional stress. Price and Cosgrove (1990) noted that fractures formed by compressive stress commonly contain gouge that would limit their transmissivity. Furthermore, shear movement along fracture surfaces may lead to partial sealing of the fracture, smaller apertures, and reduced transmissivity (Singhal and Gupta 2010).

The observed difference in fracture set populations between lithologies, coupled with the nature of fluvial deposition, has led to an improved understanding of the heterogeneous influence on groundwater flow directions. Krasny and Sharp (2007) state that detailed fracture characterization is necessary in understanding groundwater flow systems, particularly when the aquifer is also complicated by heterogeneous properties. Michalski (1990) emphasized similar heterogeneities observed in lenticular deposits in the Brunswick Formation, Newark Basin, USA, where drastic differences in hydraulic conductivities are expressed between two varying sedimentary lithologies. Although groundwater flow is predominantly along discrete fracture planes, he noted that the distribution of fracture sets

throughout the aquifer is ultimately controlled by the spatial distribution of sandstones and shale in the fluvial depositional environment.

Flow logs and televiewer imaging also identified significant bedding parallel fractured zones located at the upper and lower boundaries between the sandstone and the shale unit, denoting the upper and lower bounds of the fluvial deposit. Lateral extent of the bedding fractures are likely controlled by the width of the channel. The unfractured matrix rock contributes little transmissivity to the system but is important for groundwater storage. Carr (1964) noted the bulk of groundwater flow occurred along fractures. Rivard et al. (2008a) provide a porosity range of 5 to 10 percent for the Boss Point Formation located in other areas of the Maritimes Basin.

Evidence from calculated hydraulic conductivities, based on fracture spacing and aperture data, suggest groundwater flow directions are largely dominated by high-angle fractures trending in a NW-SE orientation (F2). It is assumed that bedding plane fractures and other oblique fracture sets contribute to groundwater flow primarily by providing connectivity pathways between the N-S trending fractures. The dominant contribution of the F2 fracture set is inferred from hydraulic conductivities estimated using the cubic law relationship, which are two to three times greater for the F2 set than for the other fracture sets. Pump test results demonstrate an extensive drawdown cone predominantly to the NW and SE of the pumping well that is consistent with preferential groundwater flow paths along the F2 fracture set orientations. The abrupt change in drawdown behavior over time, coupled with the great extension of drawdown to the south of the pumping well, are suggestive of the presence of impermeable boundaries limiting the lateral extent of the aquifer. The highly fractured sandstone aquifer may therefore be modeled as a channel-like

body (consistent with its fluvial origin) having NW-SE orientation, laterally constrained and bounded by low transmissivity, moderately to poorly fractured shale. The artesian well (07-2), coupled with evidence from geological logs and drawdown data were used to recognize that the surficial aquifer within overburden is hydraulically disconnected from the underlying fractured bedrock aquifer by a glacial till layer. However, at high elevations where overburden sediments are relatively thin (~ 4 meters thick), these areas may represent potential recharge zones.

2.8 Conclusions

The fractured sandstone aquifer underlying the Springdale Wellfield demonstrates the influence of fracture anisotropy and lithological heterogeneity present at the local aquifer-scale in the Carboniferous Maritimes Basin. The aquifer is comprised of two dominant lithology end-members in the Boss Point Formation: (i) channelized sandstone deposits which are lenticular in nature, and (ii) adjacent laminated shale units. Analysis of fracture orientation data revealed three discrete high-angle fracture sets, and one sub-horizontal bedding plane fracture set. Although sampling relied on the use of vertical boreholes, high-angle fractures nonetheless dominate the observed fracture population, whereas subhorizontal bedding plane fractures only comprise 32% of the total population. Flow meter measurements in selected boreholes indicate that the fractures located in the sandstone units are significantly more conductive than the fractures in the shale units. Flow logs combined with televiewer images also identified significant flowing intervals at bedding parallel fractured zones located at the upper and lower boundaries between the sandstone and the shale unit. The F2 set of high-angle fractures trending NW are suggested to provide the dominant flow path within the sandstone unit based on flow meter

measurements, the elongated drawdown response and the hydraulic conductivity estimates for each of the fracture sets. Other fracture sets are shown to be present in similar abundances but are considered to be most significant in providing connectivity pathways between the fractures trending along the dominant flow direction.

Structural correlations of fracture orientations observed in wells in the study area are in good agreement with the orientation of local faults and the model of a transpressional regional stress regime. Fracture spacing data for fracture sets follow a Weibull distribution, with the exception of bedding plane fractures which best follow a log-normal distribution. The F2 fracture set exhibits the highest degree of fracturing with a mean fracture spacing of 0.28 m. Apparent fracture apertures are best described by a log-normal distribution and equivalent hydraulic conductivities estimated for each of the fracture sets are consistent with the hydraulic parameters determined from pump tests.

Pump test results revealed that the aquifer is confined. The presence of one or more impervious boundaries limiting the lateral extent of the aquifer was inferred from the abrupt increase in drawdown detected after approximately 60 minutes of pumping. This is consistent with the interpretation of lenticular channelized sandstone units, given that the degree of fracturing and connectivity between fracture sets appears to decrease with the transition into the shale units. This interpretation of a less transmissive boundary is further supported by the relative scarcity of F2 fractures, which parallel the dominant flow direction, in shale units.

A combination of site-specific borehole geophysical and pump test data, interpreted in the context of the regional geology of the Moncton Subbasin, has provided an effective way of determining the hydrogeological properties and conceptual understanding of the

hydrodynamics of the region required for the design of a wellfield protection plan in a structurally complex fractured bedrock aquifer. The inherent relationship observed between the distribution of fracture sets and nearby fault orientations, generated by transpressional stresses, provides an approach which may be used to predict fracture set distributions and groundwater flow directions in other structurally similar locations along the flanks of the Caledonia Uplift throughout the Moncton Subbasin. Such insight is timely in light of growing interest in the region's Carboniferous bedrock aquifers arising from pressures associated with existing potash and gas field development as well as shale gas exploration. The study demonstrates the applicability of fracture network assessment towards the development of a hydrogeological conceptual model.

2.9 Acknowledgments

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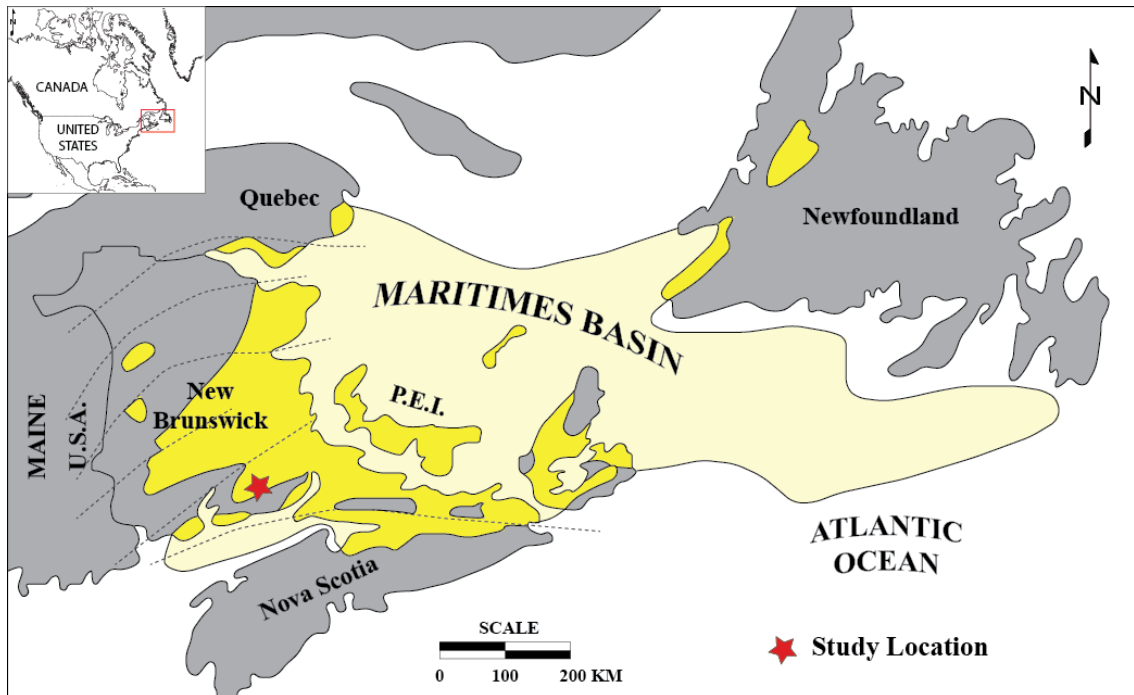


Figure 2.1 Regional distribution of the Maritime Basin through four Atlantic Canadian Provinces. The study area is located within the Moncton Subbasin in New Brunswick (Modified from St. Peter 1993)

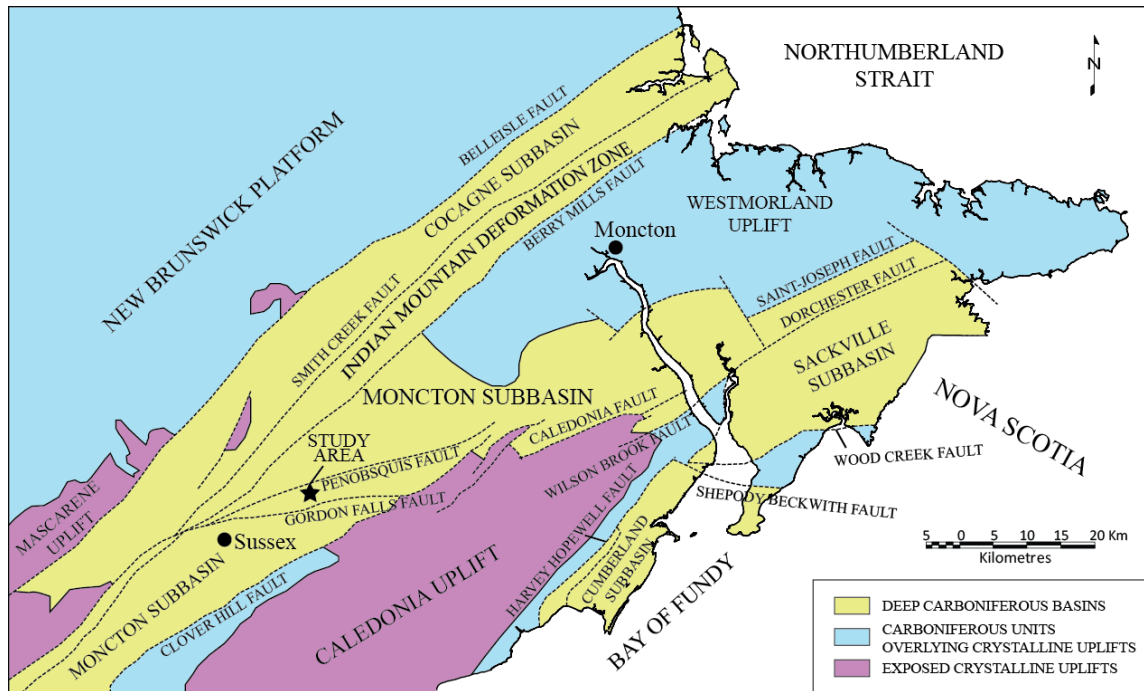


Figure 2.2 Geological map showing the major fault structures in the Moncton Subbasin along the southern margin of the Maritimes Basin in New Brunswick (modified from Wilson 2005)

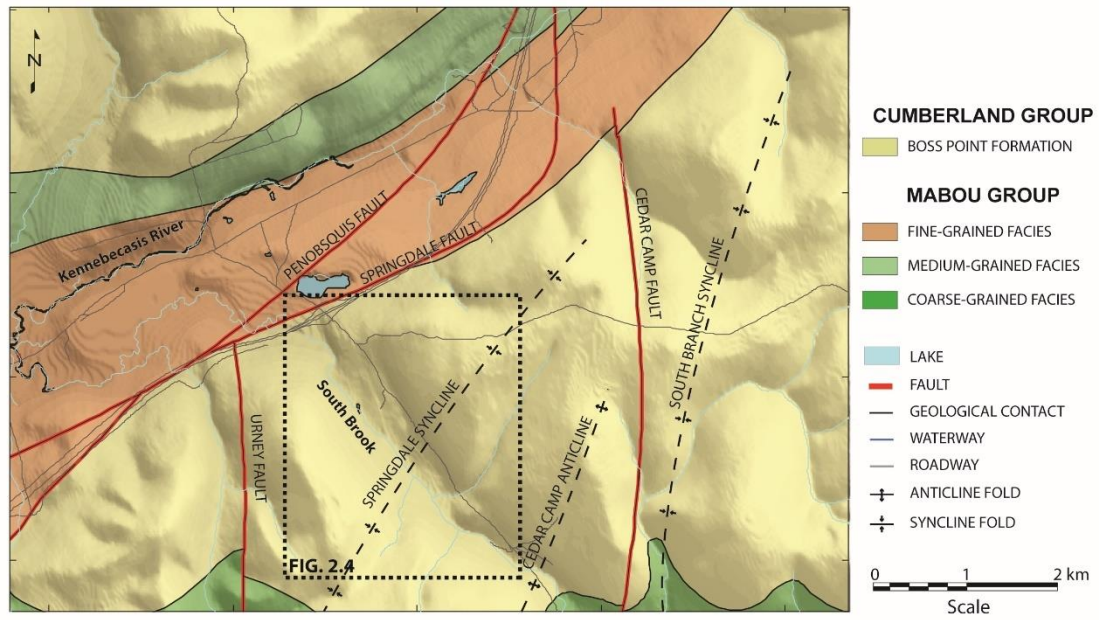


Figure 2.3 Geological map of the Springdale region with topography illustrated by shading (modified from Hinds 2008)

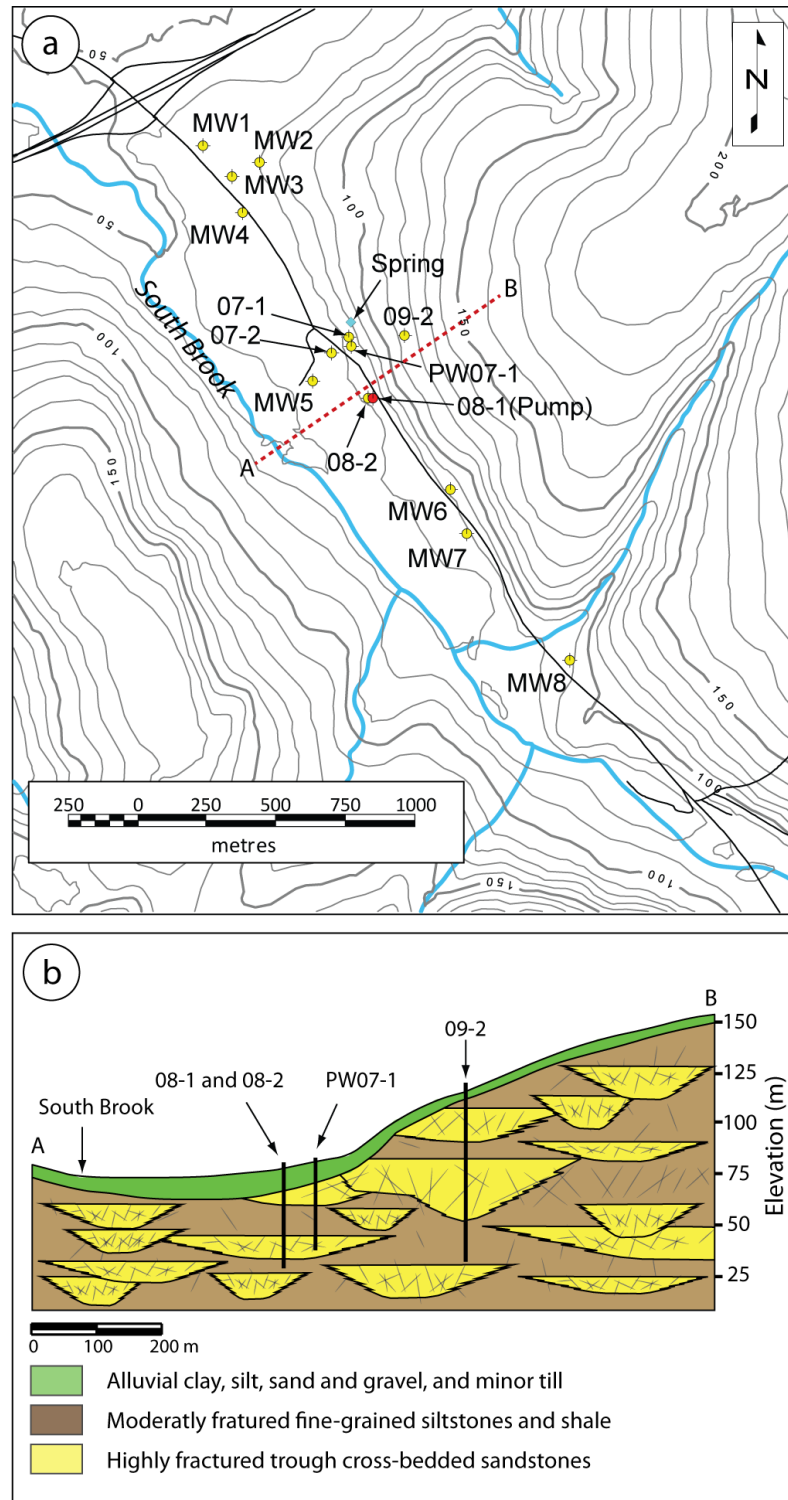


Figure 2.4 Location of the pumping and monitoring wells at the Springdale Wellfield site with 10 m topographic contours. The cross section (b) represents a schematic interpretation of lenticular sandstone channels hosted within shale units

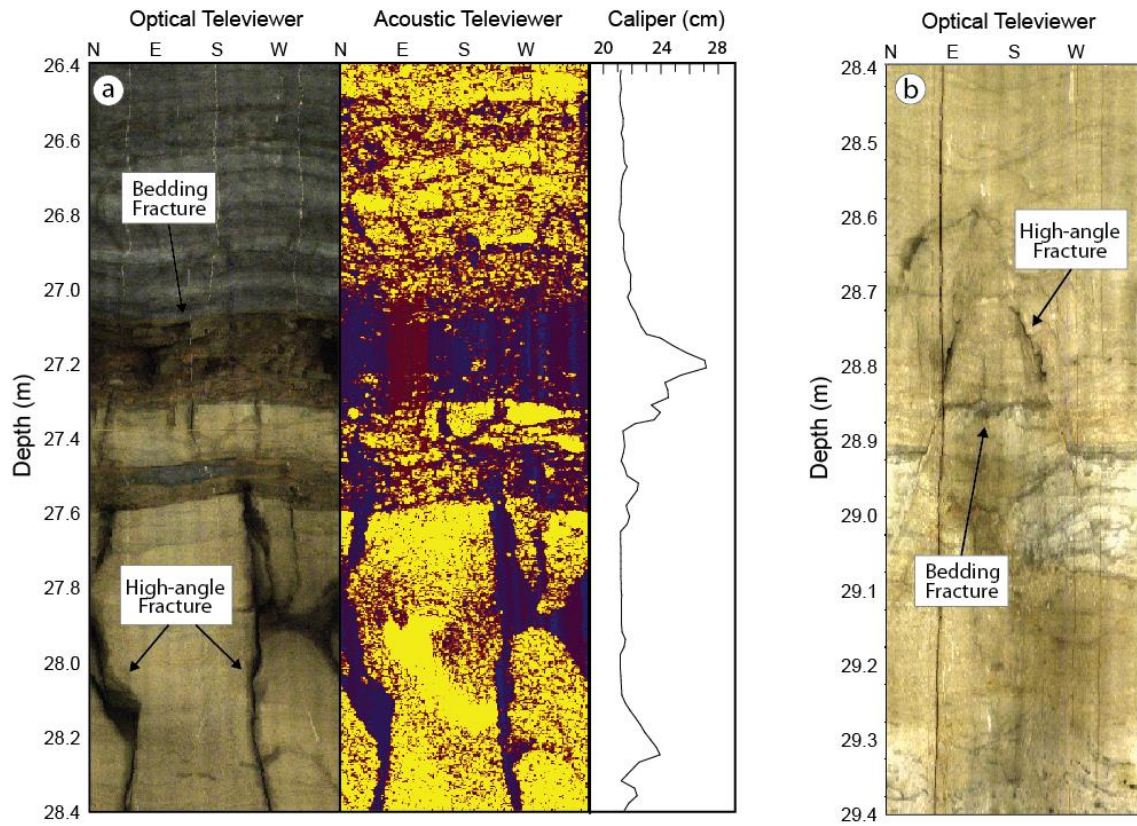


Figure 2.5 Examples of borehole televiewer data. a) Optical (left) and acoustic (right) televiewer images showing the horizontal fractured contact between the shale and sandstone, and a high-angle fracture in the sandstone in the well (08-01) used for the pump test. Caliper log of the borehole diameter shows the enlargement of the bedding parallel fracture. b) Optical televiewer image showing minor amounts of movement along a north-east trending high-angle fracture plane

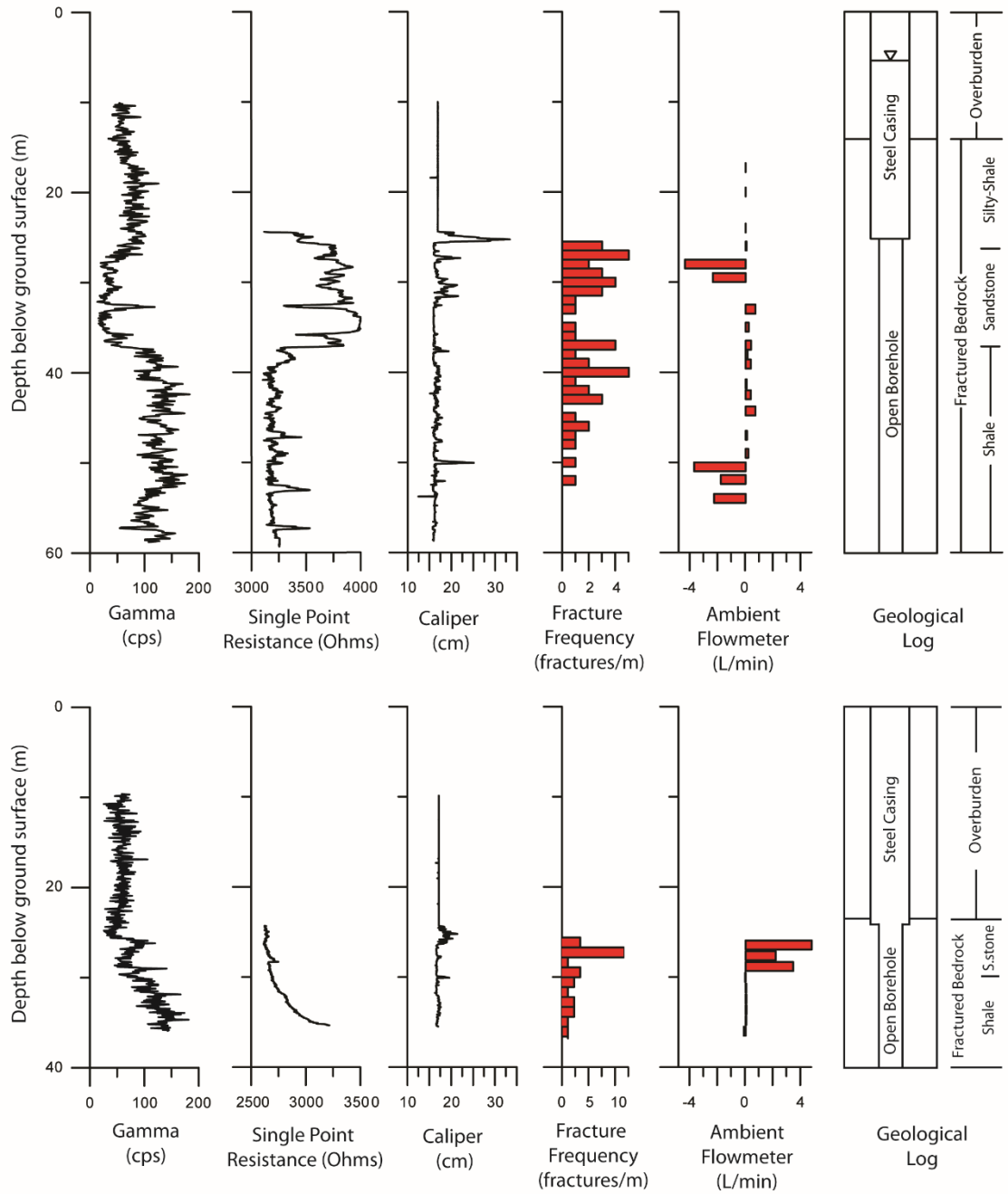


Figure 2.6 Geophysical logs from wells 08-2 and 07-2 located 11.5 m and 245 m (respectively) away from the aquifer pump test well, and 220 m and 75 m (respectively) away from the municipal well. Flowmeter data indicate that the fractured sandstone intervals and an additional fracture zone in the shale unit at depth (well 08-2) contribute to groundwater flow

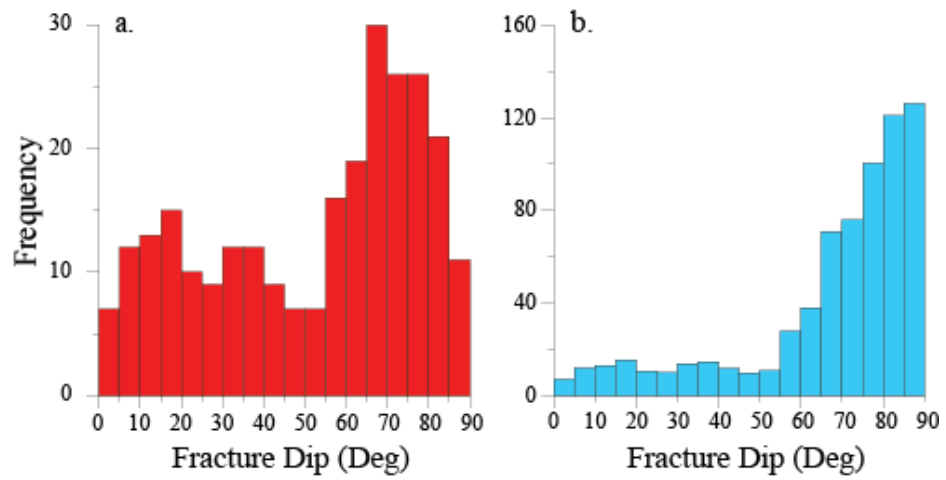


Figure 2.7 Histogram of the fracture dip angle versus frequency for the (a) raw fracture data and (b) Terzaghi corrected fracture data from five vertical boreholes

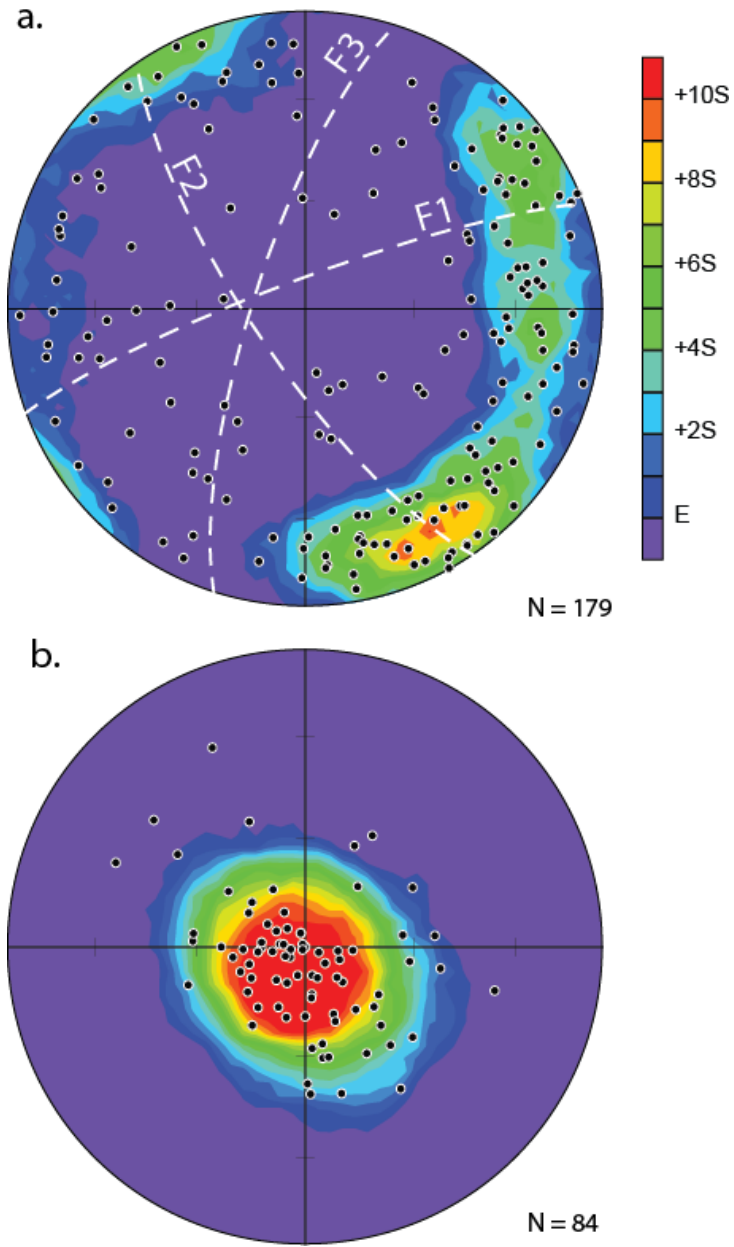


Figure 2.8 Equal-area projection of corrected poles to fracture planes measured in five boreholes. a) Three high-angle fracture sets represented by great circles and b) bedding plane fracture set with contoured pole densities by Kamb counting methods

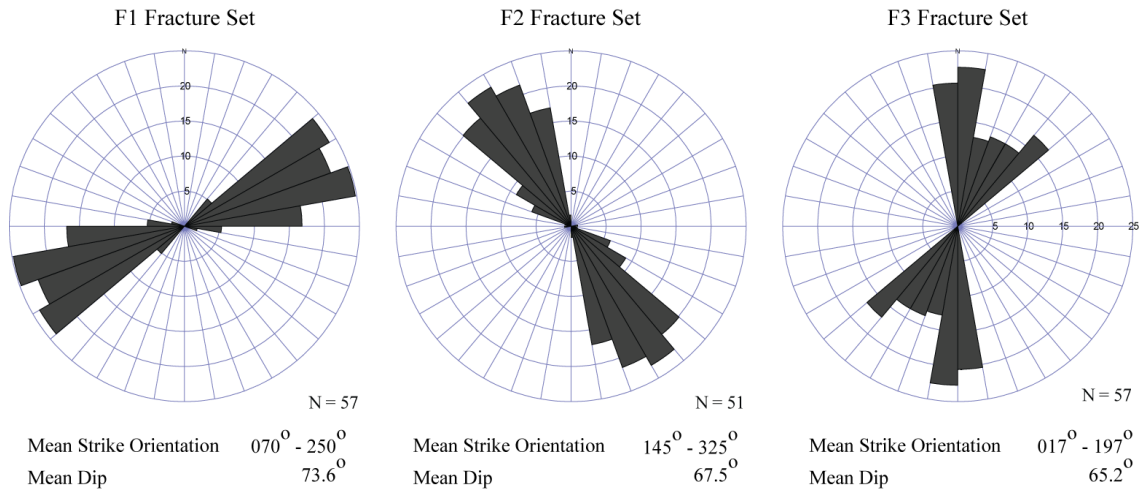
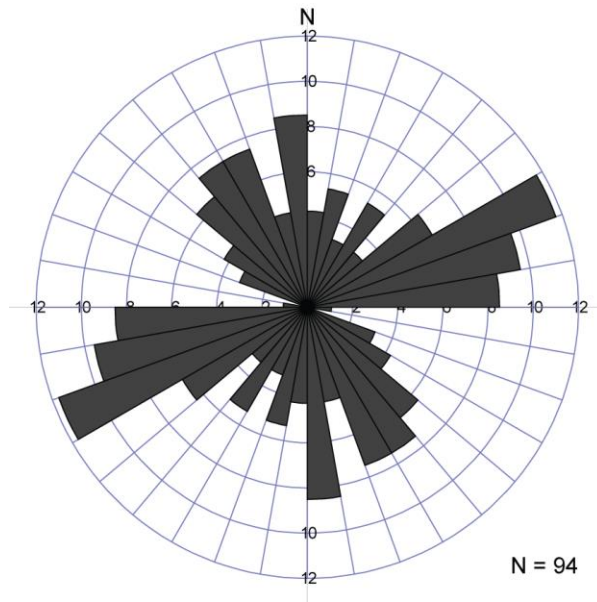


Figure 2.9 Individual rose diagrams represent the strike directions of three high-angle fracture sets (F1, F2, and F3) measured in five vertical boreholes (totaling 165 of the 179 fractures). The data are weighted to account for the sampling bias. Fourteen of the fractures contained shallow dips and therefore did not fit well into the three fracture sets.

Sandstone Fracture Orientations



Shale Fracture Orientations

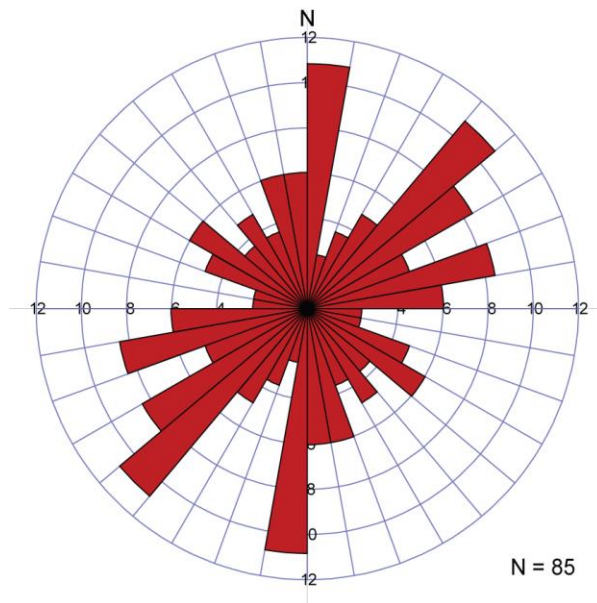


Figure 2.10 Variations observed between the high-angle fracture orientations in the sandstone and shale units. The sandstone units contain a higher abundance of F2 fractures, where the shale units comprise mainly F1 and F3 fractures, with a limited number of fractures from the F2 set.

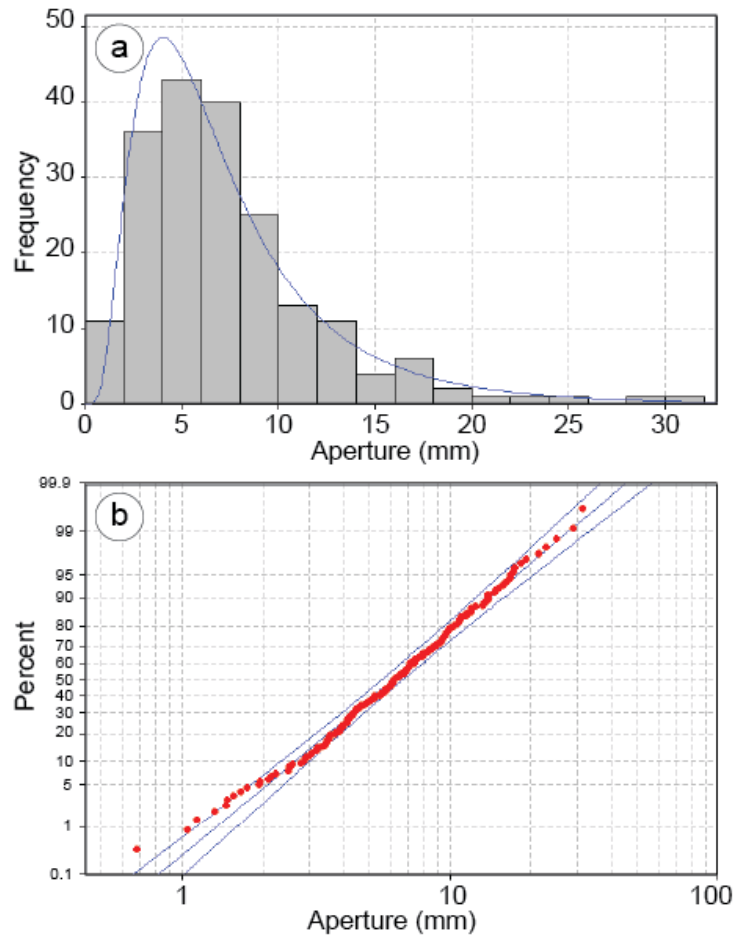


Figure 2.11 a) Apparent aperture histogram with a 2 mm interval fit with a log-normal distribution. b) The best fit model is confirmed by a quantile plot of the log-normal distribution model fit to the aperture data

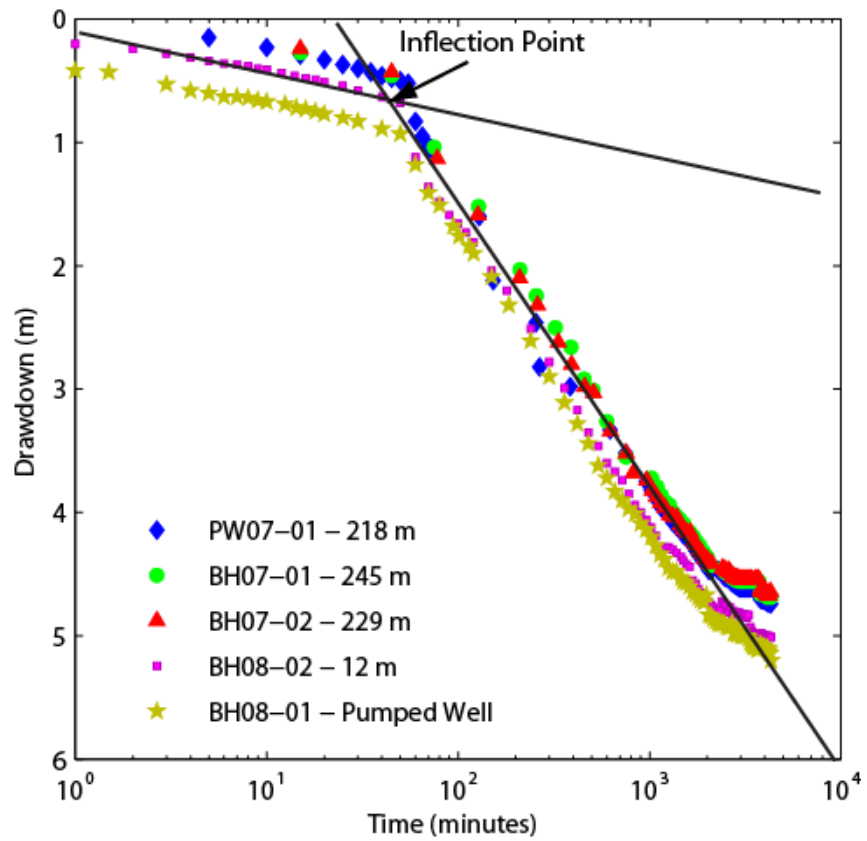


Figure 2.12 Semi-log plot of time-drawdown data from the pumping well and four monitoring wells ranging between 12 to 245 m from the pump. Transmissivity and storativity is approximated using the Cooper-Jacob straight-line method. The increased drawdown following the inflection point (~60 minutes of drawdown) represents the presence of an impermeable boundary condition

Chapter 3

Structural controls on groundwater flow in a fractured bedrock aquifer underlying an agricultural region of northwestern New Brunswick, Canada²

3.1 Abstract

A hydrogeological study was conducted in northwestern New Brunswick, Canada, to improve the predictability of fracture-dominated groundwater flow within folded bedrock composed of fine-grained turbidites. Borehole televiewer logging and outcrop mapping, integrated with hydraulic packer tests revealed enhanced hydraulic conductivity associated with northeasterly striking bedding plane fractures formed during folding and flexural slip. These fractures impart azimuthal anisotropy to the aquifer because of moderately dipping fold limbs. High-angle fractures form a well-developed non-stratabound network, comprising two open fracture sets striking NNE parallel to the current direction of principal stress, and WNW parallel to the direction of principal stress that dominated during the Acadian orogeny. The subset of fractures showing significant oxidation, deemed most important to the groundwater flow system, is dominated by bedding plane and high-angle fractures striking near-parallel to the maximum principal stress direction, resulting in

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extensional opening and enhanced hydraulic conductivities. An equivalent porous media model, incorporating anisotropy and varying hydraulic conductivity with depth, indicates that horizontal flow dominates the aquifer with relatively minor exchange between different model layers. These findings have implications for understanding flow directions in the Black Brook Watershed and elsewhere in the Matapédia Basin where fractures formed under similar stress conditions.

3.2 Introduction

It is recognized internationally that prolonged application of agricultural chemicals has an adverse effect on surface and groundwater qualities in agricultural regions (Power and Schepers 1989; Lefebvre et al. 2005; Razowska and Sadurski 2004). Nitrate levels monitored in groundwater resources across agricultural regions of Canada increased by about 25% between 1981 and 2001 (Lefebvre et al. 2005), with concentrations often in excess of drinking water guidelines (CCA 2009). The majority of these regions rely on drinking water sourced from private wells and municipal wells constructed in underlying aquifers (Corkal et al. 2004), many which are developed in fractured bedrock. Such groundwater resources require increased attention and application of watershed best management practices in order to minimize contamination. Better characterization of fracture networks and their hydraulic properties are required to develop groundwater flow models around agricultural regions.

Motivation for the current study originated from elevated nitrate levels in private and municipal wells within the Black Brook Watershed (BBW), which have been related to intense application of agricultural fertilizers (Chow and Rees 2006). BBW is located in northwestern New Brunswick, and is part of an economically important agricultural region

of the upper St. John River Valley where potatoes are the dominant crop. Municipal and residential drinking water is sourced from the underlying fractured bedrock of the Matapédia Basin which extends throughout most of northwestern New Brunswick, the Gaspé region of Québec, and into northern Maine, USA. In 2003, the village of Saint-Andre, NB which is located within the BBW relocated their water supply well due to elevated nitrate levels with concentrations occasionally exceeding the 10 mg/l NO₃-N maximum acceptable concentration for drinking water (Health Canada, 2012).

Intensive agricultural practices in the BBW have prompted several studies to assess their effects on surface and groundwater quality (e.g. Milburn 1996; Yang et al. 2010; Chow et al. 2011). The majority of the studies focused on surface water quality and soil erosion within the watershed, with little concern for the hydraulic properties of the underlying fractured bedrock aquifer. Three main groundwater research contributions include theses from the University of New Brunswick. Kierstead (1993) utilized drill core and constant head tests in five boreholes to measure fracture characteristics and hydraulic conductivity distributions in the upper 30 m of the aquifer. Results suggested that groundwater flows predominantly through the fracture network, and that hydraulic conductivity tends to decrease with depth. Gallagher (1997) and Valentin (2002) developed a watershed-scale equivalent porous media numerical model for nitrate migration through the relatively shallow fractured bedrock aquifer (5 to 30 m). However, neither study considered anisotropy and both were limited by uncertainty in the fracture distributions and their hydraulic characteristics below 30 m depth.

This study presents a comprehensive fracture network analysis in the folded fine-grained calcareous sedimentary aquifer underlying the watershed with the objective of better

predicting the dominant groundwater flow directions, and providing a first-order approximation for anisotropy into an equivalent porous media model. Such information can be used to better assess the future impact of agricultural practices on groundwater quantity and quality in the watershed. The fracture network was evaluated in boreholes and outcrop exposures using geophysical and geological methods, while hydraulic conductivity distributions were determined through borehole hydraulic testing. A subwatershed-scale groundwater flow conceptual and numerical model was also developed to further validate the hydraulic properties derived from field investigations by parameter estimation and calibration based on stream flow and hydraulic head observations. A similar approach was taken by Rayne et al. (2001) to simulate regional-scale groundwater flow and define capture zones using an equivalent porous media (EPM) model for vertical and horizontal fracture sets in a densely fractured Silurian dolomite in Wisconsin. Similarly, using an EPM approach, Gburek et al. (1999) modelled groundwater flow and agricultural contaminant transport with another similar aquifer in east-central Pennsylvania. Although spatial changes in head were determined in their model, a complete understanding of fracture anisotropy within the flow system was not achieved.

This study also describes relationships between the fold geometry, the contemporary in situ stress field, and observed fracture patterns, and draws attention to their influence on groundwater flow in the study area. Wide-aperture and oxide-stained bedding plane fractures within large-scale open folds provide the preferential groundwater flow pathways, whereas a predictable distribution of high-angle fracture sets provide vertical connectivity. Given the consistency of regional-scale fold axis orientations and contemporary in situ stress field throughout the BBW and surrounding area, the observed fracture patterns can

be broadly extrapolated to other nearby areas provided that variations in bedding dip and stress directions can be determined.

3.3 Study area

The BBW is 7.5 km long and up to 3.5 km wide covering approximately 14.5 km² (Figure 3.1) in the agricultural potato belt of northwestern New Brunswick. Black Brook is a tributary of the Little River, which in turn discharges into the St. John River. Topography is comprised of valleys and minor tributaries feeding into the Black Brook, with elevations ranging from 180 to 260 m above mean sea level. Approximately 65% of the watershed area is devoted to agriculture, with the remainder divided between forest cover (21%) and other land uses (i.e. 14% residential, transportation, wetlands) (Chow et al. 2011). BBW is part of the Upper St. John River Valley ecoregion in the Atlantic Maritime ecozone (Marshall et al. 1999) with a 3.2°C average annual air temperature and 1100 mm/yr precipitation (Environment Canada, 2012).

Overburden sediments consist of loamy lodgement till and minor amounts of ablation till, silt, sand, gravel and rubble normally ranging from 0.5 to 3 m thick (Langmaid et al. 1980; Rampton et al. 1984). Overburden thickness varied from 3 to 5 meters at the boreholes used in this study and was observed to be less than 1 m thick in nearby locations.

3.3.1 Bedrock geology

Bedrock in the study area (Figure 3.1) is situated within the NE-SW trending Aroostook-Percé anticlinorium in the composite Matapédia Basin that extends through central Maine, northern New Brunswick and the Gaspé Peninsula, Québec (Malo 2004). Along much of its length, this anticlinorium is in fault contact with the Chaleurs Bay

synclorium along the McKenzie-Gulch fault (Carroll 2003). In New Brunswick, basin sediments are largely fine-grained turbidite sequences deposited during the opening and closing of the early Paleozoic Iapetus Ocean, prior to accretion and collision of island arc and microcontinental terranes during the Appalachian Orogeny (Van Staal et al. 1998). Bedrock is weakly deformed, showing evidence of very low grade metamorphism, and consists of fine-grained Ordovician and Silurian calcareous and argillaceous sedimentary sequences of the Matapédia and Grog Brook Groups (Carroll 2003; Wilson et al. 2004). Bedrock within the Black Brook Watershed boundary is solely comprised of Matapédia Group sequences, divided into the Pabos Formation conformably overlying the White Head Formation (see Fig. 1; St. Peter 1978; Wilson et al. 2004). These formations are dominated by thin-bedded, dark grey calcareous shale, siltstone, and fine-grained sandstone, calcilutite, minor medium-bedded calcareous sandstone, and minor amounts of non-calcareous shale and siltstone deposited on a submarine slope environment (St. Peter 1978; Stringer and Pickerill 1980). The adjacent Grog Brook Group consists of a series of predominately siliciclastic turbidite deposits, composed of full and partial Bouma sequences, graded bedsets, flute casts, and other sole markings (St. Peter 1978; Carroll 2003).

Structural features observed in Grog Brook and Matapédia sequences north and northeast of the study area consist of minor NW trending, shallowly plunging open folds attributed to the Salinic Uplift (St. Peter 1978; Carroll 2003; Wilson et al. 2004). The NW trending folds are overprinted by a dominant set of NE trending upright open to locally isoclinal folds (shown in Figure 3.1) as a result of dextral transpression during the Acadian Orogeny along the McKenzie Gulch Fault system (Wilson et al. 2004). The resultant set of folds comprise a southeasterly dipping set of axial planes, and moderately to gently plunging

hinge lines towards the NE and SW, characterized as dome and basin shaped structures (Wilson 2002). These fold structures are situated within the NE-SW trending Aroostook-Percé anticlinorium in the composite Matapédia Basin. Fractures associated with the older NW trending set of folds vary from moderately to very well developed, and are oriented approximately parallel to the axial plane of the folds (Carroll 2003). These fracture planes are typically infilled by carbonate minerals, and to a lesser extent of sericite and chlorite (St. Peter 1978).

3.3.2 Hydrogeology

The primary hydrogeological unit in the study area is a highly fractured bedrock aquifer, overlain by a relatively thin layer of glacial overburden. Five vertical wells (BH7, 8, 9, 10 and 11) were drilled in 2008 using an air-rotary method to a diameter of 15.3 cm (6 inches), ranging in depth from 40 to 140 m. The wells were drilled along a NE-SW transect descending a valley wall towards the Black Brook in order to assess groundwater flow and nitrate migration through the overburden and fractured bedrock aquifer. The transect was roughly perpendicular to a line of four monitoring wells (BH1, 2, 4 and 5 in Figure 3.1) drilled in 1993 to depths of approximately 30 m to help better constrain the understanding of local hydrogeological conditions. The four shallow wells had been analyzed by Keirstead (1993) for frequency of fracturing and vertical distribution of hydraulic conductivities, and were used in this study primarily to provide hydraulic head constraints for the numerical model.

Average annual water levels within the study area range from 11.5 to 16 m below ground surface. However, water levels in some wells exhibit seasonal variations of up to 20 m. Significant water bearing zones were encountered at depths of 12 to 15 m and 27 to

33 m respectively during drilling of three of the five boreholes. Pump test results at the new municipal well location provided aquifer transmissivities from seven monitoring wells ranging from 52 to 282 m²/day, with aquifer storativity values between 2.1×10^{-4} and 1×10^{-2} . The range of storativity values indicated largely unconfined conditions ($> 10^{-4}$) with perhaps local confining conditions present (Singhal and Gupta 2010).

3.4 Methods

3.4.1 Geophysical logging and outcrop mapping

Acoustic (ALT ABI40) and optical borehole (ALT OBI40) televiewer logging provided continuous, oriented images (e.g., Figure 3.2) for 380 m of borehole wall from five different holes (diameter of 15.3 cm). Analyses of these images yielded fracture depths, dips and dip directions, and apertures. The optical televiewer was particularly useful to distinguish between open, calcite-filled, and oxide-stained fractures in the saturated and unsaturated zones. The acoustic televiewer travel-times provided caliper measurements used to avoid setting packers on damaged zones to minimize vertical leakage during hydraulic testing. Several of the damaged zones were recognized as drilling-induced borehole breakouts. Resolution of televiewer data was sufficient to identify fractures with apertures as small as 0.1 mm, although aperture estimates below 1 mm were unreliable. It is recognized that fracture apertures may display somewhat larger openings along borehole walls due to enlargements from the drilling process. Recorded fracture dips were corrected for slight borehole deviations from vertical ($< 2^\circ$) and orientations were adjusted using local magnetic declination to true north. To compensate for the sampling bias associated with measuring fracture orientations along vertical boreholes, a Terzaghi correction factor of

$1/\cos\theta$ (where θ is the dip angle at the center of each histogram bin) was applied to the observed frequency of fracture dips (Terzaghi 1965).

To investigate the character and spatial variability of fractures and their relationship to fold trends, detailed structural measurements were recorded at four outcrops located 1 to 3 km south to southwest of the borehole locations (see Figure 3.1). The outcrops provided excellent exposure of several secondary fold structures associated with the NE trending regional scale Aroostook-Percé Anticlinorium. The outcrops consisted of near-vertical road cuts and quarry walls with orientations varying between NS and EW, thereby allowing for unbiased observation of fracture orientations. Data collection consisted of recording fracture orientations relative to bedding planes and fold geometries, and noting their surface attributes, termination relationships, lengths and types (e.g. mineralization) where available. Fracture lengths were only estimated due to rock fall hazard. Most observed fractures extended beyond the extent of the visible outcrop.

Fracture orientations determined from borehole televiewer and outcrop mapping were plotted on equal-area stereographic projections, weighted to account for low probability of intersecting high-angle fractures in the vertical wells, and contoured using Gaussian methods (Robin and Jowett 1986).

Fracture spacings were determined between fractures within the same set observed along the borehole walls. To determine the spacing perpendicular to each fracture plane, S_n , a geometric correction was applied using (Priest 1993):

$$S_n = S \cos \theta \quad (3.1)$$

where S is the spacing observed along the borehole wall, and θ (deg) is the mean dip of the fracture set.

Other continuous geophysical logs recorded in the five boreholes to confirm lithology and assist in hydraulic interpretation included natural gamma, self-potential, and single point resistance.

3.4.2 Hydraulic testing

Constant head tests were completed by Environment Canada in four boreholes by injecting water at a constant pressure into discrete intervals confined by two inflatable packers, set 3.3 m apart in BH7, 8, and 11 and 2.2 m apart in BH10. Flow rates were measured following stabilization of the hydraulic head monitored using an electronic pressure transducer installed in the test interval. A total of 96 tests were completed, with 29 intervals insufficiently permeable to allow water to be injected. The remaining tests examined 44 unique intervals, and included 23 repeats to confirm accuracy. These were typically in intervals of higher interest or lower hydraulic conductivity. The standard error between repeat measurements was roughly half an order of magnitude. Tests were initiated several minutes after the packers were inflated, allowing time for the test interval to stabilize before fluid was injected. Durations of the tests were dictated by the time required for the hydraulic head to stabilize and ranged from one minute for extremely tight intervals to roughly 30 minutes for the more transmissive intervals.

Interval hydraulic conductivities K were calculated using the Thiem equation with measurements of the pressure head and flow rate at steady state (Ziegler 1976; Bear 1979),

$$K = \frac{Q}{2\pi\Delta H b} \ln \frac{r_o}{r_w} \quad (3.2)$$

where Q is the flow rate (m^3/s), H is the applied variation in head from ambient (m), b is the test interval length (m), r_o is the radius of influence of the test (m), and r_w is the well

radius (m). Because the radius of influence (r_i) approximation is within the natural logarithmic function, it has a minor influence in the outcome of the hydraulic conductivity calculation (Doe and Remer 1980). A radius of influence of 10 m was assumed for these calculations.

Additional hydraulic conductivities determined by Keirstead (1993) in 3 m packer intervals were available for four boreholes drilled to a depth of 30 m into the fractured bedrock along the southern boundary of the area selected for numerical modelling.

3.5 Results

3.5.1 Geophysical borehole logging

A composite of selected geophysical logs from BH7 is presented in Figure 3.3. The acoustic televiewer travel-times were used to calculate borehole diameters and identify intervals characterized by a high density of fracturing and drilling-induced borehole breakouts. The borehole breakouts are predominantly oriented ESE-WNW in all five boreholes, indicating the maximum compression (σ_1) of the contemporary in situ stress field is oriented in a NNE-SSW direction ($025^\circ - 205^\circ$), which is broadly consistent with other measurements in the region (Sbar and Sykes 1973, Heidbach et al. 2008). The natural gamma (NG) and single point resistance (SPR) tools were used to discriminate between lithological units. The bottom of the well (80 to 100 m) shows a higher NG response corresponding to high shale content. This interval also shows a subtle increase in self-potential (SP) and decrease in the SPR, as well as high variability in the acoustic caliper. The NG between 23 to 37 m depth shows low response consistent with low shale content. This interval is coincident with increased variability in borehole diameter associated with a

higher fracture count, and increased hydraulic conductivity values. The abrupt change in SP and SPR at 33 m corresponds to several open bedding plane fractures observed within a short interval, many of which show extensive oxidation on the fracture planes which may suggest groundwater flow. The interval from 37 to 80 m shows relatively smooth geophysical-log responses and is associated with a low fracture count. Similar borehole log characteristics are observed in the other wells.

3.5.2 Fracture network characteristics

The fracture network observed in the folded sedimentary bedrock underlying the BBW consists of four main fracture sets, which can be identified on the basis of their orientation, infilling, aperture, and relation to bedding. These fracture types consist of (i) two open aperture high-angle fracture sets that strike sub-parallel to and sub-perpendicular to fold trends, (ii) a calcite-filled high-angle fracture set, and (iii) low-angle bedding plane partings. The term “high-angle” refers to the fracture angle with respect to horizontal. Borehole televiewer images revealed 361 high-angle fractures and 164 bedding plane fractures in five vertical boreholes. The histogram of fracture dips observed in boreholes (Figure 3.4a) shows a bimodal distribution peaking at roughly 30° and 75° angles, corresponding to bedding plane and high-angle fractures, respectively. The Terzaghi corrected histogram displayed in Figure 3.4b shows a distribution that is strongly dominated by high angle dips, which is consistent with fracture dips observed in outcrop (Figure 3.4c).

Stereonet and rose plots of fracture orientations from borehole and outcrop measurements (Figure 3.5 and Figure 3.6) provide evidence for four discrete fracture sets. The stereonet plots show variability in dip and dip direction of fracture sets between separate borehole and outcrop locations. Rose diagrams highlight the well-defined strike

directions of each fracture set separately, and the mean fracture set trends for the five boreholes (Figure 3.5). Rose diagrams presented in Figure 3.6 similarly show the strike directions of each fracture set, as well as mean fracture sets for each outcrop. Properties of each fracture set are listed in Table 3.1.

Bedding plane fractures (S1) observed in the boreholes strike in a NE-SW direction and exhibit a mean dip of 31° (with a standard deviation of 9°) towards the southeast. Measurements of dip and dip direction between the boreholes (Figure 3.5) reveal subtle variability in the trend of local-scale folding. This is also evident as a small clockwise rotation of $\sim 20^{\circ}$ in bedding strike along a transect from BH7 to BH11, which may be attributed to interference patterns around the edges of the doubly plunging folds. Despite this subtle variability, the consistent dip direction towards the SE indicates that all of the boreholes penetrate only one limb of a fold.

An abundance of weathering and oxidation along bedding plane fractures identified in borehole surveys provides evidence of significant groundwater flow (Nelson 2001). A number of high-flow intervals were detected during borehole drilling by loss of drilling water corresponding to intersections with large aperture bedding plane fractures with the presence of oxidation staining. These fractures were commonly observed near the surface and tend to gradually decrease in abundance with depth.

In outcrop, a wider range of bedding dips were observed, as steep as 60° towards the SE and NW, which exposed multiple fold axes with similar NE trends. Individual beds range from <1 cm to decimeters in thickness, and extend laterally beyond the length of the exposed outcrop. Several wide-aperture fractures were discerned where calcite veins emplaced along bedding planes have been largely dissolved, presumably as a result of

groundwater flow. Figure 3.7a, for example, shows remnants of a calcite vein overlying a thin bed having a wavy surface and deformed laminations that are suggestive of a primary loading structure (e.g. ball and flame structure) formed in very soft sediments. We speculate that calcite infilling occurred during or following folding and flexural dilation along the bedding plane, where fracture dilation may have been enhanced by the rheological contrast and irregular character of the sediment contact. The wavy texture may have in turn been amplified by the development of cleavage during folding (A. Park, University of New Brunswick, pers. comm. 2013). Slickenlines were observed on a number of bedding surfaces suggesting that flexural slip movement occurred during folding. Rare pressure dissolution features, such as stylolites were also observed in thin sections and core. These features are generally characterized by irregular surfaces and act to enhance fluid flow pathways (Heald 1959). Due to their rarity however, stylolites are probably not important contributors to groundwater flow in this area.

High-angle fractures were the most abundant type observed (48%) despite the greater chance of intersecting low-angle fractures in vertical wells. These fractures were subdivided into two distinct fracture sets striking sub-parallel (F3) and sub-perpendicular (F2) to the regional NE structural trend (Figure 3.5). The F3 fracture set, accounting for 61% of the observed open high-angle fractures, includes two subsets with similar strike directions but opposing dip directions, with mean dip angles of 67° towards the ESE and 71° towards the WNW. These subsets are interpreted to form a conjugate subset intersecting the SE dipping limb of a fold with a 42° dihedral angle. As shown in Figure 3.5, the F3 fracture set strikes uniformly NNE along a transect from BH7 to BH11, although rotations in the strike of bedding (discussed above) show that the fold geometry varies locally. The F3 fracture

set strikes sub-parallel to regional fold trend, being oriented $\sim 30^\circ$ more northerly than the mean bedding strike observed in boreholes (Figure 3.5). This lack of congruence with the fold trend, together with the fact that F3 fractures exhibit two dominant dip angles, only one of which is near orthogonal to bedding indicates that the F3 set did not form simply as a consequence of fold development. The uniformity of their strike and near vertical dips despite slight changes in bedding orientation from one borehole to the next suggests that they likely formed after fold development.

The remaining open high-angle fractures (F2) exhibit a mean strike of WNW with a mean dip of 57° towards the NNE, with a minor subset dipping towards the SSW. The F2 set shows azimuthal variation in strike between adjacent boreholes, with a counter-clockwise rotation of approximately 20° along the transect from BH7 to BH11. This rotation is similar in magnitude but opposite in direction to the bedding plane rotation observed along the same transect. The fact the F2 fractures rotate whereas the F3 fractures do not, suggests that these two fracture sets developed at different times under differing stress conditions.

The F2 and F3 high-angle fracture sets were similarly represented in four nearby outcrop exposures (see Figure 3.1a), where both sets are composed of conjugate subsets with similar strikes, but opposing dip directions (Figure 3.6). Each of the conjugate subsets show a slightly broader spread in their orientation compared to the borehole fractures, likely attributable to measurements being made on fold limbs exhibiting a wider range of dips to both the NW and the SE. The strike directions of the F2 and F3 fracture sets are rotated roughly 15° counter-clockwise compared to their orientations in the boreholes 2 to 3 km

away. Despite these subtle differences, the dominant fracture sets are similarly represented between all outcrop and borehole locations.

In outcrop, the F2 and F3 fractures exhibit characteristics of regional scale joint sets, where such extensional characteristics could not be clearly discerned in boreholes. Both fracture sets possess smooth surfaces with the presence of well-developed plumose structures, and no appreciable offsets related to shear movement. These fracture sets extend near vertically through bedding planes at the same high-angle, regardless of the location along the fold geometry. The lack of offsets in F3 fractures across bedding plane partings suggests these fractures post-date bedding slip. This observation, coupled with the lack of congruence between strike of fractures and the fold geometry, provides evidence that the fractures are not genetically related to folding, and probably post-date folding (e.g. Hancock and Engelder, 1989). The high-angle fractures are observed to cross-cut bedding surfaces and, in most cases, extend for tens of meters, typically beyond the limits of the outcrop exposure (Figure 3.7b), which is consistent with a non-stratabound fracture system (Odling et al. 1999). This is expected to enhance vertical connectivity in the aquifer, between bedding plane fractures at different depths.

A less pervasive NW trending, high-angle fracture set (F1) was identified in borehole and outcrop surveys, dominated by infillings of carbonate minerals, predominantly calcite, with minor amounts of quartz. The majority of fractures in this set are oriented roughly perpendicular to the regional NE structural trend, and tends to appear in clusters, with a smaller population of fractures trending in a NE direction. The clusters are typically a few meters in length, separated by intervals with no calcite-filled fractures. Their occurrences do not appear to correlate with variations in lithology along the borehole, and their mean

orientation differs from that of the other high-angle fractures (Figure 3.5). However, similar to the F2 and F3 fracture sets, these fractures were observed in outcrop to cross-cut bedding planes and extends through tens of metres in length. The calcite filled fractures show minor offset by bedding plane flexural slip. In spite of being filled with calcite, few of these fractures are characterized by lateral shear movement evident by slickensides and slickenfibres along the fracture surfaces, and small amounts of displacement observed in outcrop. Because these fractures are filled with minerals, they are unlikely to contribute significantly to groundwater flow in the aquifer but may provide a small amount of aquifer storage which may be removed during prolonged dry or pumping periods.

Folds ranging from a few meters to hundreds of meters in wavelength were observed at the four outcrop locations. Based on measurements from all outcrops, the regional fold structure in the area has an approximate NE trend with a mean hinge line plunge of 13° towards SW, where the axial plane between the fold limbs is near vertical. The fold limbs have mean dips of 51° towards the SE and 50° towards the NW, with an interlimb angle of 79° (Figure 3.6). A second order synclinal fold observed at outcrop P3 comprises open-fold geometry with bedding dip changing across the fold axis with limbs dipping towards the NW and the SE (Figure 3.6— see outcrop P3). Although not measured, fracture intensity across the fold was observed to increase towards the axial plane, reflecting mainly an increase in both F3 fracture subsets.

The schematic diagram in Fig. 3.8 illustrates a conceptual model for the non-stratabound fracture network in this area. The schematic diagram shows the bedding plane fractures (S1) associated with the SE dipping limb of a fold, intersected by the high-angle fracture sets (F2 and F3). The SE dipping limb is intersected by boreholes BH7, 8, 9, 10,

and 11 drilled parallel to the regional fold axis trend. The NW dipping limb and fold hinge are likely located further to the NW of the current boreholes, but the distance is unknown. Fracture spacing and apertures, as well as hydraulic conductivity results are discussed below in order to identify fracture sets that dominate the flow system, for incorporation into a numerical model to simulate flow.

3.5.3 Fracture spacing and aperture

Borehole televiewer logs provided data sets for the quantification and statistical analysis of fracture spacings and apertures. Spacing data for each of the fracture sets as measured in boreholes were best represented by log-normal distributions. The geometric mean spacings (as recommended by Limpert (2001) for log-normal distributions) range from 0.21 m for the calcite-filled fractures, to 0.42 and 0.61 m for the high angle F2 and F3 fractures, to 0.87 m for the bedding plane fractures (Table 3.1).

Table 3.1. Statistics used to describe the best fit between the fracture spacing data for fracture sets recorded in boreholes.

Fracture set	Number of fractures	Fracture spacing geometric mean (m)	Fracture strike	Distribution	Arithmetic mean (m)	Standard deviation (m)	Coefficient of variation
F1 calcite-filled	110	0.21	138°SE	Log-normal	0.86	1.83	2.13
F2 high angle	98	0.61	109°SE	Log-normal	1.53	2.54	1.66
F3 high angle	153	0.42	21°NE	Log-normal	1.09	1.54	1.41
S1 bedding fractures	164	0.87	51°NE	Log-normal	1.79	3.12	1.74

The calcite-filled fractures are tightly spaced within clusters separated by tens of meters in the boreholes. This is reflected by their relatively high coefficient of variation, defined as the ratio between the standard deviation and the arithmetic mean spacing for the fracture set (Cox and Lewis 1966; Gillespie et al. 1993, Odling et al. 1999). Indeed the

coefficients of variation exceed unity for all four fracture sets suggesting that each set exhibits some degree of spatial clustering.

Measured apertures, corrected for fracture dip relative to the vertical borehole wall, ranged from < 1 mm to >50 mm, with 51% less than 1 mm. Because of their diminutive width, these fractures are thought to be insignificant with respect to the overall transmissivity of the well. However, documentation of their presence helps to determine the proper statistical distribution function and aids with the calculation of the mean fracture set spacing and orientation. Calcite-filled fractures were assigned an aperture of zero and were not included in the aperture analysis. Apertures for individual fracture sets were determined to be best fit by log-normal distributions, with geometric means ranging from 0.71 mm to 1.51 mm (Table 3.2). Statistical analyses on apertures in other studies generally have been described as following log-normal distributions (Snow 1968, Lemieux et al. 2009), although exponential and power law distributions have also been observed (Barton and Zoback 1992).

Equivalent hydraulic conductivity of the individual fracture sets were calculated by assuming uniform aperture and spacing (equal to their geometric means), and that the matrix does not contribute significantly to groundwater flow. Research in other fractured aquifers suggests that a relatively small percentage of fractures contribute to the flow system (e.g. Michalski and Britton 1997), therefore resulting in a possible overestimation of equivalent hydraulic conductivity for fracture sets. However, such an approach is valuable for providing initial estimates of hydraulic conductivities and anisotropy ratios, for comparison to various bulk hydraulic conductivity measurements and refinement through

flow model calibration (Long et al. 1982). Equivalent hydraulic conductivity for each fracture set was estimated using the cubic-law relationship (Domenico and Schwartz, 1990):

$$K = \frac{\rho_w g N b^3}{12\mu} \quad (3.3)$$

where ρ_w is water density (kg/m^3), g is gravitational acceleration (m/s^2), μ is the viscosity of water ($\text{kg/m}\cdot\text{s}$), b is the mean fracture aperture for the fracture set (m), and N is the mean fracture density of the fracture set expressed in fractures per meter (m^{-1}). Equivalent hydraulic conductivities for individual fracture sets range from 3.3×10^{-6} to 7.0×10^{-7} m/s, which is similar to values determined in the constant head packer tests presented below. The bedding plane fracture set which has the greatest mean aperture is predicted to be five times more hydraulically conductive than the two open high angle sets.

Table 3.2 Equivalent hydraulic conductivity estimates based on geometric mean fracture spacing and aperture.

Fracture sets	Geometric mean spacing (m)	Geometric mean aperture (mm)	Hydraulic conductivity (m/s)
F3 high angle	0.42	0.71	7.0E-07
F2 high angle	0.61	0.81	7.2E-07
S1 bedding fractures	0.87	1.51	3.3E-06

To evaluate the influence of fracture type on groundwater flow, apertures were grouped and plotted on a stereographic projection to show their dominant orientations (Figure 3.9). Aperture groups consist of small ($< 1\text{mm}$), medium ($1 - 5\text{ mm}$), large ($5 - 10\text{ mm}$), and very large ($> 10\text{ mm}$) fracture widths. Fractures with apertures greater than 5 mm (125 fractures) are tightly clustered on the stereonet and are predominantly associated with 86 bedding plane fractures that are moderately dipping towards the southeast in the limb of the fold penetrated by the five geophysically logged boreholes. The remaining open fractures from this group correspond to 39 high-angle fractures predominantly

comprising a mix of F2 and F3 sets. The abundance of wide aperture bedding plane fractures, combined with their large lateral extent observed in outcrop suggests that these fractures are likely to function as an important groundwater flow pathway. Since the bedding plane fractures comprise a shallow to moderate dip angle, their behavior is similar to the high-angle fractures in that they may provide an additional source of azimuthal anisotropy on the aquifer system. A number of these open fractures (90% of those with apertures greater than 5 mm) show significant oxidation and weathering along the fracture planes, predominantly corresponding to bedding plane fractures, followed by F3 high angle fractures. Oxide-staining along fracture planes may correspond to prolonged groundwater flow (Nelson 2001). Dominant orientations of these fractures are presented on a stereonet and rose diagram plot in Figure 3.10. The stained bedding plane fractures have a mean dip of 31° towards 143°SE , and strike roughly NE-SW. Similarly, the stained high-angle fractures tend to be dominated by a NNE-SSW (F3) strike orientation. Fracture spacing measured between the oxide-stained fractures (irrespective of orientation) fit well to a log-normal distribution and are characterized by a geometric mean of 0.71 m. From these results we infer that groundwater flow is controlled by the combined orientations of the transmissive bedding plane and high-angle fractures, resulting in the introduction of aquifer anisotropy along a NE–SW orientation, which is consistent with the orientation of regional fold axes. Groundwater flow is assumed to be less-significant in the NW-SE direction because of the relatively low density of high-angle fractures aligned in that direction with large aperture widths, and because a large percentage of those fractures are calcite-filled.

3.5.4 Hydraulic conductivity distribution

Constant head injection tests in four boreholes indicated that hydraulic conductivity values varied over 6 orders of magnitude from 1.2×10^{-9} to 1.3×10^{-3} m/s (1.1×10^{-4} to 110 m/day). Within an individual borehole, conductivities generally decreased by about four orders of magnitude with depth, except in BH10, which exhibited a relatively uniform vertical distribution associated with a higher degree of fracturing throughout the length of the well. Mean hydraulic conductivity from the four tested wells was 4.45×10^{-5} m/s. This is comparable to estimates of 9.6×10^{-9} to 1.3×10^{-4} m/s in the upper 30 m of bedrock determined previously by Kierstead (1993) and to estimates of 10^{-5} to 10^{-4} m/s obtained from nearby pump tests in a municipal well (Neill and Gunther 2003). However, conductivities between neighbouring test intervals varied by up to three orders of magnitude associated with changes in frequency and aperture widths of open fractures. Approximately 30% of the tested intervals had hydraulic conductivities below the minimum testing limit (10^{-10} m/s). Most of these intervals were associated with lower observed fracture frequencies.

The relationship between hydraulic conductivity and depth observed in the four tested wells contributes to a conceptual model for the aquifer. Figure 3.11 presents 67 constant head injection test results plotted against depth, with a five meter moving window used to calculate a mean hydraulic conductivity profile for all the results. Results from the conductivity profile show a general decrease in hydraulic conductivity with depth by approximately two orders of magnitude from the water table to a depth of 70 m below surface, with a clear discontinuity at 40 m depth. The conductivity decreases by an additional order of magnitude below 70 m. The conductivity profile enabled the aquifer to

be represented by three discrete hydraulic layers, which were incorporated into the numerical model. It is important to note, however, that only one of the four boreholes tested (BH7) extends below 70 m.

Figure 3.12 shows that the decrease in hydraulic conductivity with depth in borehole BH7 does not show a direct relationship with the degree of fracturing measured in boreholes. However, it does correspond with a decrease in the frequency of fractures exhibiting oxidation and weathering. Oxidation is present on fracture planes to depths up to 100 m suggesting that the resulting alteration of the fracture surface is likely not associated with fluctuating groundwater levels, but is more likely a result of enhanced groundwater flow carrying dissolved oxygen to depth along open fracture planes. These oxide-stained fractures are assumed to reflect the smaller percentage of total fractures that contribute directly to flow within the aquifer.

Fractures that intersect the packer test intervals were visually inspected using the borehole televiewer images to identify those likely to be acting as primary groundwater flow pathways in the test interval. The relative contribution of each fracture to the total K of an interval was estimated by dividing the fracture's effective K (based on its aperture and Snow's law) by the sum of the effective K 's for all fractures in that same interval. In this case, all of the fractures are assumed to contribute to flow, though only a small percentage of the fractures may actually be conductive (e.g. Morin et al. 1997; Michalski and Britton 1997; Boutt et al, 2010). Given that Snow's law models K as being proportional to aperture squared (Witherspoon et al. 1980), the approximation for K_i reduces to the following,

$$K_j = K \cdot \frac{b_j^2}{\sum_{j=1}^n b_j^2} \quad (3.4)$$

where b_j is the aperture of each open fracture and K is the total hydraulic conductivity of the interval. Based on this methodology, Fig. 3.12 shows a combination of bedding plane and high-angle fractures for BH7 contributing to the aquifer system. Hydraulic conductivities at 46 m and 69 m depth and in the interval between 20 m to 28 m are predominantly influenced by high-angle fractures. In contrast, the two intervals between 28 and 33 m, and the one from 78 to 81 m depth appear to be influenced primarily by bedding plane fractures. Of the inspected fractures, approximately 60 percent show significant signs of oxidation and weathering, mostly within the upper 40 m of bedrock.

3.6 Groundwater flow numerical modelling

A 3D groundwater flow model was developed to investigate the validity of the conceptual model for the fractured aquifer underlying the BBW, incorporating both anisotropy and variability of hydraulic conductivity with depth. An equivalent porous media (EPM) approach was selected to represent the combined effects of individual fractures and the rock matrix. This approach has commonly been used to simulate flow in fractured media and can be considered to provide an acceptable approximation of the aquifer (e.g. Khaleel, 1989; Gburek et al. 1999; Rayne et al. 2001; Scanlon et al. 2003), particularly where fractures are closely spaced and well interconnected relative to the size of the grid spacing and the thickness of the model layers. This is the case in the study area where fracture sets are roughly perpendicular and are all represented by a mean fracture spacing less than 1 meter.

Similar EPM models have successfully been employed in fractured aquifers of the Maritimes Basin in Atlantic Canada (e.g. Rivard et al. 2008, Jiang et al. 2004). In addition, previous efforts to model groundwater flow and contaminant transport in BBW also used the EPM approach (e.g. Kierstead 1993; Gallagher 1997, Craig HydroGeoLogic 1995, Valentin 2002; Li 2013), under the assumption that the aquifer behaves as an EPM due to the significant interconnectivity of the fractures.

The steady state flow model was developed in Visual MODFLOW (Schlumberger Water Services) with the model domain covering 2.6 km² of the Black Brook Watershed (see Figure 3.13). The southwestern end of the domain was defined as a constant head following the linear gradient of average measured groundwater levels between boreholes BH2 and BH5, while the northeastern boundary is delimited by the watershed limit set as impermeable. The southeastern and northwestern limits of the domain model were set as impermeable boundaries, since flow through these boundaries was expected to be minimal, as a result of the NE dominant flow direction. During the calibration process these two boundaries were also switched to constant (i.e. prescribed) head; however this change had little effect on the groundwater flow direction as well as on the water balance terms. Ground elevations of the domain were defined using a 10 m horizontal grid extracted from a provincial digital terrain model (DTM) (SNB, 2011) interpolated to a 30 m grid in Visual MODFLOW. Based on the results from hydraulic testing, three hydrostratigraphic layers were defined, with a thickness of 40 m for the upper layer (i.e. 0 – 40 m bgs), 30 m for second layer (i.e. 40 – 70 m bgs), and a fixed elevation of 0 m for the base of the bottom layer (i.e. 70 m bgs – msl). The second and third layer elevations were derived by

reinterpolating the DTM to a lower resolution (i.e. 100 m cell grid for the second layer and 500 m cell grid for the third layer).

MODFLOW allows only for the specification of hydraulic conductivities in orthogonal directions aligned with the three coordinate axes. Therefore the x-axis was aligned with the mean strike (051°) of the S1 bedding fractures. For each layer, average hydraulic conductivities obtained from the constant head tests were assigned as initial hydraulic conductivity values in the x-direction (K_x), while the average value from application of the cubic-law relationship to the F2 and F3 high angle fracture sets was assigned as the initial conductivity values in the z-direction (K_z) for all layers. Fractures trending in the y-direction consist of F1 and F2 high-angle fractures, and consequently, hydraulic conductivities in the y-direction were set to be five times lower than K_x for each layer, based on the ratio between K_x and K_y from application of the cubic law. The predominance of calcite-filled fractures, smaller fracture apertures, and a limited number of oxide-stained fractures suggests that groundwater flow in this direction may not be as significant compared to the x-direction. Initial hydraulic conductivities and anisotropy were calibrated both manually and by using the inverse parameter estimation during the model run. The MODPATH algorithm (Pollock 1998) was used for tracking three-dimensional flow paths through the aquifer.

Recharge to the aquifer was set to 525.5 mm/yr based on the daily groundwater water level data from the five wells drilled for this study and the groundwater discharge (i.e. stream baseflow) to Black Brook as presented in Danielescu et al. (2012).

The model was calibrated such that the differences between measured head in observation wells and simulated heads, as well as differences between the simulated

groundwater discharge to Black Brook and stream baseflow were minimized. For this simulation, three of the drilled wells (BH7, BH9, and BH10) and three domestic wells (DW16, DW18 and DW19) were used to assess the model performance. Data from two additional wells were available but were not used during the calibration. BH8 was located in the same grid cell as BH7 and therefore only one of the wells were selected to be used in the calibration. The average measured groundwater level in BH11, located 130 m southwest from the Black Brook, was about seven meters below the water levels measured in the stream. This difference could be attributed to geological features not considered in this study (e.g. presence of an impermeable layer between the Black Brook and the well), and consequently this well was not used for calibration. During the calibration, hydraulic conductivity (K_x , K_y , K_z) in each layer was adjusted manually, and Parameter Estimator package (PEST) was used for the refinement of final hydraulic conductivity values.

3.6.1 Numerical modelling results

The normalized root mean square error (NRMSE) between measured and calculated heads for the final model calibration was 12.9% compared to 81.1% for the initial run, while the absolute residual mean decreased from 10.6 m to 2.3 m. The calculated groundwater discharge to Black Brook was 0.212 m³/s compared to the estimate of 0.225 m³/s derived from hydrograph separation (i.e. 5.7% difference). The final values of hydraulic conductivity, shown in Table 3.3, are on the same order of magnitude as those reported for fractured carbonates (Singhal and Gupta 2010) and are similar to the results from hydraulic packer tests. Calibrated results show a decrease of one order of magnitude in K_x for the uppermost layer (L1) compared to the initial estimates. During calibration, various ratios between K_y and K_x were tested, including the isotropic case (i.e. $K_y/K_x=1$), however

increasing the anisotropy of the flow system significantly improved the model fit (NRMSE 18.5% for $K_y/K_x = 1$, 16.3% for $K_y/K_x = 0.2$ and 12.9% for $K_y/K_x=0.01$, respectively). The model is most sensitive to changes of K_x in layer 1 followed by changes of K_x in layer 2, whereas changes in conductivities in other directions and layers have little effect on the fitness of the model. For example, changing K_x in layer 1 from $5E-6$ to $3E-6$ m/s results in a change of NRMSE from 12.9 to 39%, which is equivalent to the NRMSE resulted by changing K_x for layer 2 by three orders of magnitude. Similarly, the model was not sensitive to changes in K_z during calibration, such that changes of over four orders magnitude had little to no impact on the flow budget. This could be attributed to the significant thickness of each layer coupled with the influence in the K_x and K_y changes with depth, resulting in vertical movement within each model layer and increasing horizontal flow near the layer boundaries, evidenced by the particle tracking simulation (Figure 3.13). The resulting decreases in K with depth are consistent with the trend identified in hydraulic test results (Table 3.3).

Table 3.3 Hydraulic conductivities used for the groundwater numerical flow model.

Model Layer	Initial values			Calibrated values		
	K_x (m/s)	K_y (m/s)	K_z (m/s)	K_x (m/s)	K_y (m/s)	K_z (m/s)
L1	7.5E-05	1.5E-05	7.1E-07	5.8E-06	5.8E-08	7.1E-06
L2	5.2E-06	1.0E-06	7.1E-07	5.3E-08	5.3E-10	7.1E-07
L3	5.2E-08	1.0E-08	7.1E-07	5.1E-09	5.1E-11	7.1E-08

For the top layer, 92% of the groundwater flows in the x-direction either towards the brook (65%) or the constant head boundary (27%). The remaining 8% ($280 \text{ m}^3/\text{d}$) is lost due to evapotranspiration. The flux exchange between the two upper layers of the system is very small ($31 \text{ m}^3/\text{d}$), while the exchange between the second and the bottom layer is even

smaller ($11 \text{ m}^3/\text{d}$). The best fit between the simulated and observed head values for all wells is shown in Figure 3.13.

3.7 Discussion

Three fracture types were identified corresponding to open high-angle fractures, calcite-filled high-angle fractures and moderately dipping bedding plane fractures. Together, these fractures form a well-developed non-stratabound network with fracture spacings ranging from 0.21 m to 0.87 m and lengths that extend for tens to hundreds of meters. To our knowledge, few studies have reported hydraulic properties of non-stratabound fracture networks in layered sedimentary formations (e.g., Odling et al. 1999; Gillespie et al. 2001; Gross and Eyal 2007), although these fracture networks have been commonly studied from a structural point of view (Odling et al. 1999). Lemieux et al. (2006) concluded that sets of relatively short high-angle fractures played a minor role in a regional groundwater flow system, acting locally as vertical pathways between adjacent bedding plane fractures; the bulk of groundwater flow was confined to particular bedding plane fractures which possessed enhanced hydraulic conductivity. This observation differs from high-angle fractures identified in the BBW, which persistently extend for tens of meters in length (commonly beyond the top and bottom of the outcrops). In this study, the high-angle fractures extend through multiple bedding plane fractures, and are therefore expected to contribute to vertical permeability of the flow system and enhance connectivity at depth. Other studies show similar results where the distribution of high-angle fractures, which formed as extensional joints, improves the vertical connectivity of the bedrock (e.g. Singhal and Gupta 2010).

High-angle fractures in the BBW site are interpreted to be joint sets having three dominant strike directions, reflecting temporal changes in the horizontal in situ stress direction during multiple deformation stages in the northeastern Appalachian belt. They may have formed from any of a variety of mechanisms resulting in paleotectonic or neotectonic, hydraulic, unloading and release fractures (e.g. Engelder 1985). Tectonic fractures generally trend parallel to the direction of maximum compressive stress at the time of propagation, and in some cases can be normal to the strike of regional structure (i.e. fold belt). Hydraulic fractures form under increased vertical compression from burial, resulting in increased pore fluid pressure, and are typically in-filled with minerals (i.e. calcite). Fractures related to uplift and erosional unloading have orientations reflecting the principal stresses during the uplift. In some cases, their preferential orientation may be controlled by a pre-existing structural fabric (fold axes, cleavage, foliation), providing the illusion of a causative relationship.

The F3 fractures have a strong preferential direction to the NNE, which is perpendicular to the orientation of borehole breakouts, and sub-parallel to the regional- and local-scale fold axes. Regardless of the changing strike of local-scale fold geometry, the strike of the F3 fracture set remains uniform towards the NNE. This disparity as well as the through-going nature of fractures suggests that the F3 set is not related to the fold development in this area. Despite the lack of cross-cutting relationships observed in the field, the F3 fracture set likely formed most recently as neotectonic joints, aligned parallel to the current direction of maximum compressive stress as inferred from the orientation of borehole breakouts. This late-formed fracture interpretation is consistent with the relationship between other NE-trending fractures and the contemporary stress field across

the Appalachian belt (e.g. Nickelsen and Hough 1967; Engelder and Geiser 1980; Engelder 1982), where Irving (1977) suggested the current stress field initiated no longer than 70 Ma. The ESE-WNW oriented borehole breakouts in all five boreholes identify the maximum compression (σ_1) of the contemporary stress field oriented in a NNE-SSW direction ($025^\circ - 205^\circ$). This orientation is broadly consistent with other regional stress field measurements in the area (i.e. Sbar and Sykes 1973; Adams 1989; Heidbach et al. 2008) and consistent with the trend of the F3 fracture set. Alternatively, Faure et al. (1996) interpreted a NNE-directed paleostress direction during the latest stage of Alleghenian compression, which may have resulted in NNE-trending F3 fractures. Although this relation is plausible, the NE trending folds would have presumably formed prior to fracture propagation.

Although timing relationships are difficult to discern, both the WNW trending F1 and F2 fractures are suggested to be related to an earlier paleostress field, likely associated with WNW-directed compression during the Acadian or Alleghenian orogenies. These fracture orientations are consistent with WNW paleostress directions calculated by Faure et al. (2004) for the Acadian orogeny in northern New Brunswick, as well as the WNW paleostress directions calculated by Faure et al. (1996) for the later Alleghenian compression. These paleostress directions are generally consistent with the WNW to NW strike direction of the F1 and F2 fracture sets. Kirkwood et al. (2001) proposed that calcite-filled fractures in the Gaspé region of Québec were developed during burial and concurrent Acadian compression. This Acadian deformation resulted in the development of NE-trending folds and regional-scale dextral strike-slip faults in the region (Carroll 2003; Faure et al. 2004). The counter-clockwise rotation observed in the F2 fracture pattern, in particular along the transect from BH7 to BH11, might indicate a change in the direction of

paleostress field during late compression within the Appalachian. This interpretation may suggest a possible late paleostress field rotation, or may reflect secondary stress directions as a result of local perturbations of the regional stress field associated with specific geological or tectonic features (e.g. Zoback 1992). For example, a uniformly directed regional stress field may have been modified by the bedding attitude of the doubly plunging fold structures.

The F1 fractures may have formed and infilled during an early deformation phase pre-dating folding the development of F2 and F3 fractures. Evidence for this is provided by calcite-filled fractures being cross-cut by F2 and F3 fractures, as well as minor offset by flexural-slip movement along S1 bedding plane fractures. Similar filled-fractures are typically attributed to propagation during hydraulic fracturing in the presence of mineral-rich fluids (e.g. Beach 1977). Mineral infilling is assumed to limit the hydraulic conductivity and groundwater flow through this fracture set.

Fractures elsewhere in the Appalachian belt consist of similar fracture patterns and interactions with fold axis trends and contemporary stress directions (e.g., Geiser and Engelder 1983). Both Scheidegger (1981) and Engelder (1982) suggest there is a genetic relationship between the regional fracture trends and the contemporaneous stress field within eastern North America, where they are found to propagate parallel to the principal stress plane of the maximum horizontal stress.

High hydraulic conductivity, abundance of oxide staining and weathering, and great lateral extent provide evidence that bedding plane fractures represent the most important feature controlling groundwater flow in the aquifer underlying the BBW. Large apertures typically measured for bedding plane fractures may be the result of flexural-slip movement

during Acadian fold development, leading to vertical extension of the bedding plane (Price and Cosgrove 1990). Fractures showing significant weathering and oxidation are deemed the most important to the overall groundwater flow system, most of which are bedding plane fractures (S1), and to a lesser degree F3 high-angle fractures. The strike of these two fracture sets is predominantly oriented in a NNE to NE direction, parallel to the maximum principal stress direction. As a result these fractures are held open under extensional stress conditions, thereby enhancing their hydraulic conductivities. Similar relationships between contemporary stress field and groundwater flow pathways have been documented in the Newark Basin, within the Appalachian belt (Morin and Savage 2001). This interpretation is validated by the common occurrence of oxidation and weathering of bedding plane fractures coincident with the enhanced hydraulic conductivity determined in the packer test intervals. As well, high-flow intervals detected during borehole drilling largely correspond to intersections with large aperture bedding plane fractures and few F3 high-angle fractures with an abundance of oxidation staining. Hydraulic conductivities from packer tests corresponding to the high-flow intervals range between 10^{-5} and 10^{-3} m/s, which are among the highest of the 67 measured intervals. Several studies have found bedding plane fractures to be important pathways in flat-lying fractured sedimentary aquifers (Merin 1992; Michalski and Britton 1997; Morin et al. 1997; Muldoon et al. 2001). Merin (1992) concluded that groundwater flow in a relatively homogeneous sedimentary basin in the Appalachian Plateau occurs predominantly along bedding plane fractures. Similar to the BBW, results showed a decrease in bedding plane fracture frequency with depth resulting in the groundwater flow system at depth to behave differently than the shallow aquifer. Morin et al. (1997) presented similar results for the Passaic Formation, New Jersey, USA, where

bedding plane fractures were more transmissive than high-angle fractures, but were of diminishing importance with depth.

A combination of the borehole fracture characteristics and hydraulic test results were used to separate the aquifer into three hydrogeological layers, which were incorporated into a numerical model to refine our estimates of anisotropy and hydraulic conductivity for the fractured aquifer. Overall, the calibrated hydraulic conductivities from the numerical model are consistent with those calculated using the cubic law for individual fracture sets, as well as hydraulic conductivities determined from packer test intervals, with average discrepancies of less than one order of magnitude. The main differences between hydraulic conductivities obtained from constant head tests and calibration of the numerical model were noticed for x-direction in the second layer of the model (i.e. two orders of magnitude; 5.2×10^{-6} m/s initial value vs. 5.3×10^{-8} m/s calibrated value). Sensitivity analysis indicates that the K_x of layer 1, in conjunction with K_x for layer 2 dominates the relationship between measured and observed heads. The difference in hydraulic conductivities between the two layers, leads to a “boundary” effect (Fig. 3.13) at the bottom of L1, with groundwater paths bending to run sub-parallel to the boundary. This effect was also noticed at the boundary between L2 and L3; however, it is less significant due to the smaller difference in hydraulic conductivities. During the calibration, K_z for each of the layers was modified independently of the changes in K_x and K_y . However, the model did not show significant changes in mass balance or flowpaths (e.g. less than 1% NRMSE change) for over four orders of magnitude adjustment of K_z (5×10^{-4} m/s to 5×10^{-8} m/s). Physically, this may be attributed to the fact that horizontal conductivities in the lower two layers were two to three orders of magnitude smaller than those in the upper layer, thereby limiting flow into the lower layers irrespective

of the values assigned to K_z . The relatively low sensitivity to variations in K_z is also consistent with findings of other studies conducted in fractured bedrock aquifers (e.g. Rivard et al. 2008) suggesting that the near surface flow in these systems is controlled by the highly conductive bedding plane fractures, while the high angle fractures might play a more important role in the deeper aquifer (e.g. Merin 1992; Gallagher 1997; Morin et al. 1997).

Results from the 3D steady-state numerical model sufficiently reproduce the observed hydraulic head values in the boreholes. However, as shown in Fig. 3.13, the modelled heads tend to be lower than the measured water levels in the upslope areas and higher in the downslope areas. This could be attributed to the insufficient resolution of the modelled grid and lateral variability in hydrogeological properties of the model, as well as the dip of bedding plane fractures not considered in the model. Results of the numerical model indicate that flow is predominantly in the NE-SW direction, corresponding to the strike of the bedding plane fractures and slope of local topography. Although several high-angle fractures are present within the aquifer, vertical fluxes observed in the model between the shallow aquifer layer (L1) and deeper layers of the model remain insignificant. Model calibration, resulting in an increase in K_z for layer L1 compared to the deeper layers of the model, provided only minor increases to vertical groundwater flow path through the aquifer. Consequently, model results suggest that groundwater flow is predominantly controlled by the horizontal components (K_x and K_y) within in the top layer (L1) of the model. Such conditions indicate that the shallow aquifer may be vulnerable to contamination being transported over large distances, with relatively minor vertical movement between the different model layers. These findings have implications for

developing agricultural best management practices and wellhead protection plans for the Black Brook Watershed and surrounding region.

3.8 Conclusions

A conceptual model for the fracture network controlling groundwater flow in the Black Brook Watershed (BBW) has been developed based on a combination of geophysical logging, outcrop mapping, and constant head hydraulic testing. Borehole logging and hydraulic packer tests indicate groundwater flow is dominantly controlled by the bedding plane fractures with groundwater flow paths parallel to the regional northeasterly structural trend. The high-angle fracture sets tend not to be related to fold development in this area, but likely propagated parallel to the principal stress directions during time of formation. The F3 fracture set presumably formed most recently compared to the other fractures as neotectonic joints aligned parallel to the current direction of maximum compressive stress as inferred from the orientation of borehole breakouts. Both the F1 and F2 fractures are suggested to be related to earlier paleostress fields, likely associated with the Acadian and Alleghenian compression. The through-going nature of the high-angle fracture sets cross-cut multiple bedding plane fractures, acting to enhance the vertical connectivity between highly transmissive bedding plane fractures. Fractures showing large apertures, significant weathering and oxidation are deemed the most important to the overall groundwater flow system, most of which are bedding plane fractures (S1), and to a lesser degree F3 high-angle fractures. These fractures trend NNE to NE, parallel to the maximum principal stress direction, which results in these fractures being held open under an extensional condition. The hydraulic test results show a general decrease in hydraulic conductivity with depth, which corresponds with a decrease in the frequency of fractures exhibiting oxidation and

weathering. Groundwater flow modelling suggests that most flow occurs within the shallow aquifer, mainly controlled by the horizontal hydraulic conductivity, with preferential flow occurring in the NE-SW direction following the strike of the bedding plane and F3 fractures, and the slope of local topography.

The genetic relationship observed between the regional fracture trends, large-scale fold geometry, and the contemporaneous in situ stress field within the region provides a valuable approach to predict azimuthal anisotropy in hydraulic conductivity in the remaining portions of the BBW, as well as adjacent watersheds within the Matapédia Basin. Such insight is important within this agricultural region to help minimize the potential risk of contaminating the shallow groundwater aquifer by agricultural chemicals, and to develop watershed best management practices. Careful extrapolation of this approach can be helpful in locations where boreholes and outcrop exposures are limited, whereby a simpler understanding of the regional structural trends may allow for the dominant fracture pathways to be predicted in similar structural settings throughout the Matapédia Basin which extends from northern New Brunswick and into neighboring Maine and Québec.

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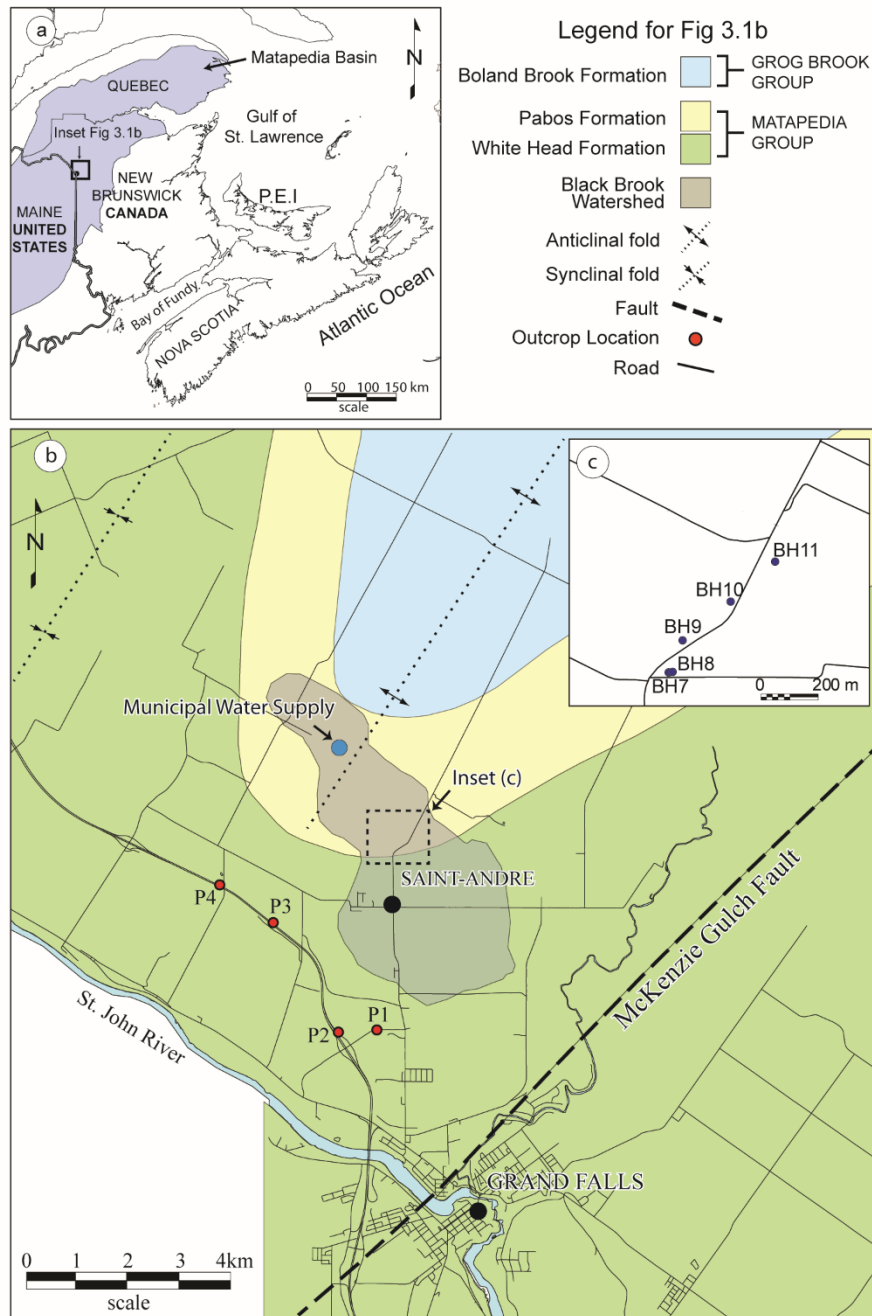


Figure 3.1 a) Location of the Black Brook Watershed study area and the extent of the Matapédia Basin through New Brunswick, Québec and Maine; and b) bedrock geology in the vicinity of the BBW near Grand Falls, New Brunswick. The dashed box outlines the inset figure, which shows the location of tested boreholes in the study area within the Pabos Formation. A series of anticlinal and synclinal structures trend in a northeast direction roughly parallel to the McKenzie-Gulch Fault. The BBW boundary encompasses the Pabos and White Head Formations, and the Community of Saint-André

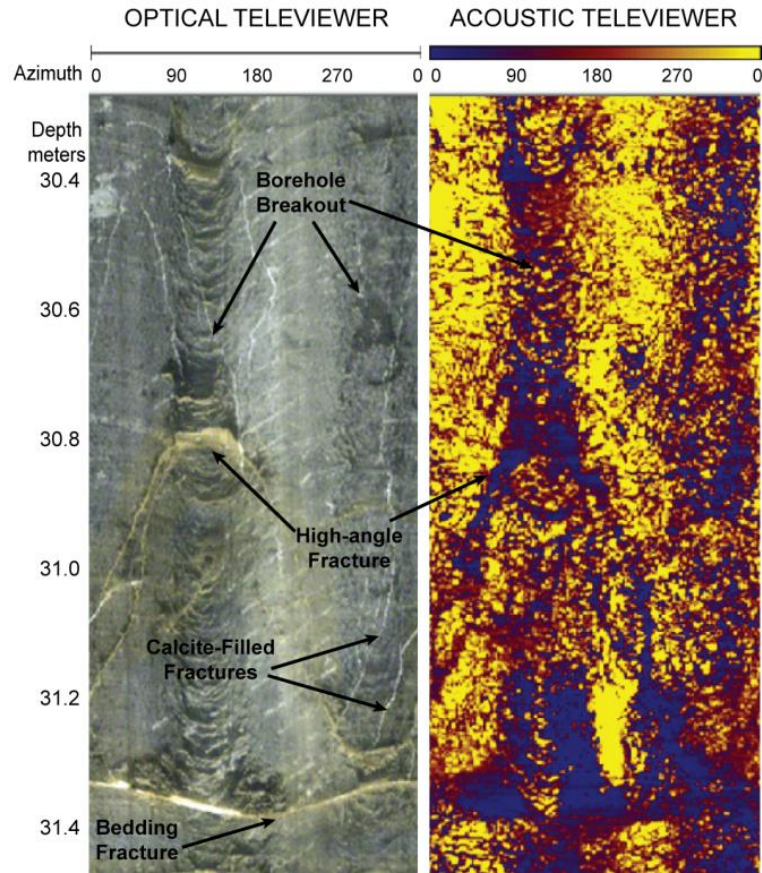


Figure 3.2 Acoustic and optical televiewer images showing presence of vertical borehole breakouts oriented ESE-WNW, thin calcite-filled fractures, and oxidized high-angle and bedding plane fractures intersecting a 1 m interval of BH10

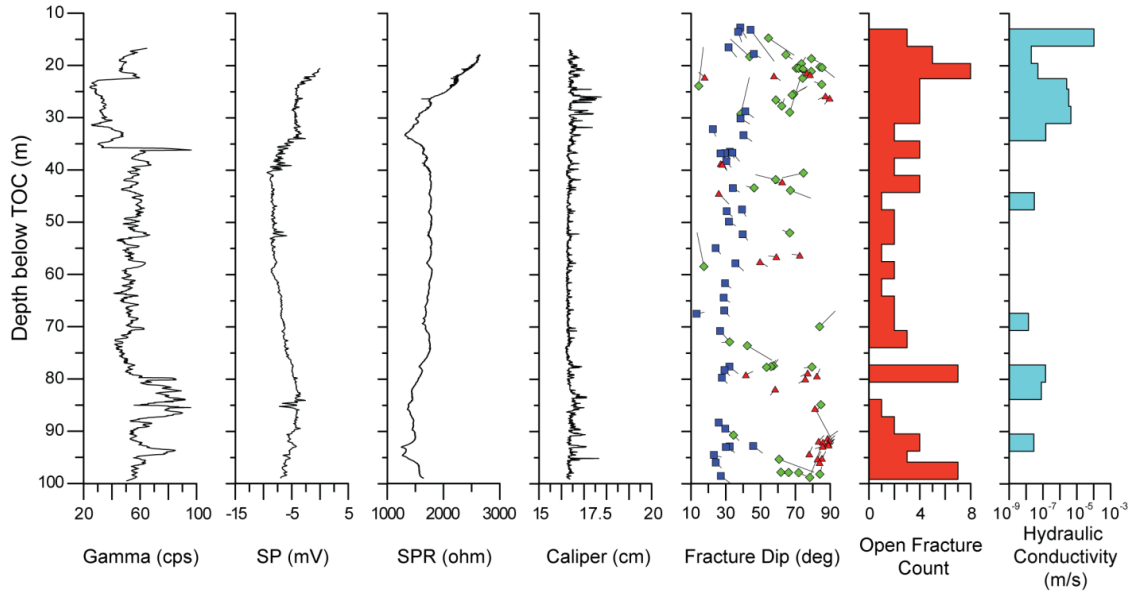


Figure 3.3 Composite image of geophysical logs and hydraulic packer test profile from BH7. Fracture orientations are represented by tadpole plots (blue – bedding plane fractures; green – high-angle open fractures; red – high-angle calcite-filled fractures) where the tail points in the dip direction and the aperture magnitude is represented by the tail length. The interval length for the open fracture count and hydraulic conductivity log is 3.3 m. TOC represents top of casing.

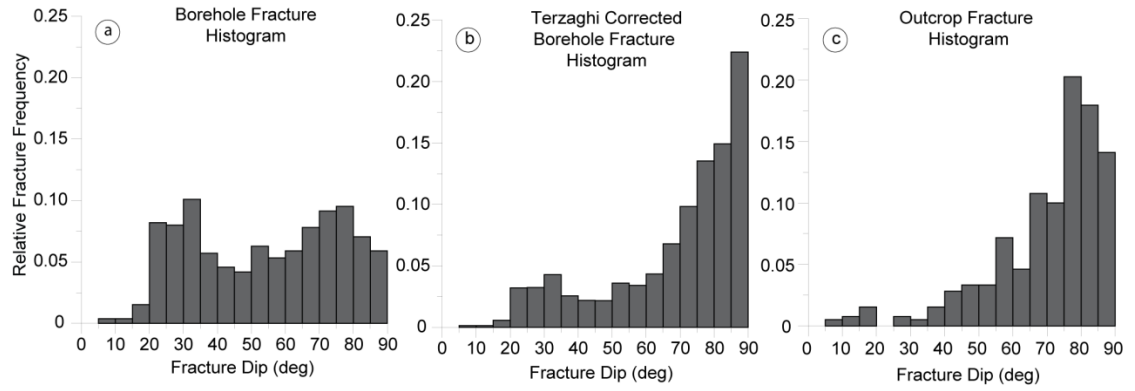


Figure 3.4 Histogram of dip magnitude from 525 fractures measured in 5 vertical boreholes before (a) and after (b) the dip angle bias correction, and (c) 575 fractures from outcrop mapping

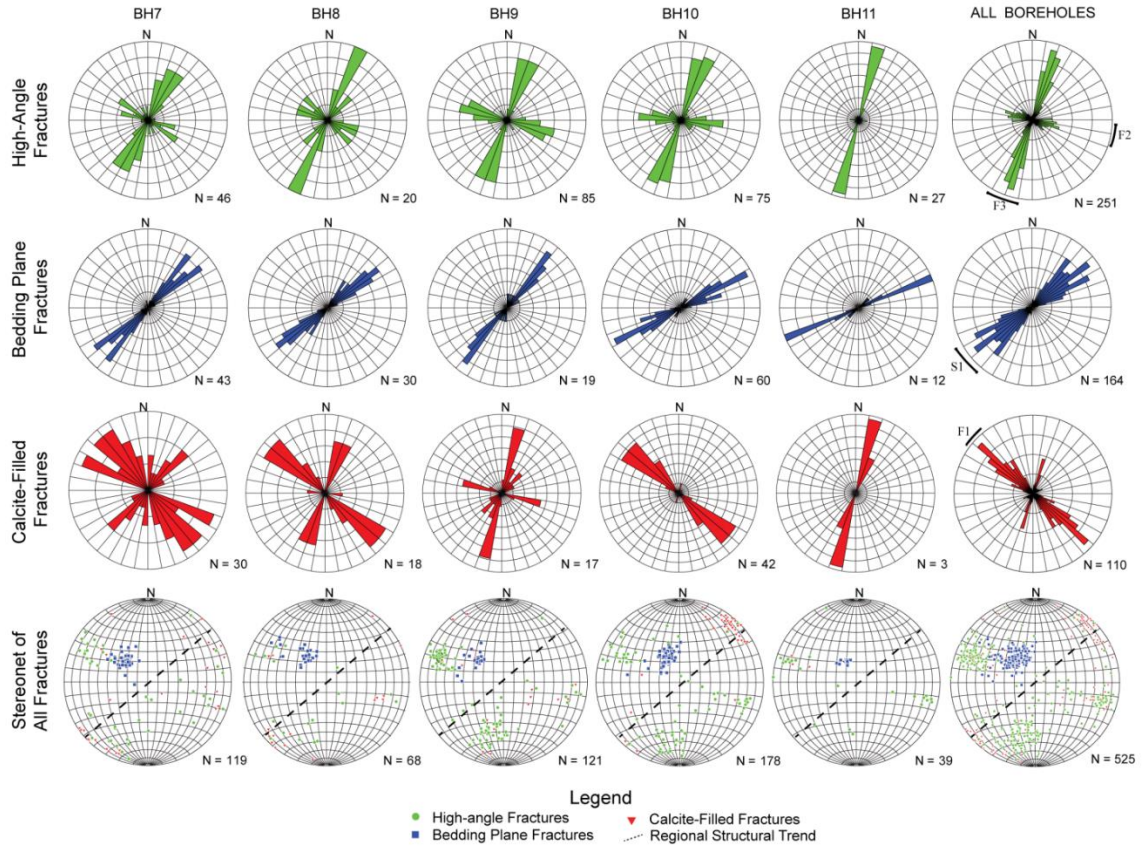


Figure 3.5 Rose diagrams and equal-area stereonet projections showing orientations for high angle fractures, bedding plane fractures and calcite-filled fractures for individual boreholes, and combined for all boreholes. High-angle fractures show two distinct sets (F2 and F3). The F3 set shows a uniform trend towards the NNE, whereas F2 set shows azimuthal variability between different boreholes as a counter-clockwise rotation. Bedding plane fractures show variability between boreholes as a clockwise rotation. Calcite-filled fractures show significant variability between wells, with a dominant trend towards the NW, and a minor trend NNE. The data are weighted using the Terzaghi correction to adjust the frequency of high-angle fractures intersecting a vertical well.

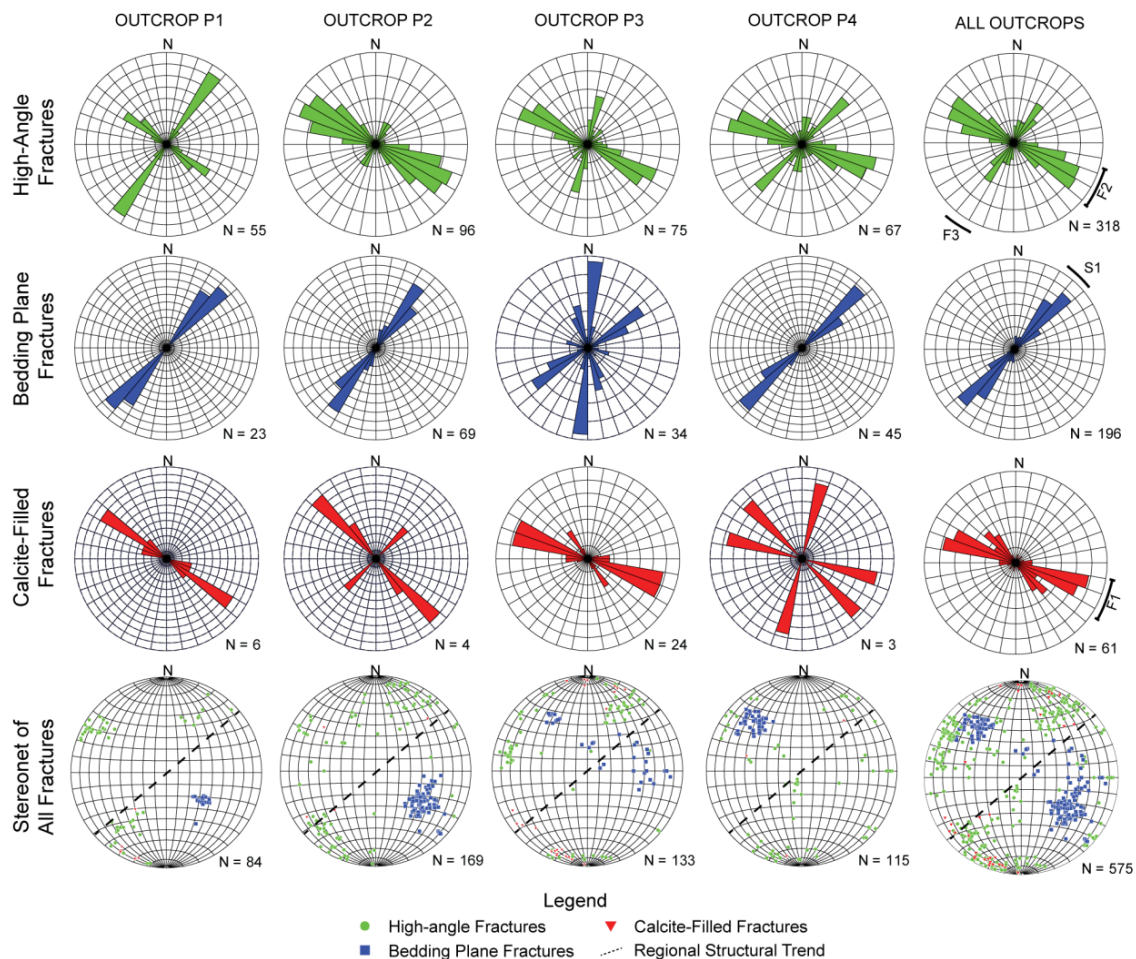


Figure 3.6 Rose diagrams and equal-area stereonet projections showing orientations for high angle fractures, bedding plane fractures and calcite-filled fractures for individual outcrop locations. Locations of outcrops are shown in Fig 1. Survey lines at the outcrops consisted of variable azimuths, therefore removing the requirement to correct the fracture population for azimuthal orientation bias.

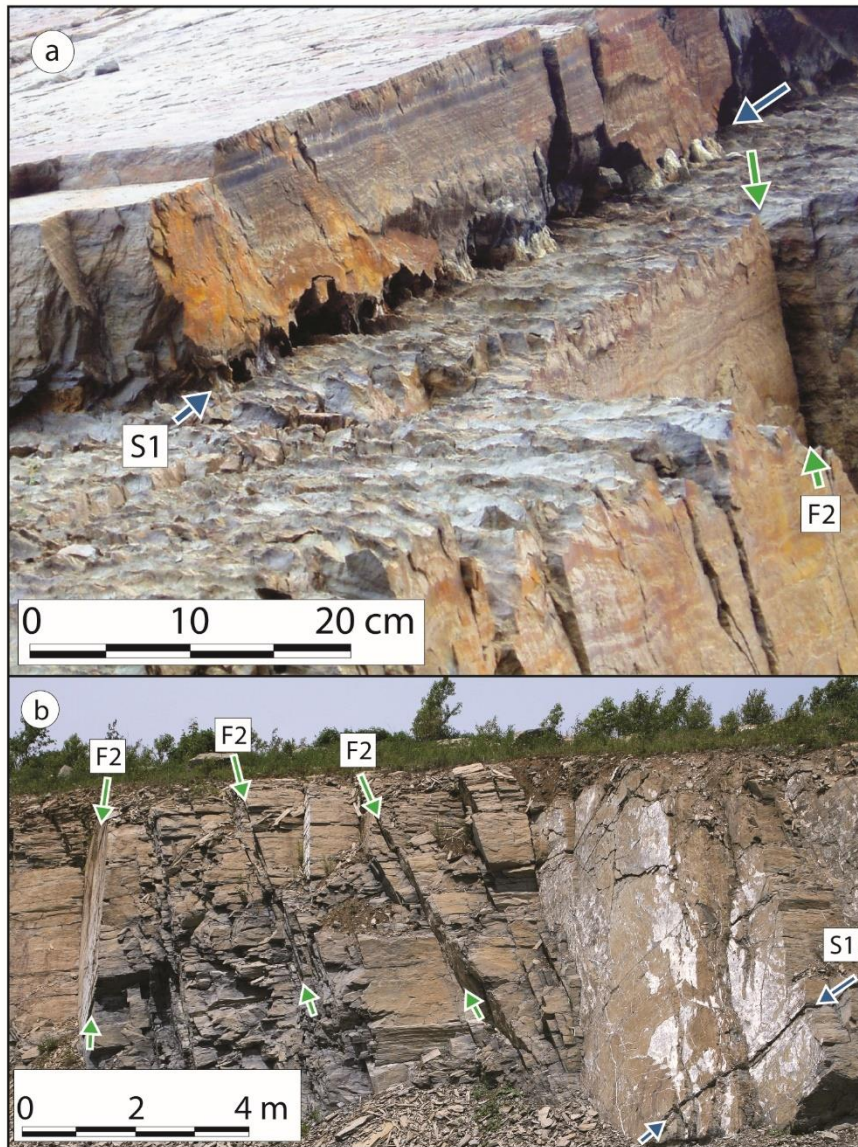


Figure 3.7 Photographs of bedding (S1, blue arrows) and high-angle fractures (F2, green arrows) in outcrop. a) Calcite-filled vein parallel to bedding at outcrop P3 dissolved to form a wide-aperture bedding plane fracture dipping (S1, blue arrow) roughly 45° towards the NW. b) Continuity of bedding plane (S1, blue arrow) and F2 high-angle (F2, green arrows) fractures observed at outcrop P1 (For outcrop locations see Fig. 1.)

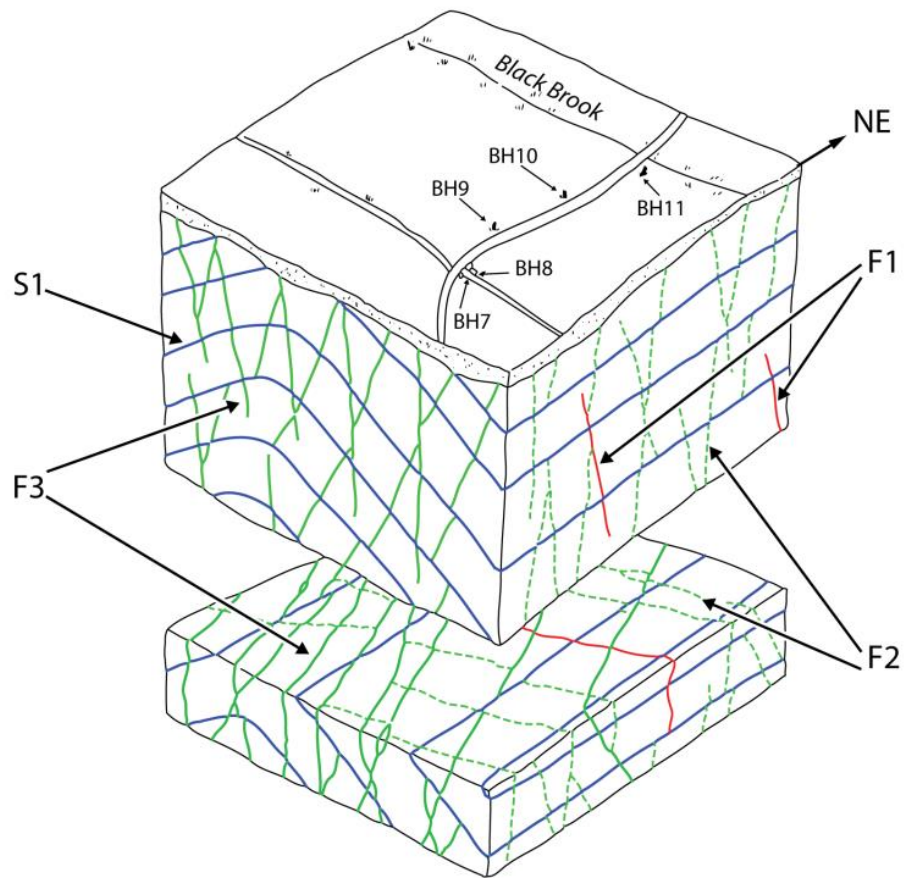


Figure 3.8 Schematic representation of the non-stratabound fracture network through the study area (~1 km²). Units are characterized by 3 - 5 m of glacial overburden overlying the fractured bedrock aquifer. Fracture network consists of F1 calcite-filled fractures, open F2 and F3 high-angle fractures, and a bedding parallel fracture set. F3 fractures trend NNE at an oblique angle to the fold axis, consisting of two subsets dipping ESE and WNW. F2 fractures trend NW roughly normal to the fold axis, consisting of two subsets dipping predominantly NE, with few towards the SW. Mean fracture spacings range between 0.21 and 0.87 m. The groundwater flow system is mostly controlled by the open-aperture and oxide-stained bedding plane fractures, and is enhanced by high-angle fractures providing vertical flow pathways

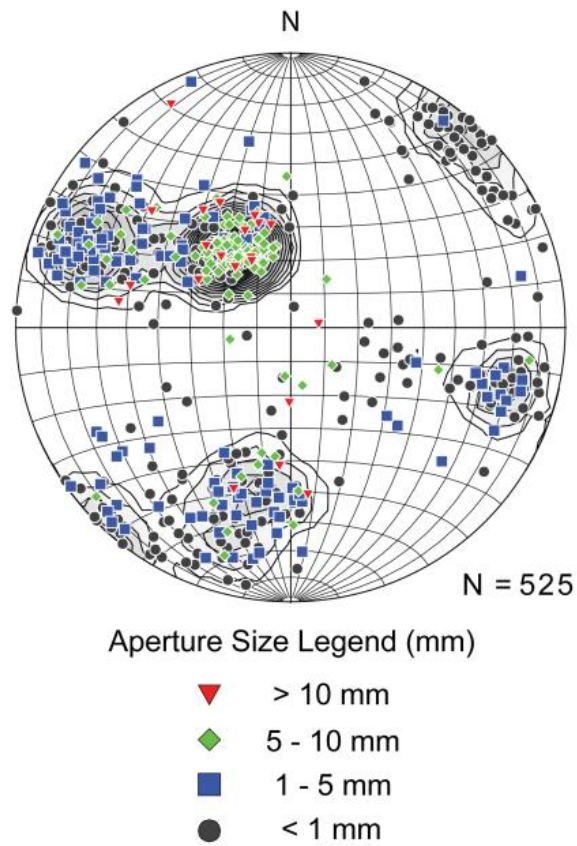


Figure 3.9 Equal-area stereonet diagram showing distribution of fracture apertures from borehole measurements. Bedding plane fractures are dominated by open apertures greater than 5 mm, whereas all other fracture sets tend to comprise aperture widths less than 5 mm

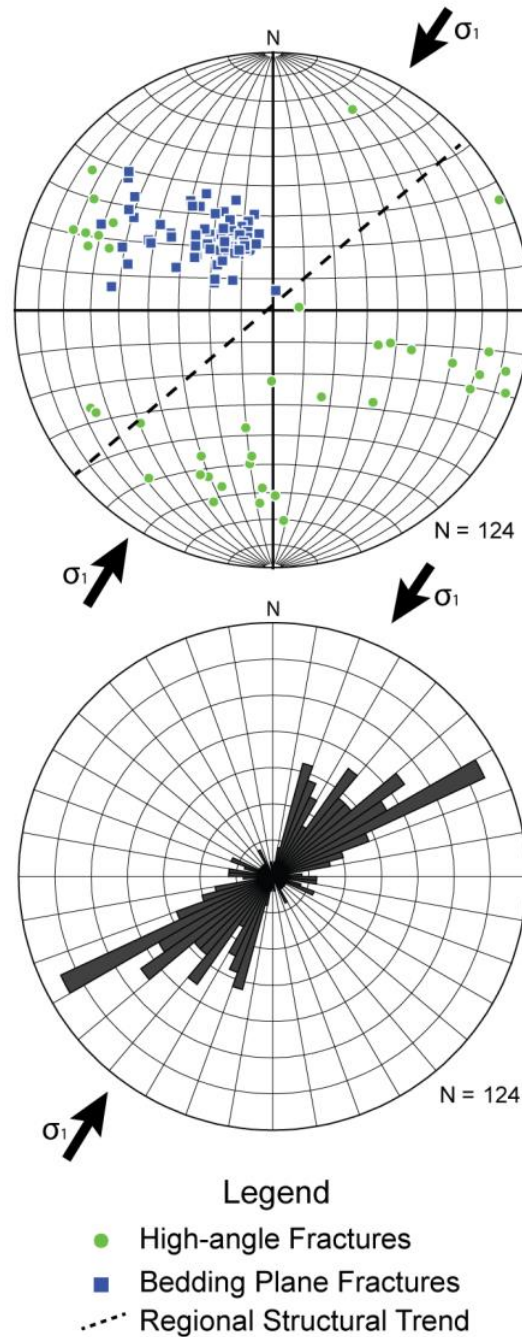


Figure 3.10 Equal-area stereonet and rose diagrams of the highly weathered and oxide-stained high-angle fractures (circles) and bedding plane fractures (squares) identified by inspection of borehole televiewer images. These fractures are aligned sub-parallel to the maximum horizontal direction (σ_1) of the contemporary stress field, as inferred by borehole breakout directions. The horizontal compression directions are interpreted to hold the oxide-stained fractures open under extensional conditions, resulting in a larger contribution to groundwater flow.

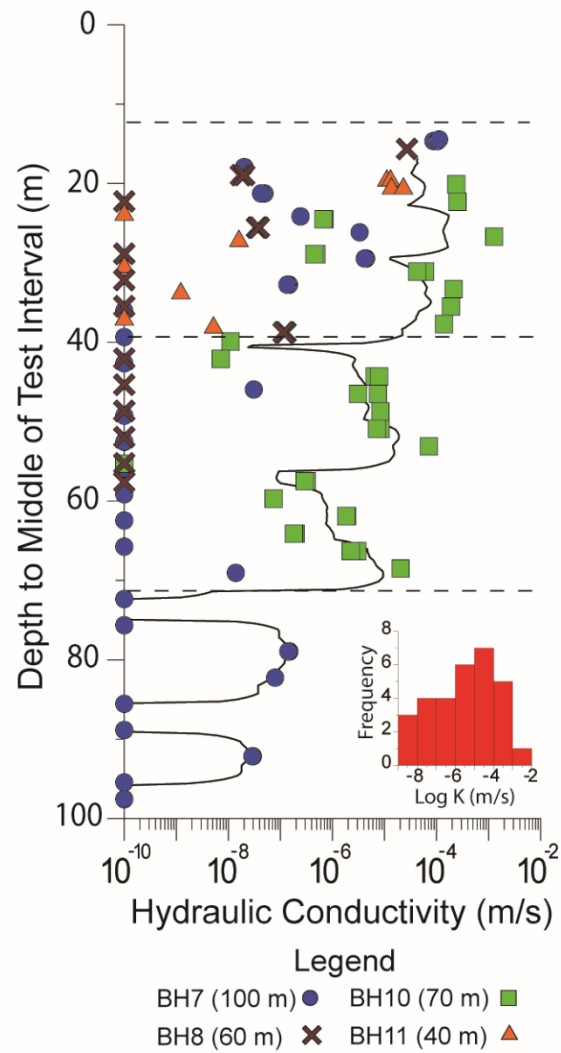


Figure 3.11 Hydraulic conductivities as a function of depth determined by constant head tests in four wells. Well depths are listed in the legend. Black line represents the mean conductivity profile using a 5 m moving window, and can be subdivided into three hydraulic intervals. Note that there are only three measurements below 70 m depth. The combined hydraulic conductivity measurements fit to a slightly skewed log-normal distribution.

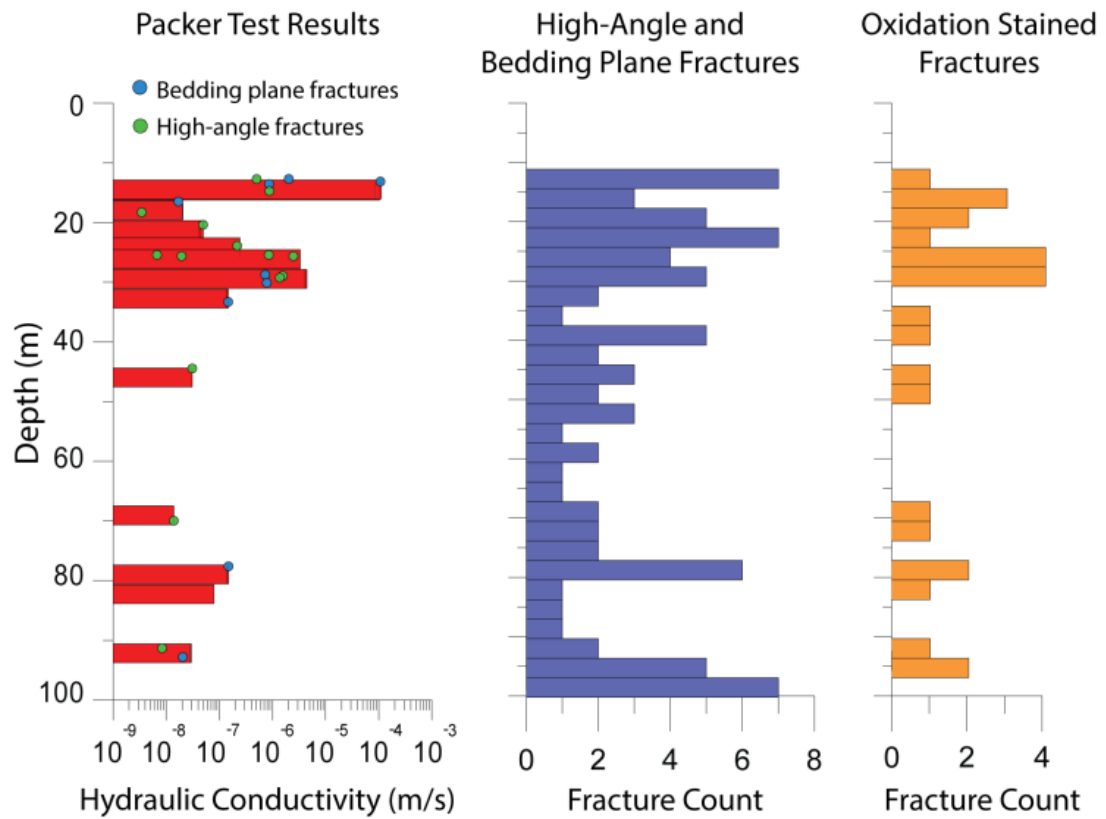


Figure 3.12 Relationship between hydraulic test results, the number of open fractures, and the number of fractures showing evidence of oxidation and weathering in each of the 3.3 m intervals tested in BH7. Blue and green dots superimposed on the hydraulic conductivity measurements indicate the contributions that individual bedding plane and high angle fractures are estimated to make based on their relative apertures.

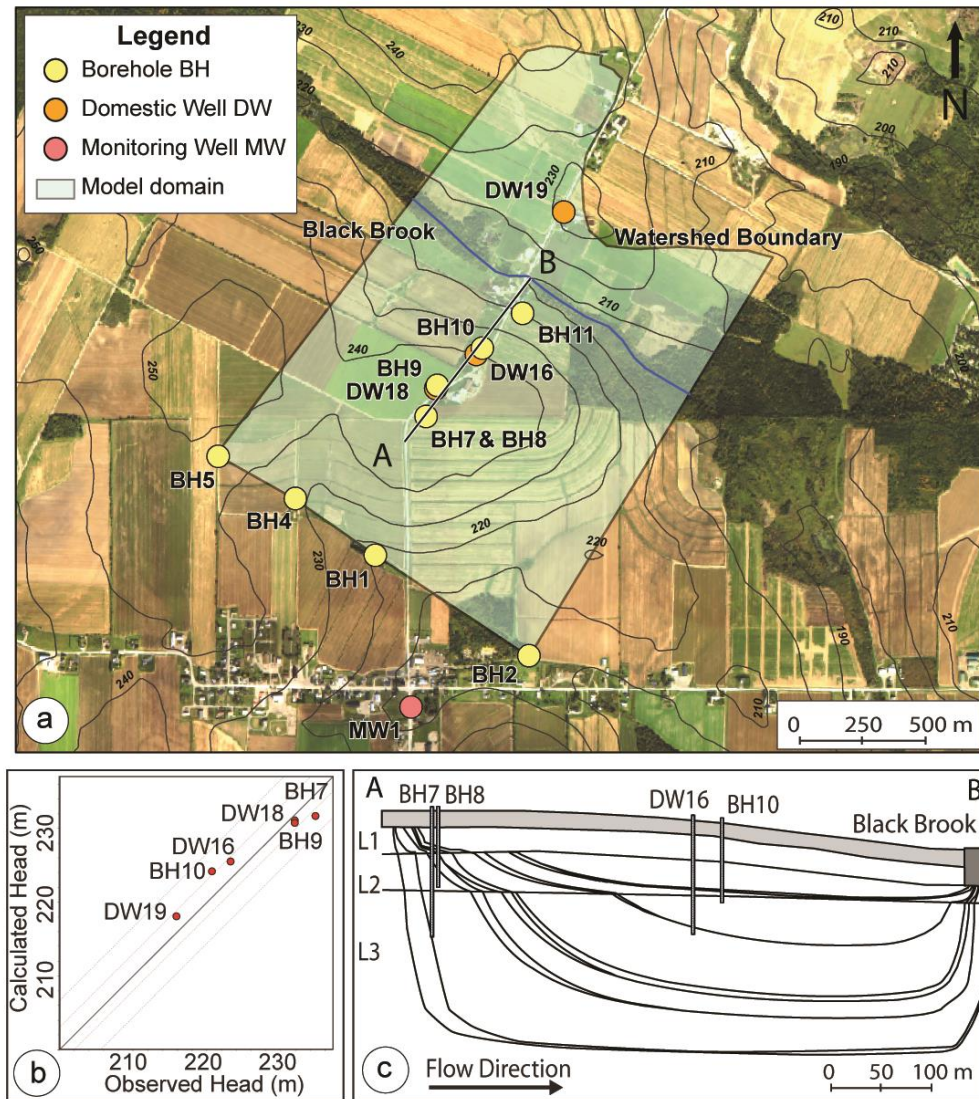


Figure 3.13 Plan view (top) of the anisotropic modelling domain including topography and well locations. The southern boundary is defined by a series of wells and the northern boundary is the watershed limit. The municipal well drilled in 2003 is located roughly 2 km northwest of the figure extent. Lower panels show modeled vs observed heads, and predicted flow paths from the topographic high to Black Brook in section AB which is oriented parallel to the inferred direction of maximum hydraulic conductivity.

Chapter 4

Monitoring and modelling of pumping-induced self potentials for transmissivity estimation within a heterogeneous confined aquifer³

4.1 Summary

Variations in self-potentials (SP) measured at surface during pumping of a heterogeneous confined fractured rock aquifer have been monitored and modeled in order to investigate capabilities and limitations of SP methods in estimating aquifer hydraulic properties. SP variations were recorded around a pumping well using an irregular grid of 31 non-polarizing Pb-PbCl₂ electrodes that were referenced to a remote electrode and connected to a commercial multiplexer and digitizer/data logger through a passive lowpass filter on each channel. The lowpass filter reduced noise by a factor of 10 compared to levels obtained using the data logger's integration-based sampling method for powerline noise suppression alone. SP signals showed a linear relationship with water levels observed in the pumping and monitoring wells over the pumping period, with an apparent electrokinetic coupling coefficient of $-3.4 \text{ mV} \cdot \text{m}^{-1}$. Following recent developments in SP methodology, variability of the SP response between different electrodes is taken as a proxy

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for lateral variations in hydraulic head within the aquifer and used to infer lateral variations in the aquifer's apparent transmissivity. In order to demonstrate the viability of this approach, SP is modeled numerically to determine its sensitivity to (i) lateral variations in the hydraulic conductivity of the confined aquifer and (ii) the electrical conductivity of the confining layer and conductive well casing. In all cases, SP simulated on the surface still varies linearly with hydraulic head modeled at the base on the confining layer although the apparent coupling coefficient changes to varying degrees. Using the linear relationship observed in the field, drawdown curves were inferred for each electrode location using SP variations observed over the duration of the pumping period. Transmissivity estimates, obtained by fitting the Theis model to inferred drawdown curves at all 31 electrodes, fell within a narrow range of $(2.0 - 4.2) \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$ and were consistent with values measured in the pumping and monitoring wells. This approach will be of particular interest where monitoring wells are lacking for direct measurement, and SP on the surface can be used to quickly estimate hydraulic properties.

4.2 Introduction

The evaluation of aquifer transmissivity and storativity typically relies heavily on an adequate network of wells to measure drawdown of the water level or piezometric surface during hydraulic pump tests (Kruseman and de Ridder 2000). However, networks of monitoring wells are expensive, and are typically not available for smaller characterization programs. Measurement of self-potentials (SP) generated by drawdown has recently been proposed as a non-invasive alternative, using a dense network of electrodes to infill between wells (e.g. Revil and Jardani 2013). A number of investigations have demonstrated that SP signals on the ground in the proximity of a pumping well can be used to infer groundwater

flow pathways or to estimate hydraulic properties of the underlying aquifer (e.g. Bogoslovsky and Ogilvy 1973; Suski et al. 2004; Rizzo et al. 2004; Maineult et al. 2008; Malama et al. 2009).

Sill (1983) presented a theoretical basis for rigorous forward modelling of SP in terms of convection current induced by a primary hydraulic gradient, resulting in the generation of conduction currents. The hydraulic problem is first solved and used to determine the distribution of current sources in the subsurface. These sources are incorporated with the subsurface electrical conductivity structure in order to determine the resulting SP. Using this approach, several studies have modelled SP signals generated around pumping wells in both homogeneous and heterogeneous isotropic aquifer scenarios (e.g. Sill 1983; Wurmstich and Morgan 1994; Titov et al. 2005; Sheffer and Oldenburg 2007). Wurmstich and Morgan (1994) modeled a deep oil-saturated reservoir bounded by units with lower permeability and higher electrical conductivity in order to predict SP signal on the surface during pumping. They show that the electrical conductivity assigned to the confining units significantly influence the magnitude of SP on surface. Sheffer and Oldenburg (2007) modeled SP in a heterogeneous aquifer while pumping. Their model assumed a homogeneous distribution of electrical properties, where only hydraulic conductivity changed between units. Titov et al. (2005) evaluated the relationship between SP measured at surface and hydraulic head measured by Rizzo et al. (2004) in an underlying confined aquifer undergoing pumping, where over the course of pumping the hydraulic head level in the aquifer changed from confined to unconfined. They demonstrated that despite the aquifer being heterogeneous in hydraulic conductivity, modelling predicted a laterally consistent linear relationship between drawdown and surface SP that was proportional to the electrokinetic coupling coefficient.

The rigorous forward modelling approach may not always be feasible for the interpretation of SP data in terms of hydraulic properties, as it requires prior knowledge of the spatial distribution of electrical properties of the aquifer (electrical resistivity and electrokinetic current coupling coefficient). However, a semi-empirical approach proposed by Revil et al. (2003), based on prior work by Fournier (1989), provides a simplified form of SP modelling that can be used to estimate spatial variability in drawdown during pumping, and from there, lateral variations in hydraulic properties of an aquifer. Revil et al. (2003) showed that, for the simple case of steady state pumping in a homogeneous, unconfined aquifer, lateral variations in SP should be proportional to drawdown of the water table below the SP electrode. Applying the semi-empirical approach with an electrokinetic coupling coefficient $-3.2 \text{ mV} \cdot \text{m}^{-1}$ they analyzed the SP response measured by Bogoslovsky and Ogilvy (1973) during pumping of an unconfined aquifer and obtained estimates for the depth and shape of the water table in reasonable agreement with the reported piezometer measurements. Darnet et al. (2003) performed an SP inversion of the same dataset to deduce the bedrock depth, the hydraulic conductivity, and the electrokinetic coupling coefficient in a homogenous unconfined aquifer, and concluded that hydraulic conductivity was the best determined parameter. By incorporating resistivity structure into their inversion, they showed clearly that the SP anomaly on surface was influenced by channeling of conduction currents to surface along the conductive well casing. Rizzo et al. (2004) recognized a linear relationship (approximately -0.8 mV/m) between the SP signal and hydraulic head around a pumping well in a confined silty sand aquifer during the recovery phase following a pumping test. SP data recorded using an array of electrodes on surface above the aquifer were converted to a hydraulic head distribution and combined

with direct measurement of water levels in a small network of piezometer wells. By assuming a homogeneous and unconfined aquifer under steady-state, they interpreted the SP data to determine an average hydraulic conductivity (Rizzo et al. 2004). Straface et al. (2007) integrated the hydraulic head estimates that were calculated by Rizzo et al. (2004) with direct observation of head values in wells using a geostatistical gridding method (Yeh et al. 1996) to solve for the spatial distribution of effective transmissivity throughout the aquifer. The transient (pump test recovery) SP data collected by Rizzo et al. (2004) were also used by Malama et al. (2009) who developed a semi-analytical solution for the SP response of a three-layered confined aquifer system consisting of a homogeneous aquifer with upper and lower confining units. Malama et al. (2009) and Suski et al. (2006) both showed that where monitoring wells are unavailable, transient SP data collected on the surface over the duration of a pump test may be used to determine aquifer hydraulic conductivity and specific storage.

This study presents transient SP data collected on surface over a confined and heterogeneous fractured bedrock aquifer during a pump test. Clear temporal SP anomalies with magnitudes of 2 - 9 mV were monitored at 31 different electrode locations using a simple 0.8 Hz lowpass filter on each channel to improve the noise rejection characteristics of a commercial data logger. The SP anomalies generated by pumping are used in combination with a measured electrokinetic voltage coupling coefficient to estimate spatial variations in drawdown around the well. Apparent transmissivity and storativity values are then obtained by fitting the Theis model to the SP-derived drawdown data. The range of apparent transmissivity results compare favorably with values derived from monitoring-well drawdown during long-period pump tests (DesRoches et al. 2012). Finally, rigorous forward

modelling of the relationship between SP and hydraulic head for a confined aquifer is used to illustrate conditions under which the simplified semi-empirical approach can be expected to yield reliable results.

4.3 Self-potential theory

Self-potential (SP) associated with subsurface water flow is an electrical potential field that is a consequence of coupling between flows of water and electrical charge in porous or fractured media. The coupling arises due to drag of excess electrical charge contained in the mobile portion of the electrical double layer at the rock-water interface (e.g., Davis et al. 1978; Ishido and Mizutani 1981; Revil et al. 1999; Jouniaux and Ishido 2012). For the pH and geochemical conditions present in most aquifers, the solid surfaces hold an excess negative charge, which is balanced by excess positive charge in the pore fluid that is concentrated adjacent to the pore wall. Some of the positively charged ions in this electrical double layer are free to move with the pore fluid resulting in the establishment of convection currents which can in turn lead to the development of a potential gradient driving conduction currents in the opposite direction (Revil and Leroy 2001). Self-potential gradients may also arise from heat fluxes, geochemical diffusion and redox reactions within the aquifer system (e.g., Sill 1983; Rubin and Hubbard 2005; Maineult et al. 2006; Jouniaux et al. 2009; Hubbard and Linde 2011).

Under isotropic conditions the electric and fluid flows are described by transport equations operating at the scale of a representative elementary volume (e.g. Titov et al. 2005; Straface et al. 2010):

$$\mathbf{j} = -\sigma \nabla \varphi - L \rho g \nabla h \quad (4.1)$$

$$\mathbf{u} = -L\nabla\varphi - \frac{k}{\mu}\rho g\nabla h \quad (4.2)$$

where $\mathbf{u}$ is the volumetric fluid flux ($\text{m}\cdot\text{s}^{-1}$), and $\mathbf{j}$ is the total electrical current density ($\text{A}\cdot\text{m}^{-2}$) represented by the sum of a conduction current term given by Ohm's Law $\mathbf{j}_c = -\sigma\nabla\varphi$ and a convection or source current term $\mathbf{j}_s = -L\rho g\nabla h$ representing the electrokinetic contribution. Here σ is the bulk electrical conductivity ($\text{S}\cdot\text{m}^{-1}$) of the rock, φ is the electric potential (V), k is the intrinsic permeability (m^2), μ is the fluid dynamic viscosity ($\text{Pa}\cdot\text{m}$), ρ is the fluid density ($\text{kg}\cdot\text{m}^{-3}$), g is gravitational acceleration ($\text{m}\cdot\text{s}^{-2}$), and L is the electrokinetic current coupling coefficient ($\text{A}\cdot\text{m}^{-1}\text{Pa}^{-1}$). Hydraulic head (m) is given as $h = \psi + z$, where $\psi = p/\rho g$ is the pressure head (m), p is the pore fluid pressure (Pa), and z is the elevation head (m) (Domenico and Schwartz 1997).

Equation (4.2) can be simplified under the assumption that the only source of electric potential gradient is that generated by electrokinetic coupling, and that the resulting electro-osmotic contribution to fluid flow $-L\nabla\varphi$ is several orders of magnitude smaller than that produced by the hydraulic gradient associated with pumping (Sill 1983; Ishido 1989), resulting in the classical Darcy's equation:

$$\mathbf{u} = -\frac{k}{\mu}\rho g\nabla h = -K\nabla h \quad (4.3)$$

where $K = \frac{k}{\mu}\rho g$ is hydraulic conductivity ($\text{m}\cdot\text{s}^{-1}$).

The following governing equation can be solved to determine the time-dependent spatial distribution of hydraulic head h driving groundwater flow as a consequence of pumping:

$$S_s \frac{\partial h}{\partial t} + \nabla \cdot \mathbf{u} = Q \quad (4.4)$$

where t (s) represents time, S_s is the specific storage (m^{-1}), and Q is the pumping rate (m^3s^{-1}).

Equations (4.1) and (4.2) are linked through the electrokinetic current coupling coefficient (Revil et al. 2003; Darnet et al. 2003; Straface et al. 2010):

$$L = -\frac{\sigma C_h}{\rho g} \quad (4.5)$$

where C_h is the electrokinetic voltage coupling coefficient (V m^{-1}) representing the voltage that would develop across a sample of the material in response to fluid flow driven by a unit change in hydraulic head across the sample. The electrokinetic coupling coefficient can vary as a function of water saturation (Perrier and Morat 2000; Allègre et al. 2012; 2015; and Giampaolo et al. 2014).

By substituting eq.(4.5) into eq.(4.1), the total current density can be presented as:

$$\mathbf{j} = \mathbf{j}_c + \mathbf{j}_s = -\sigma \nabla \varphi + \sigma C_h \nabla h \quad (4.6)$$

Following Sill (1983), we now appeal to the continuity equation for steady currents $\nabla \cdot \mathbf{j} = 0$ or $\nabla \cdot \mathbf{j}_c = -\nabla \cdot \mathbf{j}_s$, to identify the sources (non-zero divergence) of conduction currents arising from electrokinetic coupling:

$$\nabla \cdot \mathbf{j}_c = \nabla \cdot (\sigma C_h \nabla h) = \nabla(\sigma C_h) \cdot \nabla h + \sigma C_h \nabla^2 h. \quad (4.7)$$

Taking the divergence of eq.(4.3) to obtain an expression for $\nabla^2 h$ and expanding the two terms on the right hand side of eq.(4.7) reveals four distinct sources:

$$\begin{aligned} \nabla \cdot \mathbf{j}_c = \sigma(\nabla C_h \cdot \nabla h) + C_h(\nabla \sigma \cdot \nabla h) - \left(\frac{\sigma C_h}{K}\right)(\nabla K \cdot \nabla h) - \\ \left(\frac{\sigma C_h}{K}\right)(\nabla \cdot \mathbf{u}). \end{aligned} \quad (4.8)$$

Interpretation of the three source terms dependent on the hydraulic gradient ∇h is more physically meaningful if we use Darcy's Law (eq. (4.3)) to reformulate them in terms of the fluid flux $\mathbf{u}$:

$$\begin{aligned} \nabla \cdot \mathbf{j}_c = & -\left(\frac{\sigma}{K}\right) (\nabla C_h \cdot \mathbf{u}) - \left(\frac{C_h}{K}\right) (\nabla \sigma \cdot \mathbf{u}) + \left(\frac{\sigma C_h}{K^2}\right) (\nabla K \cdot \mathbf{u}) - \\ & \left(\frac{\sigma C_h}{K}\right) (\nabla \cdot \mathbf{u}). \end{aligned} \quad (4.9)$$

The first three terms on the right hand side of eq. (4.9) show that SP sources can be generated by flow (hydraulic gradients) directed perpendicular to boundaries (or parallel to gradients) in electrokinetic voltage coupling coefficient C_h , electrical conductivity σ , or hydraulic conductivity K . The fourth term, arising from any divergence in fluid flow, corresponds to electrical sources which occur at fluid sources or sinks (i.e. injection or pumping wells).

Following the methodology proposed by Sill (1983), a rigorous forward modelling approach can be used to solve for the SP distribution. The first step is to determine the distribution of hydraulic head through the model domain using the governing differential eq. (4.4), using specified hydraulic conductivities, and appropriate boundary conditions. Given the hydraulic head distribution, the sources of conduction currents $\mathbf{j}_c = -\sigma \nabla \varphi$ can be solved using eq.(4.8), provided the distributions of the electrokinetic coupling coefficient (C_h), and the electrical conductivity of the bedrock (σ) are known. Lastly, the sources of conduction currents generated in response to fluid flow can be used to solve for SP distribution (φ) through the model. If the distribution of electrical conductivity can be

determined by 3D resistivity imaging, and electrokinetic coupling coefficients reasonably estimated, it may be possible to invert SP data rigorously to solve for the depth and shape of the underlying water table (e.g. Jardani et al. 2009), and ultimately determine the hydraulic properties of the aquifer.

A simplified approach using the SP response generated at surface by water well pumping can be used to infer the associated drawdown of hydraulic head around the well by assuming (i) homogeneous electrical and hydraulic conductivities within the flowing portion of the aquifer, (ii) sediments overlying the water table have uniform thickness and electrical conductivity, and (iii) electrokinetic voltage coupling coefficient is homogeneous. Based on these simplifying assumptions, the semi-empirical approach given by Revil et al. (2003) predicts that changes in hydraulic head associated with a drawdown cone around a pumping well are simply proportional to observed changes in SP at surface according to the equation,

$$\partial\varphi = C_h(\partial h)_{\mathbf{j}=\mathbf{0}} \quad (4.10)$$

where $\partial\varphi$ is a change in the potential difference (V) measured on the ground surface between an observation point located within the zone of drawdown and a reference point far away from the pumping well. The subscript $\mathbf{j} = \mathbf{0}$ indicates that the equation holds provided total current flow is zero, meaning that convection currents are exactly balanced by conduction currents everywhere in the subsurface. Physically, this means that the equation is valid provided that conduction currents follow the same paths (in the opposite direction) as convection currents associated with water flow. The equation may therefore be expected to break down in situations where resistivity heterogeneities cause conduction currents to follow different paths than the water flow; such a situation could arise where

water flow is focused in electrically resistive sandy units that are underlain, overlain or separated by electrically conductive clay-rich units that act as shunts for the conduction currents (cf. Wurmstich and Morgan 1994).

Eq. (4.10) states that under suitable conditions $\partial\phi$ and ∂h are linearly related through the electrokinetic voltage coupling coefficient C_h ($V \cdot m^{-1}$). Rigorous modelling presented by others (e.g. Titov et al. 2005; Jardani et al. 2009) and in this paper indicates that the proportionality remains approximately correct for moderate deviations from the assumptions listed above. The coupling coefficient C_h can be determined by laboratory measurements (e.g. Morgan et al. 1989), or in the field using eq. (4.10) in the case where the change in hydraulic head is observed in a monitoring well and SP is measured directly beside the well. This latter approach results in an apparent electrokinetic coupling coefficient C_{ha} ($V \cdot m^{-1}$) for the aquifer (e.g. Revil et al. 2003) that takes into account the in situ fluid and rock electrical conductivities, thereby at least partially compensating for confining layer shunting effects as discussed above. With such field or lab measurements of C_h or C_{ha} , and measurements of SP at numerous electrodes spread over the surface, eq.(4.10) may be used to approximate the drawdown distribution during the pumping period (e.g. Revil et al. 2003, and Rizzo et al. 2004).

4.4 Description of the experiment

4.4.1 Site Description

Monitoring of SP was completed around a municipal pumping well in the Springdale Wellfield, which is located in the South Brook valley approximately 18 km east of the town of Sussex, New Brunswick, Canada (Figure 4.1). This region has experienced significant

groundwater supply issues in recent years, and is adjacent to both a natural gas field and an underground potash mine, leading to a need to better understand the groundwater flow system (DesRoches et al., 2012).

Local geology within the wellfield comprises fractured sedimentary bedrock formed by two dominant lithology end-members in the Boss Point Formation, within the Moncton Subbasin. These include a highly fractured coarse-grained sandstone unit comprised of channelized deposits that are lenticular in nature (Figure 4.2) surrounded by fine-grained moderately fractured shale to siltstone unit (DesRoches et al. 2012). The geometry of the channelized unit appears to exert major control on groundwater flow directions during pumping. Within the bedrock, fractures comprise three discrete high-angle sets, and one sub-horizontal bedding plane set. The northwest-trending set provides the dominant flow path within the channelized sandstone unit. Other fracture sets are shown to be present in similar abundances, but are only considered to provide connectivity pathways between the fractures trending along the dominant flow direction. Adjacent to the sandstone units, bedrock comprises moderate to sparsely fractured shale units that are thinly laminated. The shale units appear to limit lateral and vertical connectivity in the aquifer (DesRoches et al. 2012). Results from a long-duration pump test provided aquifer transmissivity estimates of $(5.1 - 6.3) \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$ and storativity values on the order of 10^{-5} (DesRoches et al. 2012). Water levels in wells MW07-1 and PW07-1 indicate that the bedrock aquifer is fully saturated, with the water levels representing a potentiometric surface extending into overburden that acts as a confining layer.

Overburden sediments consisting of alluvial sand, gravel, silt and clay blanket the study area to depths of 10 to 12 m. Geological logs from MW07-1 and PW07-1 identified a

clay-rich till located at the base of the overburden deposits. This till layer hydraulically isolates a surficial aquifer within the alluvial sand overburden sediments from the aquifer in the underlying fractured bedrock (DesRoches et al. 2012). This model is supported by two monitoring wells completed in the overburden sediments (~ 10 and 200 m from pumping well) which showed zero drawdown during a long-duration pump test, as well as by a nearby bedrock well (180 m from the pumping well) found to be under artesian pressure with ambient water level slightly above ground surface.

Resistivity logs acquired in monitoring well MW07-1, and in an additional well outside the SP study area (210 m from the pumping well), were used to estimate the vertical resistivity structure of the bedrock aquifer. Resistivity measurements averaged $120\ \Omega\text{m}$ ($0.0083\ \text{S}\cdot\text{m}^{-1}$) in the lower part of the wells (32 to 40 m depth) associated with shaley bedrock, and $180\ \Omega\text{m}$ ($0.0055\ \text{S}\cdot\text{m}^{-1}$) in the depth range (~ 22 m to 32 m depth) corresponding to the channelized sandstone unit. This change in lithology is also expressed in the natural gamma, single point resistance, and optical televiewer logs (DesRoches et al. 2012). Borehole fluid conductivity logs were consistent throughout the boreholes, with a mean value equal to $0.0147\ \text{S}\cdot\text{m}^{-1}$ (resistivity of $68\ \Omega\text{m}$) at a fluid temperature of 6.3°C ($\sim 0.0245\ \text{S}\cdot\text{m}^{-1}$ at 25°C).

4.4.2 Self-potential measurements

SP measurements were recorded around the pumping well (PW07-1) and a nearby monitoring well (MW07-1) during a three day period, from August 10-12, 2010. Both wells measure 43 m deep by 15.2 cm in diameter and have roughly 12 m of steel casing through the overburden sediments. Water levels were continuously recorded in both for the duration of the experiment. Pumping was activated by an automated control system

(SCADA) to fill an above-ground water tank located 200 m away to the east. During this field study, the pump ran for approximately 300 consecutive minutes each day at a rate (Q) of 7.8 L s^{-1} ($0.0078 \text{ m}^3 \text{ s}^{-1}$ or $673 \text{ m}^3 \text{ day}^{-1}$), each time yielding drawdowns of approximately 1.4 m and 1.1 m in the pumping and monitoring wells respectively. At the end of each of pumping period the drawdown approached steady state conditions.

SP measurements from an array of 31 non-polarizing Pb-PbCl₂ electrodes (Petiau 2000) were recorded at a 3 second (0.33 Hz) sample interval using a Campbell Scientific AM32B multiplexer connected to a CR1000 data logger having an input impedance of 20 GΩ. The electrodes were arranged in an irregular grid pattern on the surface around the perimeter of the pumping well (Figure 4.3), avoiding surface obstacles such as tree roots and a steel fence enclosing the pump facility. Typical spacing between adjacent electrodes was 5 to 10 m, at distances of 8.7 to 35.7 m from the pumping well. All electrodes were buried in shallow holes augured to a depth of approximately 20 cm, and seated in bentonite mud in order to promote uniform surface contact, lower contact resistance, and minimize temperature and moisture variations during the experiment.

Electrodes in the array were referenced to a single remote electrode 175 m away from the pumping well where drawdown was expected to be negligible over the duration of the pumping test. It was also removed from areas expected to exhibit elevated electrical noise levels associated with powerlines. Two additional electrodes were deployed 25 m southwest and northeast of the reference electrode in order provide two remote dipoles for monitoring telluric fields and any other ambient electrical noise. Those measurements were analyzed to identify any long period telluric signals that would also affect electrodes within the wellfield array.

SP measurements were acquired by the CR1000 data logger in a mode that integrated voltage over 16.66 ms (one cycle of 60 Hz) in order to cancel stable powerline noise occurring at 60 Hz and its harmonics. To further reduce noise, the leads from each SP dipole were terminated, at the inputs to the multiplexer, with a passive RC filter to attenuate noise at frequencies higher than ~ 1 Hz, including powerline harmonic and telluric fields (discussed in more detail below).

To emphasize the electrokinetic contribution captured during the pumping experiment, this study attempted to remove the influence of spatial variations in SP observed across the array in the absence of pumping. Such spatial variability can be associated with heterogeneity in the subsurface resistivity structure (particularly in the near surface), variations in soil temperature, pH, and moisture content. In addition, oxidation-reduction reactions may generate SP anomalies in the presence of corrosion and generation of biofilms around the steel casing in the wells (Rizzo et al. 2004). None of the above influences tend to vary over the duration of a pump test. Therefore, as argued by Naudet et al. (2003) and Rizzo et al. (2004), their impact on the SP at each electrode in the array can be removed by subtracting temporal reference values which we calculated as the average SP recorded at each electrode over a 1 minute period immediately before the onset of pumping.

4.4.3 Noise reduction

One limitation of the SP method for monitoring groundwater flow is the effect of electrical noise recorded concurrently with the streaming potentials (Corwin 1989). SP data are considered difficult to interpret due to multiple sources of signal and noise (e.g. Revil et al. 2006). Sources of noise include powerline harmonics, telluric currents, long term thermal

changes in soil temperature, and electrode drift (Revil et al. 2012). Allegre et al. (2014) recently showed that variability of water saturation at the electrode can also generate subtle electric signals which may be considered signal or noise depending on the application. Magnitudes may be comparable to, or even much greater than SP arising from electrokinetic coupling. In most monitoring studies, SP data are digitally filtered to deal with high frequency noise using a variety of lowpass filters (e.g. Rizzo et al. 2004), and wavelet based filters (e.g. Moreau et al. 1996). However, noise occurring at frequencies exceeding half the sample rate can be aliased to appear at lower frequencies in the same bandwidth as expected electrokinetic effects. This highlights the importance of removing noise sources in the field before the raw data are digitized.

In this study, the objective of the noise reduction was to minimize high-frequency interference suspected to be dominated by residual powerline harmonic noise not cancelled by the data logger's 16.66 ms integration sampling mode. Powerline harmonic noise can commonly be hundreds of times stronger than the SP signals of interest (Corwin 1989). Residual powerline noise is to be expected if the amplitude or phase of the powerline fields in the earth varies over time (typically in response to changes in load) or if the fundamental frequency deviates from exactly 60 Hz. Previous studies (Butler and Russell 1993; 2003) have documented that powerline field amplitudes in many areas of the world can be on the order of a few mV/m – corresponding to 60 Hz noise levels on the order of 500 mV between SP electrodes and a reference electrode located 175 m away as in this study, and that frequencies can deviate from the nominal fundamental by up to a few hundredths of a Hz. Even if the harmonics are stable, noise may arise from settling time artifacts that occur when a multiplexer switches between channels with vastly differing voltages.

In order to minimize high-frequency interference each SP dipole was terminated with a passive lowpass RC filter at its input to the multiplexer. The design of this filter is presented in Appendix A. A test was conducted at the Springdale site to determine the filter's effectiveness. Measurements were recorded between the reference electrode and three electrodes located within the SP monitoring array immediately prior to the start of monitoring. As shown in Figure 4.4, the RC filter reduced high-frequency noise, and also low frequencies presumably appearing as a consequence of aliasing, by a factor of approximately 10 (e.g., from about 5 mV to 0.5 mV peak to peak) over the reduction obtained using the CR1000 data logger's 60 Hz integration sampling mode alone.

The final noise reduction step involved minimal digital post-processing using a non-linear Savitzky-Golay filter which removed intermittent high frequency spikes while preserving the primary frequency content of the SP signal. This filter involved least-squares fitting of a 4th order polynomial to each 11 point (33 second) window of each SP time series. Generally, this filter is well suited to remove intermittent high amplitude noise spikes without smoothing all higher frequency content in the manner of standard moving average digital filters (e.g. Maineult et al. 2008; Schafer 2011).

4.5 SP results from Springdale site

4.5.1 SP during pump test

Figure 4.5 shows the filtered SP response to pumping observed at four different distances (19 to 32 m) and azimuths from the pumping well, and the drawdowns recorded in the pumping well and monitoring well. Hydraulic head in the pumping well dropped 0.6 m within a few minutes of the onset of pumping, followed by a gradual decrease in the

drawdown rate. The rapid drop at the start of pumping observed in both the pumping well and monitoring well is presumably due to very low storativity of the confined aquifer ($\sim 10^{-5}$ estimated in DesRoches et al. (2012)). Because the pump triggers on and off based on the water level within a storage tank, the pump shut off before the aquifer had reached a steady state drawdown condition. Maximum drawdown in PW07-1 after 300 minutes of pumping was 1.4 m. In MW07-1 located 29 m away, water levels declined a total of 1.1 m following a similar drawdown curve. Results show a strong inverse correlation between drawdown in the wells and SP observed at the surface electrodes (Figure 4.5). This correlation suggests the recorded rise in SP is associated with the drop in hydraulic head. At electrodes e1 and e2 the SP changed abruptly by up to 1.5 mV within the first minute of pumping and thereafter changed more gradually - similar to the pattern observed in pumping well water level. The total rise in SP was approximately 7 mV at electrode e1, and 2.5 – 5.5 mV at the other three electrodes shown. SP measurements were smoothed by fitting the data to a third-degree polynomial curve to highlight the data trend. The overall SP changes observed during the test were significantly larger than the high frequency scatter of approximately ± 0.5 mV evident in Figure 4.5. It was noted however that noise levels tended to be higher for electrodes near the pumping well (e.g. Electrodes e1 and e2), compared to electrodes at greater distances (e.g. Electrodes e9 and e10). Suski et al. (2004) concluded that noise levels recorded with self-potential signals are sensitive to the rate which water is drawn from the aquifer.

Plotting the SP recorded at electrode e10 versus drawdown in the adjacent monitoring well (Figure 4.6) yields a reasonably linear relationship as predicted by eq.(4.10). The slope of the best fitting straight line through the entire data set was used to determine

an apparent voltage coupling coefficient C_{ha} of -3.4 mV/m for the confined aquifer. The line fit produced a root mean square error (RMSE) of 0.18 mV with a coefficient of distribution of 0.76.

The smoothed SP data from the electrode array were interpolated, as shown in Figure 4.7, to show their spatial and temporal evolution after 10, 60, 100 and 300 minutes of pumping. After 60 minutes (Figure 4.7b), a weak positive SP response had developed northwest to south of the pump. Progressively, SP signals intensified across the electrode array, with the anomaly remaining strongest northwest to south of the pump (Figure 4.7d). This differs from the radially symmetric pattern that would be expected from a homogeneous aquifer. The asymmetry is potentially related to hydraulic heterogeneity within the confined aquifer; for example, SP sources would be expected where hydraulic conductivity contrast exists along the boundary between the channelized sandstone unit and the lower permeability surrounding rocks.

4.6 Numerical modelling

Numerical modelling following the approach of Sill (1983) outlined above, was carried out using Comsol Multiphysics (COMSOL v4.4a) to investigate the sensitivity of SP to pumping in a heterogeneous and confined bedrock aquifer similar to the one at Springdale. The goal of our modelling was to assess hydraulic and electrical conditions under which the SP distribution at surface could be expected to produce a linear correlation with modeled hydraulic head. Sensitivity analyses are performed to explore the effects of aquifer heterogeneity, confining layer presence and conductivity, and the influence of metallic well casings on surface SP.

For computational efficiency this study considered an axisymmetric model centered on a pumping well with a radial extent of 200 m and depth extent of 100 m. The upper 10 m of the model was intended to represent the overburden sediments encountered at Springdale with the upper 8 m representing alluvial sand and 2 m of clay-rich glacial till along the base. The underlying aquifer was represented by fractured shale throughout the majority of the model. In the heterogeneous model a 10 m thick permeable channelized fractured sandstone unit extends 50 m from the well between 16 to 26 m depth. The pumping well, representing PW07-1, was cased through overburden to 12 m depth.

SP modelling requires the assignment of both hydraulic and electrical boundary conditions. The vertical boundaries located 200 m from the pump were assigned a constant hydraulic head of 0 m ($h_o = 0$ m) meaning the water level in a monitoring well completed in the aquifer at that location would be at ground surface. SP was calculated relative to a reference point on the surface on this boundary, located far enough to not be influenced by pumping. Both recharge from surface and discharge have been neglected in the model. Accordingly, the upper and lower domain boundaries were defined as no flow. Initial hydraulic head conditions throughout the entire domain were set to the ground surface ($h_o = 0$ m). Gravitational forces influencing groundwater flow through the domain were neglected. For purposes of presentation we show simulation results within a portion of the modelling domain, extending 80 m away from the well and to a depth of 40 m.

A mass flux condition was applied to the boundary of the well at a constant pumping rate (Q) of $7.8 \text{ L}\cdot\text{s}^{-1}$ ($0.0078 \text{ m}^3\cdot\text{s}^{-1}$) for 300 minutes, similar to the pumping conditions during SP monitoring at the Springdale site. The assigned flux was weighted along the well boundary to account for differences in hydraulic conductivity and layer thickness (i.e.

transmissivity) between the channelized sandstone unit and the surrounding shale units (e.g., Neville and Tonkin 2004). In all model calculations the well casing is represented in the electrical model as a highly conductive 12 m long cylinder with inner radius of 7.5 cm, and wall thickness of 1 cm. The casing is given an electrical conductivity of roughly 7 orders of magnitude higher than the channelized sandstone unit ($10000 \text{ S}\cdot\text{m}^{-1}$), which provides a pathway for current to reach the surface (Ishido et al. 1983; Kawakami and Takasugi 1994; Titov et al. 2005; Darnet et al. 2006).

Each of the model layers were assigned hydraulic and electrical conductivities, storativity, and electrokinetic current coupling coefficients. We used approximate hydraulic conductivities determined through analysis of pump test data in DesRoches et al. (2012), and mean electrical conductivities of the two bedrock types determined from normal resistivity logging in borehole MW07-1. The channelized fractured sandstone unit was assigned a hydraulic conductivity of $10^{-3} \text{ m}\cdot\text{s}^{-1}$, with an electrical conductivity of $5.5 \times 10^{-3} \text{ S}\cdot\text{m}^{-1}$ ($180 \text{ }\Omega\text{m}$ resistivity), compared to the surrounding fractured shale unit with a hydraulic conductivity of $10^{-6} \text{ m}\cdot\text{s}^{-1}$ and electrical conductivity of $8.3 \times 10^{-3} \text{ S}\cdot\text{m}^{-1}$ ($120 \text{ }\Omega\text{m}$ resistivity). Geological logs, combined with the observed lack of hydraulic connection between the surficial and bedrock aquifers, led us to conclude that the bottom 2 m of the overburden was clay-rich till of relatively high electrical conductivity and low hydraulic conductivity compared to the underlying bedrock aquifer. Although we did not measure the electrical conductivity of the alluvial sand and till units, we assigned them values of $3.3 \times 10^{-3} \text{ S}\cdot\text{m}^{-1}$ ($300 \text{ }\Omega\text{m}$ resistivity) and $5.0 \times 10^{-2} \text{ S}\cdot\text{m}^{-1}$ ($20 \text{ }\Omega\text{m}$ resistivity).

The till layer is made effectively impermeable by imposing a very low hydraulic conductivity ($10^{-14} \text{ m}\cdot\text{s}^{-1}$). This isolates the surficial alluvial aquifer from the effects of

pumping with the result that hydraulic head in the surficial aquifer does not deviate from the initial condition $h = h_0 = 0$ m. Aquifer storativity is uniformly assigned to all model domains as 8.0×10^{-5} , which is consistent with values of storativity determined through pump tests (DesRoches et al. 2012). For simplicity, we are assuming that all geological units have the same electrokinetic voltage coupling coefficient which we set equal to the apparent voltage coupling coefficient (G_{ha} of $-3.4 \text{ mV} \cdot \text{m}^{-1}$) measured in the field.

A finite element mesh extends across the entire model, with mesh refinements in locations where hydraulic and SP gradients were expected to be high (e.g. near the pumping well and steel casing, the edges of the channelized sandstone unit, and the confining layer).

4.6.1 Heterogeneous and confined model results

Hydraulic heterogeneity was included as a channelized sandstone unit that is pierced by the well and has a hydraulic conductivity three orders of magnitude ($10^{-3} \text{ m} \cdot \text{s}^{-1}$) greater than the surrounding bedrock ($10^{-6} \text{ m} \cdot \text{s}^{-1}$). The hydraulically conductive channel was modeled as a confined aquifer in order to examine what influence the confining layers, and their electrical conductivities, have on the distribution of SP measured on the surface (Figure 4.8).

After 300 minutes of simulated pumping, hydraulic head at the top of the confined aquifer (base of the confining layer) reaches a maximum drawdown of 2.0 m at a distance of 1 m from the well (Figure 4.8(a)). Over most of the channelized unit, the drawdown remains constant (at ~ 2.0 m), showing how the very high hydraulic conductivity of that unit dominates head distribution at the top of the confined aquifer. The high-weighted flux assigned to the well boundary along the channel results in the vast majority of water being drawn through the channel. An inflection in the drawdown surface and strong gradients in

head (see Figure 4.8(a) and (b)) are observed around the edges of the channelized unit at a distance of 50 m from the well. Hydraulic head rises relatively abruptly beyond that point reflecting the higher hydraulic gradients associated with the lower hydraulic conductivity of the shale.

Similar to the drawdown profile, SP modeled along the ground surface (Figure 4.8(c)) clearly shows an inflection at a distance of roughly 50 m from the well that corresponds to hydraulic and electrical conductivity contrasts at the boundary between the channel and the shale unit (Figure 4.8(d)). Based on eq.(4.9), SP sources are generated by convection currents crossing the boundary of the sandstone channel as a consequence of flow $\mathbf{u}$ perpendicular to contrasts in electrical conductivity σ , and hydraulic conductivity K . Positive current sources are also generated along the producing length of the pumping well which represents a fluid sink. As a result, SP on the surface reaches a maximum magnitude of ~ 2.6 mV in the vicinity of the pumping well and conductive casing, and shows a decrease away from the well (Figure 4.8(c)) similar to the modeled drawdown. Convective current sources are not generated within the confining till and alluvial units because there is no flow generated across the boundary during pumping.

The presence of a conductive casing used in the model generated a maximum difference in SP magnitude on surface of only 0.2 mV or approximately 3% despite the casing being located directly above the primary hydraulic source. On the surface the influence of the conductive casing on the SP magnitude drops off at a distance of approximately 3 m from the well. On the other hand, the shunting effect of the electrically conductive clay till layer above the pumping interval is expected to reduce the SP response observed on the surface (c.f. Revil et al. (2003)). Titov et al. (2005) modeled this effect by

varying the conductance (conductivity-thickness product) of a confining layer by factor of 10, from 0.4 to 4 S, and showed that the SP magnitudes expected at surface decreased by a factor of six. In the model the electrical conductivity was varied for the 2 m thick confining till layer by three orders of magnitude, from 0.003 to 1 $\text{S}\cdot\text{m}^{-1}$ (conductance of 0.006 to 2 S); and found that the predicted SP profile on surface retained the same shape but varied in amplitude from roughly 3.0 to 0.5 mV next to the pumping well.

4.6.2 Model sensitivity analyses

We now consider the influence of the confining layer and the channelized sandstone unit on the distribution of hydraulic head and SP through the model. Removing the confining layer and overburden from the model produces hydraulic head contours that extend throughout the upper portion of the aquifer (Figure 4.9). Drawdown in this model reached ~ 1.7 m after 300 minutes of pumping, at a distance of 1 m from the well. For simplicity, due to the short pumping duration we ignored the effects of the vadose zone that would develop in this unconfined aquifer case since the magnitude is considered negligible compared to the current sources generated due to pumping (e.g. Darnet et al. 2003). However, if pumping durations were longer then the effect of water table drawdown then pore drainage would need to be explored in greater detail. The lower drawdown modeled compared to the confined aquifer case is a consequence of water being drawn from the entire domain, as opposed to being limited to a confined aquifer only. Despite the reduced drawdown response, SP magnitude on the surface increased by approximately 1 mV (e.g. to 3.4 mV from 2.6 mV near the pumping well). This increase in SP is in response to removing the glacial till layers with higher electrical conductivity from the model (e.g. Titov et al. 2005). Similar to the heterogeneous confined model, SP sources arise from flow

$\mathbf{u}$ perpendicular to contrasts in electrical conductivity σ , and hydraulic conductivity K at the boundary of the sandstone channel, and divergence of flow induced by pumping. The SP results from this unconfined case still exhibit an inflection in the SP profile line, 50 m from the well overlying the edge of the heterogeneous sandstone unit. This is consistent with modeled results from both Titov et al. (2005) and Sheffer and Oldenburg (2007) who showed that hydraulic heterogeneities can greatly influence the distribution of SP magnitude and character measured on the surface.

Removal of the channel unit from the confined model produces a distinct influence on the distribution of head and SP (Figure 4.10). This model now represents a homogeneous confined aquifer with water drawn from the low permeability ($K = 10^{-6} \text{ m}\cdot\text{s}^{-1}$) fractured shale. In the absence of the channel unit, a maximum drawdown of ~ 22 m is reached after 300 minutes of pumping at a distance of 1 m from the well. Drawdown rapidly decreases as a function of distance, and approaches zero roughly 100 m from the well. The resulting SP calculated on the ground surface exhibits an inverse correlation with the hydraulic results, and decreases away from the well (Figure 4.10(c)), although, the rate of SP decline with distance away from the well is more gradual than the rate of drawdown decline measured at the base of the confining layer. Near the well, SP magnitudes are approximately 6.5 times higher (~ 17.0 mV at 1 m away from the well) compared to the results including the sandstone channel. This change is attributed to increased hydraulic gradients within the less permeable shale unit. Titov et al. (2005) shows that a decrease in the hydraulic conductivity of an aquifer for a constant pumping rate results in an increase in hydraulic gradient in the unit, and an increase in hydraulic source term. This can be appreciated from inspection of the conduction current source terms in eq.(4.8); the first

three terms all increase with ∇h and the fourth term increases with $\frac{1}{K}$ if the divergence of fluid flux $\mathbf{u}$ remains unchanged. For the homogeneous and confined aquifer modelled in this case, there is no flow across layer boundaries; the pumping well, responsible for a divergence in fluid flow, is the sole SP source (eq.(4.9)). Therefore, SP at surface decreases monotonically away from the pumping well, in contrast to the inflection observed above the edge of the channelized sandstone in the heterogeneous models.

Figure 4.11 indicates that, in all three scenarios, modeled SP and drawdown 25 m from the well shows strong linear correlations. This linearity during pumping is in agreement with the simple relationship between SP and drawdown predicted by eq.(4.10) even though the assumption of homogeneity and zero net current at each point in the subsurface is not satisfied. The linearity observed in our modelling results is also in agreement with our observations at Springdale MW07-1 (Figure 4.6), and with observations made by Revil et al. (2003), Rizzo et al. (2004), and Suski et al. (2004). We note however, that the proportionality breaks down for smaller drawdowns occurring early in the pumping period, in particular for the homogeneous confined aquifer case (Figure 4.11). Although drawdown in the model, as at the monitoring site, had not achieved steady-state conditions, results show that the linear correlation stabilized over the duration of the drawdown period. Poor linearity observed during early drawdown period is consistent with observations by Straface et al. (2010).

The slopes of the linear relationships shown in Figure 4.11 represent modeled apparent electrokinetic voltage coupling coefficients (C_h). Slopes for the two confined aquifer models, determined by linear regression over the entire range of drawdowns, are -1.2 and -1.3 mV·m⁻¹ for the heterogeneous and homogeneous cases, respectively (Figure

4.11). The coefficient for the unconfined and heterogeneous aquifer model is nearly two times larger ($-2.0 \text{ mV} \cdot \text{m}^{-1}$). This can be explained by the removal of the electrically conductive confining till layer from the model, reducing the overall conductance of the aquifer system (e.g. Titov et al. 2005). Physically, the conduction currents flowing near surface give rise to larger voltage drops (SP) because of the increased resistivity of the near surface region.

Figure 4.12 shows the relationship between the modeled SP on the surface and drawdown at the top of the heterogeneous confined aquifer. Values are taken at different distances from the pumping well (1m, 20m, 40m, 60m and 80m) and are plotted for times between 0 and 300 minutes after the start of pumping. At larger distances from the well, the magnitudes of both drawdown and SP decrease. However, the slope of SP versus drawdown in the model (C_h) varies only subtly from -1.3 to -1.4 mV m^{-1} . Subtle changes in apparent C_h likely reflect subtle variability in heterogeneity associated with thinning of the channel and the small influence from the conductive well casing away from the well.

The modelling results presented in Figures 4.11 and 4.12 confirm that a linear relationship between SP and drawdown can be expected for both confined and unconfined aquifers under the conditions of two dimensional hydraulic heterogeneity combined with predominantly one-dimensional (layered) resistivity heterogeneity. We next apply this linear approximation to the Springdale SP data set to infer spatial variations in drawdown and for estimations of apparent (or equivalent) aquifer properties.

4.7 Interpretation of field data

4.7.1 Estimation of aquifer transmissivity and storativity

Others have used the voltage coupling coefficient and measured variations in SP during pumping to infer drawdown of the water table surface beneath arrays of electrodes using eq.(4.10) (e.g. Revil et al. 2003; Straface et al. 2007). Hydraulic properties of the aquifer have been estimated by treating the electrodes as monitoring wells.

Following a similar approach, drawdown and apparent transmissivity and storativity values are estimated here for the Springdale aquifer by utilizing drawdown curves from the two wells, and the apparent drawdown curve from 31 SP electrodes, and determining best-fits to the Theis equation for a confined aquifer (e.g., Domenico and Schwartz 1997). The Theis solution assumes that the aquifer is confined, homogeneous, isotropic, and of uniform thickness. Although these assumptions are rarely satisfied, the Theis solution has been widely used for heterogeneous aquifers as a quick approximation of transmissivities using data from monitoring wells while pumping (Wu et al. 2005). Drawdown below each electrode was estimated from eq. (4.10) using the apparent voltage coupling coefficient $C_{ha} = -3.4 \text{ mV} \cdot \text{m}^{-1}$ measured in the field and the change in SP during pumping. Each electrode effectively acts as an additional monitoring well in which the drawdown over the duration of the pumping test was determined.

The Theis solution was solved by optimizing the values of transmissivity (T) and storativity (S) using the HyTool MATLAB routines (Renard 2003; Mejías et al. 2009). The best-fit was obtained by a non-linear least square procedure, which required initial estimates for T and S that were iteratively modified until the difference between predicted and measured/inferred head values as a function of pumping time was minimized.

Aquifer transmissivity determined from the drawdown data measured in MW07-1 was $2.4 \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$. This value is very comparable to transmissivity estimates of $(5.1 - 6.3) \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$ calculated for this monitoring well, PW07-1, and three other nearby wells, using early-time drawdown data from a 72-hour pump test in the same confined aquifer (DesRoches et al. 2012). The late-time drawdown behavior differed, suggesting the influence of an impermeable barrier, attributed to the limited spatial extent of the channelized sandstone unit that was being pumped. In this case, where pumping was limited to just 5 hours, we consider it reasonable to apply the early time analysis approach to approximate apparent transmissivity of the aquifer using the Theis model.

Figure 4.13 shows a log-log plot of the apparent drawdown time series determined using SP data from electrode (e1) located 19 m from the pumping well. The plot shows the solution for the best-fit to the Theis model, which converged after 5 iterations with a mean residual of 0.013 m. Apparent transmissivity estimates from all 31 electrodes located between 9 to 36 m from the pumping well were similar ranging from $(2.0 - 4.2) \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$, with a mean value of $2.2 \times 10^{-3} \text{ m}^2 \cdot \text{s}^{-1}$. These apparent transmissivity estimates are the same order of magnitude as the afore-mentioned values determined from actual drawdowns measured in PW07-1, MW07-1 and three other nearby monitoring wells during the 72 hour pump test. The use of the apparent coupling coefficient of $-3.4 \text{ mV} \cdot \text{m}^{-1}$ resulted in a slight over-estimation of the inferred drawdown in some electrode locations (e.g. electrode e5) compared to the drawdown observed in the pumping well (PW07-1). This may be a consequence of lateral variation in resistivity and/or coupling coefficient below those locations.

Apparent storativity estimates range from $(2.9 - 4.9) \times 10^{-3}$. These storativity values were consistent with the highest of four storativity estimates (2.9×10^{-3} to 3.5×10^{-5}) determined from drawdown data measured during a 72 hour pump test (DesRoches et al. 2012). We expect that storativity estimates derived from the 72 hour pump test were reduced by the influence of channel boundaries observed after 60 minutes in the late time drawdown data.

Apparent transmissivities determined for each electrode location were interpolated to show the distribution over the confined aquifer (Figure 4.14). The interpolated area measures roughly 40 x 65 m with estimates from the 31 electrodes varying by only a factor of two. In Figure 4.14, the relative consistency of apparent transmissivity values across the study area is encouraging. Apparent transmissivities show a pattern of consistently higher values in the northeastern part of the study area, and lower transmissive regions south and west of the pumping well. Areas of higher SP response during pumping (see Figure 4.7) correspond to areas that are interpreted to have relatively lower transmissivities. It is recognized that the homogeneous, isotropic, and uniform thickness assumptions of the Theis solution make interpreting the transmissivity distribution difficult for a heterogeneous aquifer (Wu et al. 2005). More sophisticated hydraulic tomography techniques may be required to fully understand the distribution.

4.8 Discussion

4.8.1 The apparent voltage coupling coefficient (C_{ha})

For the purpose of SP modelling, the electrokinetic voltage coupling coefficient is an important input parameter to determine the magnitude of conduction current sources. We

found the apparent voltage coupling coefficient $C_{ha} = -3.4 \text{ mV} \cdot \text{m}^{-1}$ reasonable since it fell within -1 to $-10 \text{ mV} \cdot \text{m}^{-1}$, which is typical of fresh groundwater systems (Revil et al. 2003). This apparent voltage coupling coefficient assigned to the heterogeneous and confined aquifer is nearly three times larger than the value determined by linear regression of the modeled SP and hydraulic data ($-1.2 \text{ mV} \cdot \text{m}^{-1}$ (see Figure 4.11)). Differences were also noted by Titov et al. (2005) who found that the apparent coupling coefficient predicted by modelling ($-1.5 \text{ mV} \cdot \text{m}^{-1}$) was 2.7 times smaller than the value of C_h they assigned to their model layers. The difference between the value of C_h assigned to the model layers and the apparent C_{ha} determined from the modeled SP-drawdown relationship can be attributed to subsurface heterogeneity which causes the dominant flow paths for conduction currents and convection currents to differ, in violation of the assumption implicit in eq.(4.10) that total current $j = 0$ everywhere in the subsurface. The influences of an impermeable and electrically conductive confining layer and of the electrically conductive well casing are expected to be particularly important.

The discrepancy between the apparent coupling coefficient measured at Springdale and that obtained by numerical modelling of a heterogeneous confined aquifer could have been eliminated by assigning the geological units a coupling coefficient C_h that was higher than the measured C_{ha} . The application of this field measurement (C_{ha}) may be questioned given that the monitoring well was open between 12 and 43 m depth and therefore did not sample drawdown exactly at the base of the confining layer. In that respect, it is interesting to consider other estimates we might have selected for representative values of C_h in rocks at the Springdale site.

Voltage coupling coefficients were determined on five crushed silty sandstone samples from the field area in the laboratory using the methodology in Morgan et al. (1989). The measurements ranged from -3.0 to $-9.7 \text{ mV}\cdot\text{m}^{-1}$ with a mean value of $-6.2 \text{ mV}\cdot\text{m}^{-1}$ after correcting for the fact that water used in the lab was more electrically conductive than the water in the borehole that is thought to be representative of in situ groundwater ($270 \text{ }\mu\text{S}\cdot\text{cm}^{-1}$ in the lab compared to $147 \text{ }\mu\text{S}\cdot\text{cm}^{-1}$ in the borehole). Revil et al. (2003) showed that, given knowledge of groundwater conductivity σ_f ($\text{S}\cdot\text{m}^{-1}$), the voltage coupling coefficient could be approximated for differing sedimentary and volcanic lithologies using the log-log correlation, $\log_{10} C_h = -0.921 - 1.091 \log_{10} \sigma_f$, where C_h is given in $\text{mV}\cdot\text{m}^{-1}$. Jouniaux and Ishido (2012) recognized a similar trend for sands and sandstones (in water of pH 7-8) where the coupling coefficient was approximated by $C = -1.2 \times 10^{-8} \sigma_f^{-1}$, with C given in $\text{V}\cdot\text{Pa}^{-1}$. Using both approaches with the observed borehole fluid electrical conductivity of $0.0147 \text{ S}\cdot\text{m}^{-1}$ yields voltage coupling coefficients of -6.4 and $-7.8 \text{ mV}\cdot\text{m}^{-1}$, which are nearly consistent with the mean voltage coupling coefficient determined in the lab, and are also within a factor of two of the apparent voltage coupling coefficient determined in the field.

The apparent voltage coupling coefficient measured in the field using eq. (4.10) was initially used to estimate drawdown below each electrode and determine hydraulic parameters using the Theis solution. If we instead estimate drawdowns using the mean voltage coupling coefficient from lab measurements ($-6.2 \text{ mV}\cdot\text{m}^{-1}$), or using the coupling coefficient from our numerical model ($-1.2 \text{ mV}\cdot\text{m}^{-1}$), rather than the apparent voltage coefficient measured in the field ($-3.4 \text{ mV}\cdot\text{m}^{-1}$), the mean transmissivity estimates increase or decrease by factors of 1.8 or 1.4, respectively. Despite these changes, transmissivity

estimates derived from the SP data remain within an order of magnitude of those derived from actual drawdown measurements.

4.8.2 Influence of conductive casing

Titov et al. (2005) and Darnet et al. (2003) showed that a conductive well casing in the upper part of the model acts to shunt electrical currents to the surface and thereby increase SP response near the well. Darnet et al. (2003) modeled this effect for a homogeneous unconfined aquifer using a 20 m long conductive casing, with an inner radius of 9 cm, wall thickness of 1.3 cm and electrical conductivity of $10^8 \text{ S}\cdot\text{m}^{-1}$. Their conductive casing was positioned directly above a 5 m interval (between the casing bottom at 20 m and the base of the aquifer at 25 m) where water was drawn during pumping. Such a well casing configuration would be a significant difference from our bedrock aquifer case where water is drawn into a very long section of open hole (between the bottom of casing at 12 m and the bottom of the hole at 43 m) that is 2.5 times longer than the conductive casing. The fact that their pumping was all concentrated directly below the casing over a shorter interval than we considered, coupled with the fact that their conductive casing was nearly twice as long may explain why the effects of their conductive casing on surface SP was so much more important than we found. This is consistent with Schenkel and Morrison (1990) who showed that the well casing effect on the electrical response on the surface would have been more profound if the open section of hole had been shorter, therefore concentrating the current source closer to the base of the casing.

In Figure 4.16, we compare surface SP profiles extending from the well head to 40 m radius, calculated using conductive and insulating casing in each of our three aquifer models. In the case of the homogeneous confined aquifer (Figure 4.15(a)), the effect of

conductive casing immediately adjacent to the well is significant, increasing SP up to 25%. However, the SP enhancement decays rapidly with distance, decreasing to less than 2% at a radial distance of 10 m. The casing is relatively important in this case as the only current source is the water being pulled into the well beneath it. Current from this source is readily channeled up through the metallic casing to enhance SP at surface. The effect would presumably be even more pronounced if the open section of the pumping well was shorter, concentrating more flow close to the base of the casing.

In the case of the heterogeneous confined and unconfined aquifers (Figure 4.15(b and c)), the effect of conductive casing is significantly reduced compared to the homogeneous confined case, with a less than 3% increase of SP on the surface immediately adjacent to the well. One major difference in the heterogeneous cases is that the channel unit has a higher hydraulic conductivity, thereby reducing the strength of the induced hydraulic gradient near the well and current source generated beneath the casing, as shown by the fourth term in eq. (4.8). In addition, current sources generated in the heterogeneous cases exist along the edges of the channelized sandstone extending far away from the casing, with the result that their pathways are not strongly influenced by the presence of metallic casing. It is apparent that the conductive casing is more important for conveying conduction currents to surface when there is an electrically conductive confining layer partially screening or shunting currents originating below it. By removing the overburden units from the heterogeneous model, the overall SP magnitude on surface is enhanced but the effect of the conductive casing on the SP magnitude is further reduced (Figure 4.15(c)).

4.8.3 Interpreting transmissivity distribution

SP surveys offer great potential to supplement sparsely-distributed monitoring wells in defining drawdown patterns. It is up to hydrogeologists to determine how to best make use of this rich new data source for the estimation of hydraulic properties. In heterogeneous aquifers, hydraulic tomography or geostatistical approaches, such as those implemented by Yeh et al. (1995), Wu et al. (2005), and Straface et al. (2007), have been used to recover the spatial distribution of transmissivity. Confined aquifer transmissivity estimates based on individual SP (or monitoring well) time series and the Theis equation (with its assumptions of homogeneity, isotropy and constant thickness) can be useful, but maps of such apparent transmissivities must be interpreted with caution as they can be misleading for heterogeneous aquifers (Wu et al. 2005). For example, the large amplitude drawdowns modelled over the permeable sandstone aquifer in our two heterogeneous models (Figures 8 and 9) would yield low transmissivities if inverted using the Theis equation with its assumption of confined aquifer homogeneity. For that reason, other authors (e.g., Rizzo et al. 2004; Straface et al. 2007) have worked on methods to solve for constant or spatially variable hydraulic conductivity or transmissivity that provide the best fit to a whole set of SP-derived inferred drawdown data. Those analyses are beyond the scope of this study which was to explore the applicability of the simple linear SP-drawdown relationship to confined aquifer models with limited heterogeneity. Despite the above limitations, the Theis analysis provides a valuable first approximation for transmissivity in a heterogeneous aquifer (e.g., Straface et al. 2007).

4.9 Conclusions

The use of surface SP monitoring to infer drawdown during pumping, and spatial variations in the apparent transmissivity of a heterogeneous and confined fractured bedrock aquifer has been investigated through both field measurements and numerical modelling. Pumping-induced SP anomalies were measured at 31 locations around a municipal production well. Signal-to-noise ratios were improved by a factor of ten with the aid of passive analog lowpass filters compared to those obtained using the data logger's powerline noise cancelling sampling mode alone. SP anomalies reached magnitudes of 2 - 9 mV at distances up to 36 m from the well during pumping at a rate of $7.8 \text{ L}\cdot\text{s}^{-1}$ for 300 minutes. Using an apparent electrokinetic coupling coefficient obtained from a single electrode-monitoring well pair, apparent drawdown curves were inferred below each electrode location using the SP signals over the duration of the pumping period. Inferred drawdown curves for all electrode locations were interpreted using the Theis model to provide estimates of apparent transmissivity ranging from $(2.0 - 4.2)\times 10^{-3} \text{ m}^2\cdot\text{s}^{-1}$, which were only 1.5 – 3 times smaller than estimates obtained using conventional pump tests and direct drawdown measurements.

Conditions under which a simplified semi-empirical approach to SP interpretation could be expected to yield reliable estimates of drawdown during pumping were assessed by modelling three different aquifer scenarios inspired by the field site. The models considered hydraulic heterogeneity that was strongly two-dimensional and electrical heterogeneity that was predominantly one-dimensional (layered). All three involved a conductive casing that has been shown in previous studies to increase current flow near surface and thereby increase SP effects on surface. Numerical modelling results demonstrated that, for three

relatively shallow aquifer scenarios, the effect of conductive casing was modest to negligible, amounting to at most 25% immediately adjacent to the well for a homogeneous confined case where the only SP source was flow into the well bore immediately below the casing. The hydraulic conductivity contrast between the permeable sandstone channel and the surrounding shale unit gave rise to a prominent inflection in the modelled SP profiles directly above the edge of the channel. This demonstrates that surface SP measurements are directly sensitive to hydraulic heterogeneities. It was also noted that under the same electrical conditions and pumping rate, the magnitude of SP simulated on the surface increased when the hydraulically conductive channel unit was removed, and is associated with a hydraulic gradient increase in the unit. A change in electrical conductivity structure, in particular removing the electrical conductivity of the clayey glacial till confining layer, also caused the SP magnitude to increase on the surface. SP modeled on the surface near the pumping well in each of the aquifer settings revealed linear correlations with modeled hydraulic head, despite the aquifer being both electrically and hydraulically heterogeneous.

Forward modelling and field monitoring have shown that SP data measured on the surface can be used to complement a limited number of monitoring wells to infer the evolution of a drawdown pattern during pumping and estimate the hydraulic properties of an aquifer. Our analyses suggest SP monitoring should be effective where confined aquifers are overlain by overburden of uniform conductance. In addition, our study shows that SP monitoring during pumping can be a valuable way of obtaining a high density of hydraulic data needed for tomographic studies to determine hydraulic properties of a heterogeneous aquifer. This can be conducted in a way that is non-invasive and significantly less expensive than installing tens or hundreds of monitoring wells.

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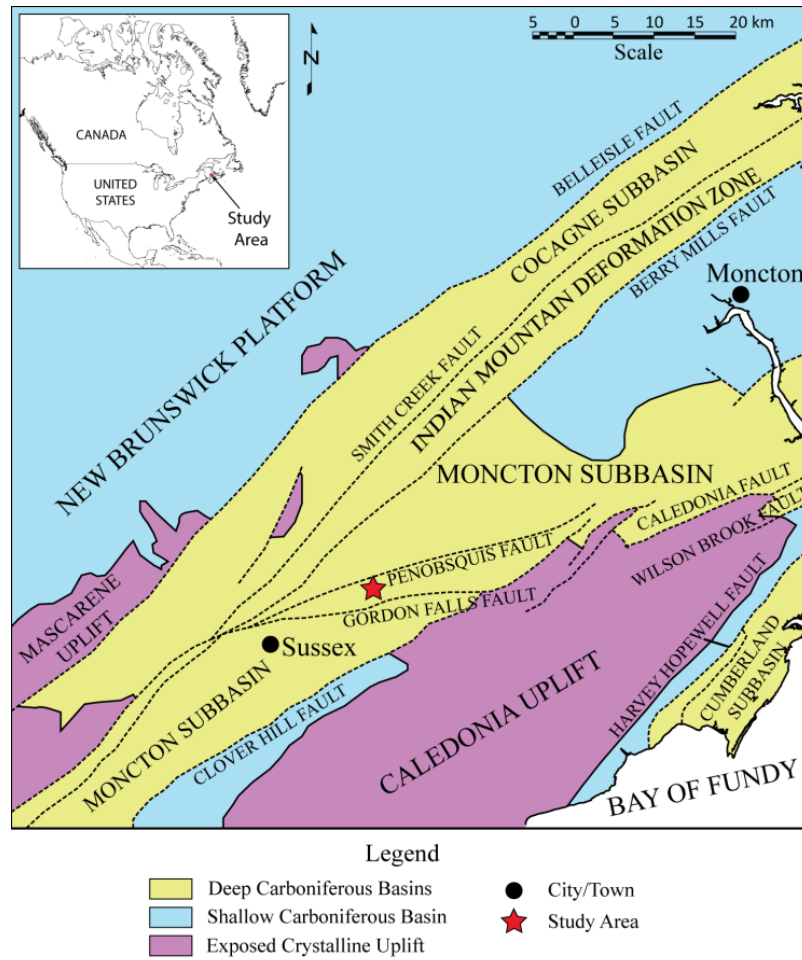


Figure 4.1. Geological map showing the study area location within the Carboniferous Moncton subbasin

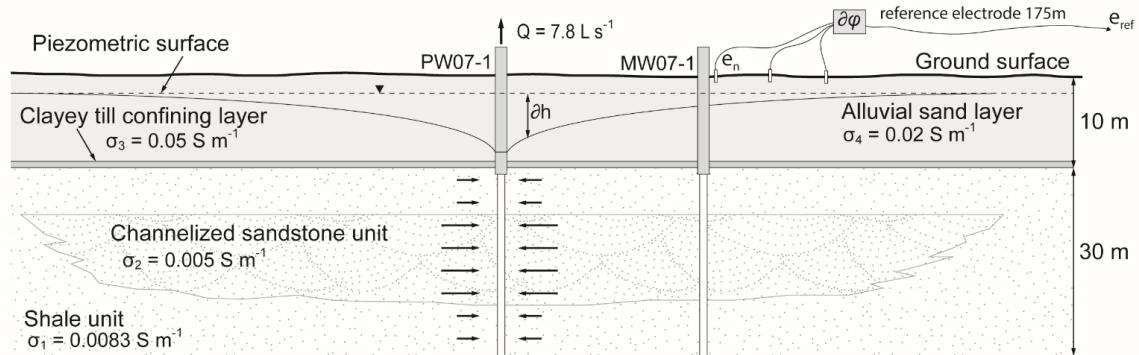


Figure 4.2. Conceptual model for the confined heterogeneous aquifer at the Springdale site. The piezometric surface extends into the overburden layers. Majority of the flow is through the channelized sandstone unit. SP was measured between observation electrodes (e_n) situated on surface within 40 m of the pumping well (PW07-1) and a reference electrode (e_{ref}) located 175 m from the pumping well

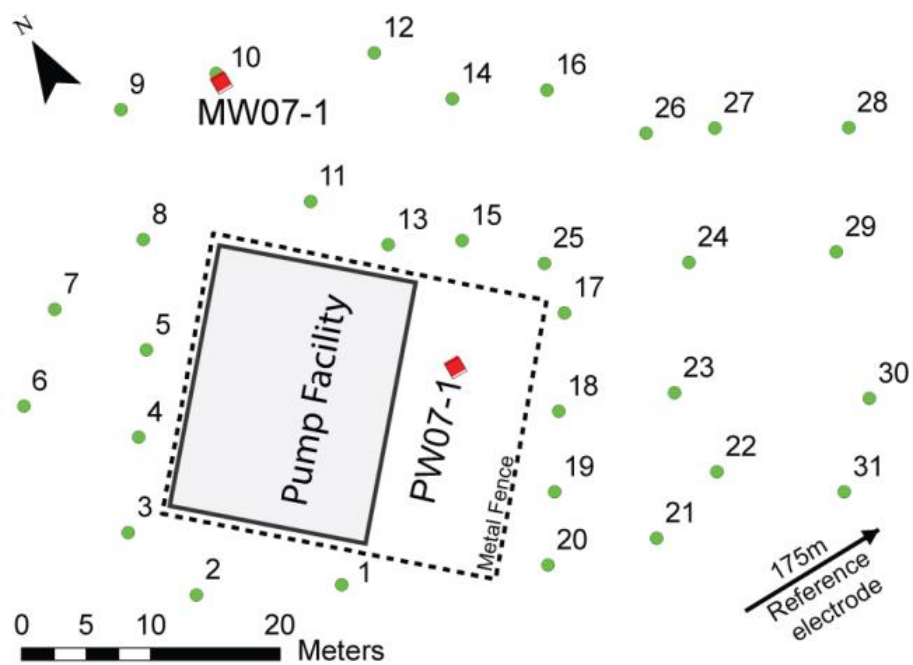


Figure 4.3. Location of electrodes (green circles) and the monitoring and pumping wells (red diamonds) for the SP survey

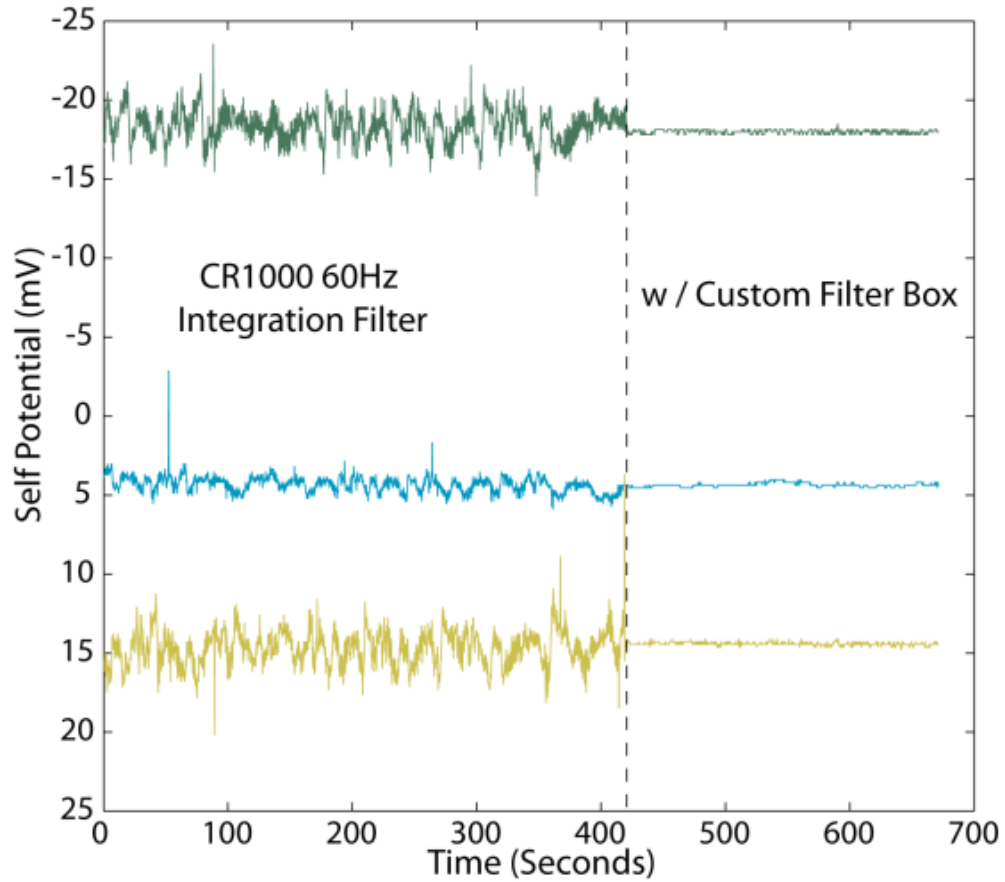


Figure 4.4. The effectiveness of the custom-built filter box using electrodes (e1, e5, and e9) in the array measured against the reference electrode (175 m away) at the Springdale aquifer monitoring site (DC offsets have been added to the traces to prevent trace overlap). The test recorded approximately 700 seconds, at a sample rate of 10 Hz, directly into three channels of the CR1000 data logger using 60 Hz integration sampling. High frequency noise was attenuated by a factor of approximately 10 by switching in the lowpass RC filter across each channel.

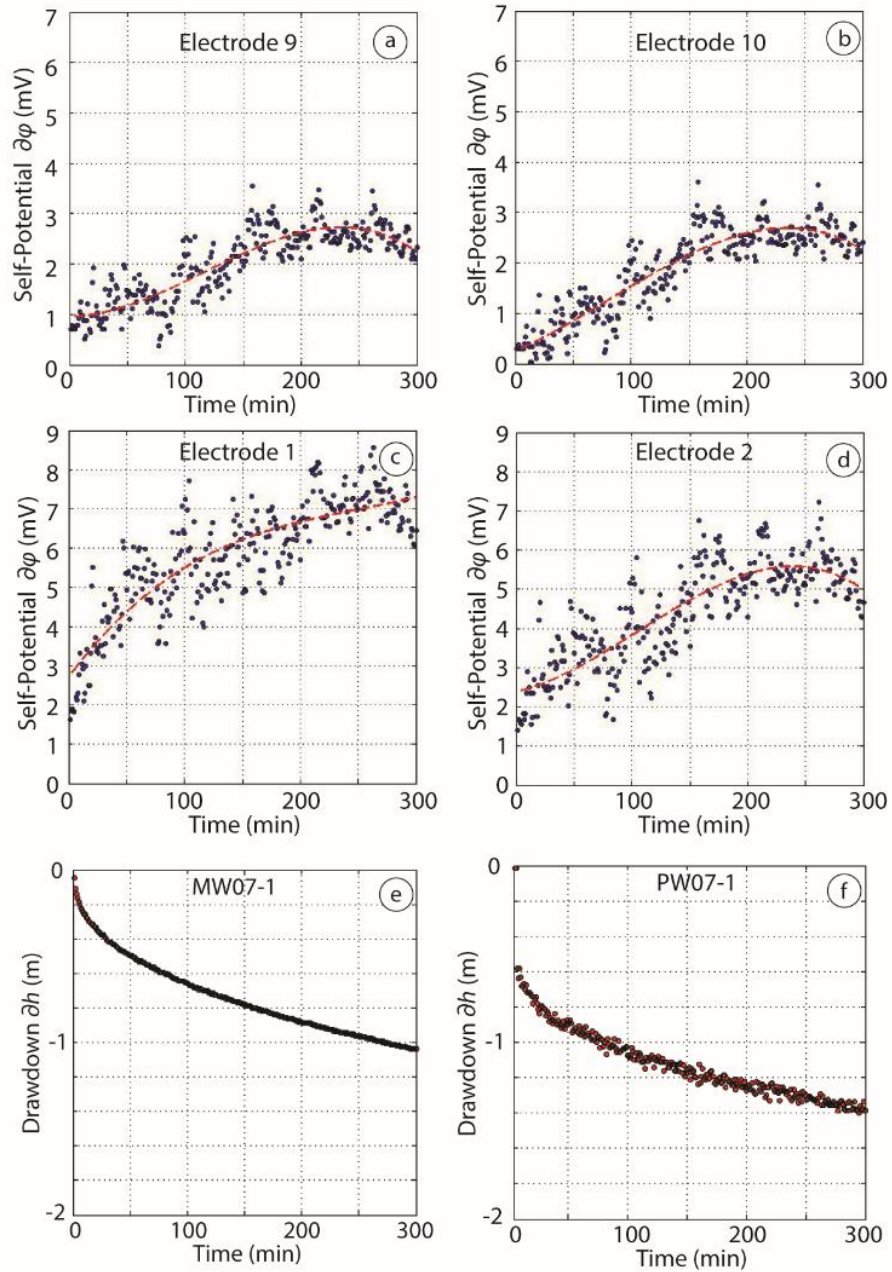


Figure 4.5. (a – d) SP data observed at four electrodes located 19 to 32 m from the borehole collar during 300 minutes of pumping. The smoothed SP curves (third-degree polynomial curves shown in red) show an abrupt change when the pump is turned on, and reach a maximum shortly before pumping is terminated. Drawdown curves (e and f) from the monitoring well (MW07-1) and pumping well (PW07-1) are consistent with the shape of the SP data.

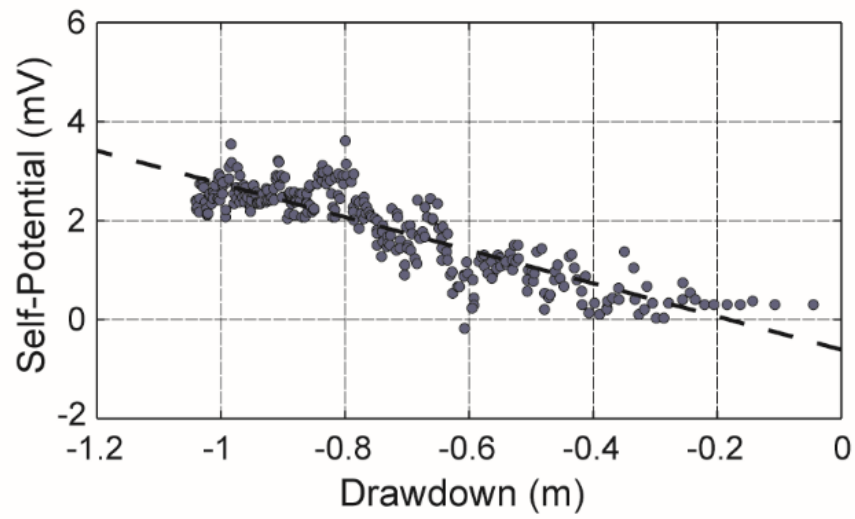


Figure 4.6. Linear regression between drawdown in the monitoring well and SP at electrode $\epsilon 10$ located 1 m from the monitoring well MW07-1 collar provides an apparent voltage coupling coefficient of -3.4 mV/m, with an R-squared value of 0.76 and root mean square error (RMSE) of 0.18 mV.

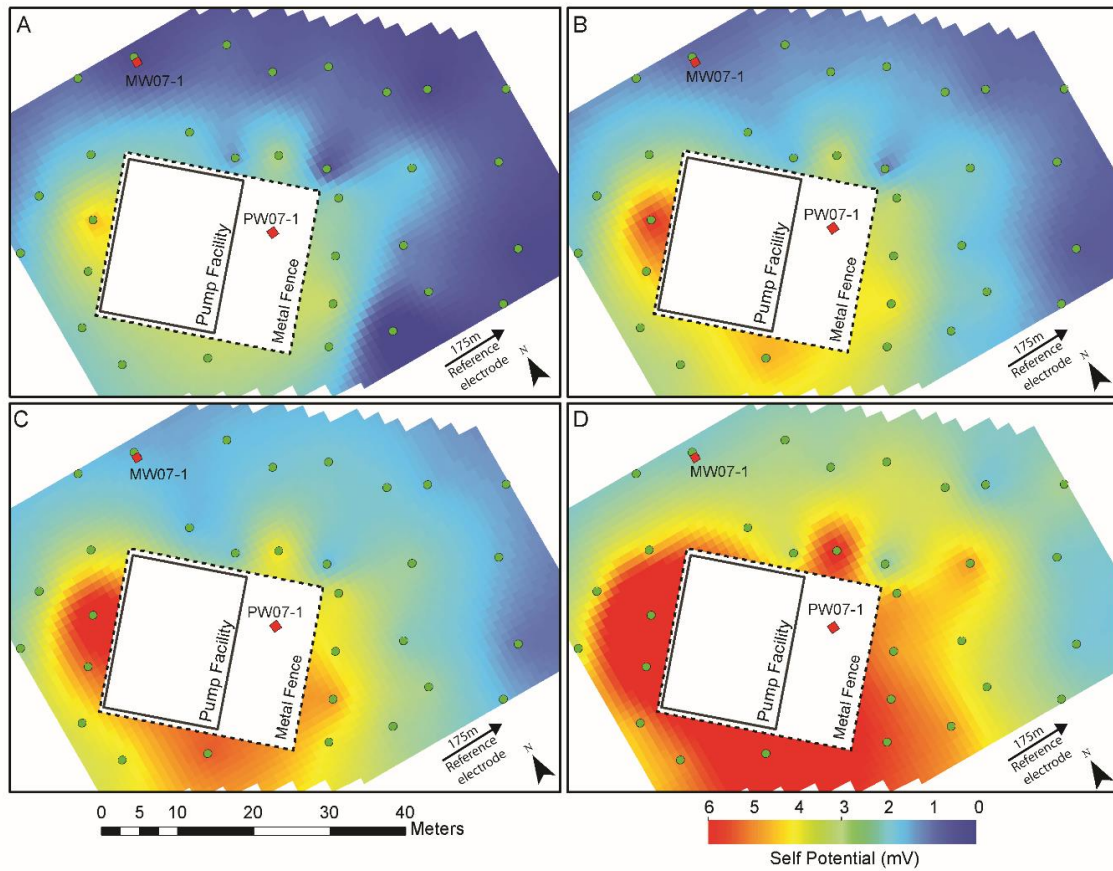


Figure 4.7. SP distribution measured with 31 electrodes around the pumping well at four different times during a 300 minute pumping cycle; (A) 10 min, (B) 60 min, (C) 100 min, and (D) 300 min. Magnitude was strongest northwest to south of the pump, unlike the radially symmetric pattern one would expect above an isotropic, homogeneous aquifer. Pumping well (PW07-1) and monitoring well (MW07-1) contain 12 m of metallic well casing through overburden units. No electrodes were located within the pump facility area. Reference electrode is located 175 m to the east of pumping well.

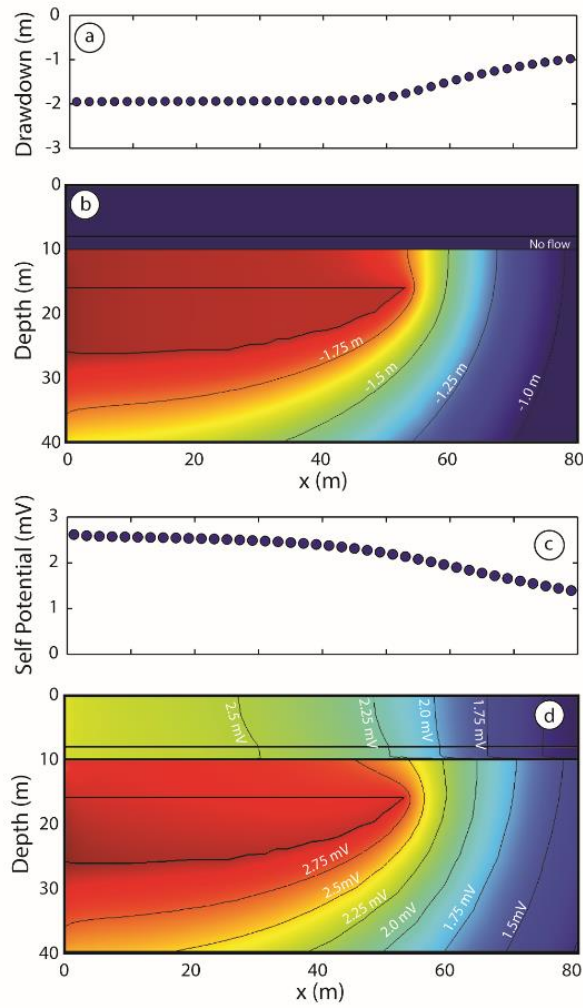


Figure 4.8. Modelling results showing the hydraulic head (a and b) and SP distribution (c and d) through a heterogeneous confined aquifer after 300 minutes of pumping. Profiles (a) and (c) show hydraulic head and SP values modeled along the base of the confining layer and on the ground surface, respectively, as a function of distance away from the pumping well.

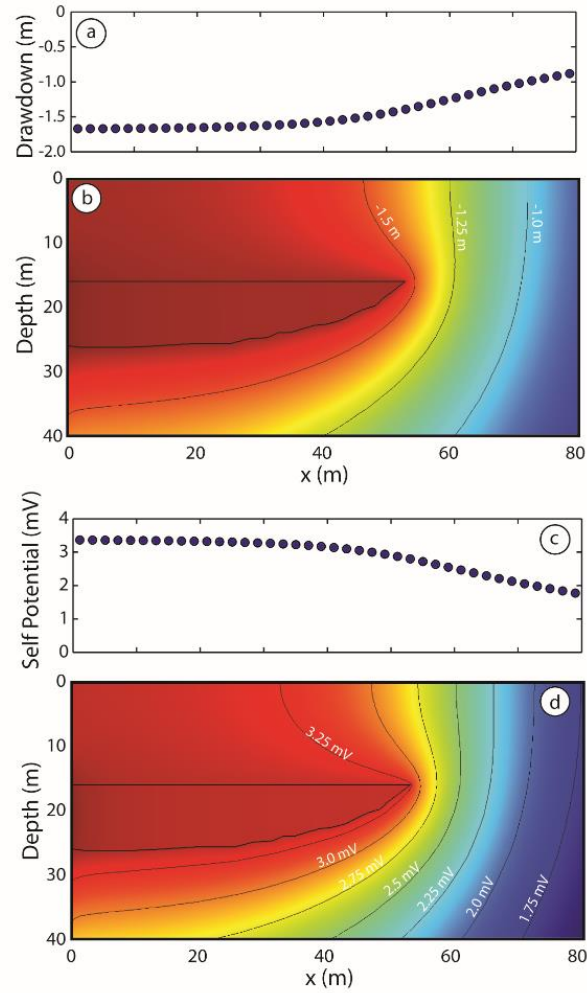


Figure 4.9. Modelling results showing the hydraulic head (a and b) and SP distribution (c and d) through a heterogeneous unconfined aquifer while pumping. Profiles (a) and (c) show the drawdown and SP values modeled along the ground surface, respectively, as a function of distance away from the pumping well.

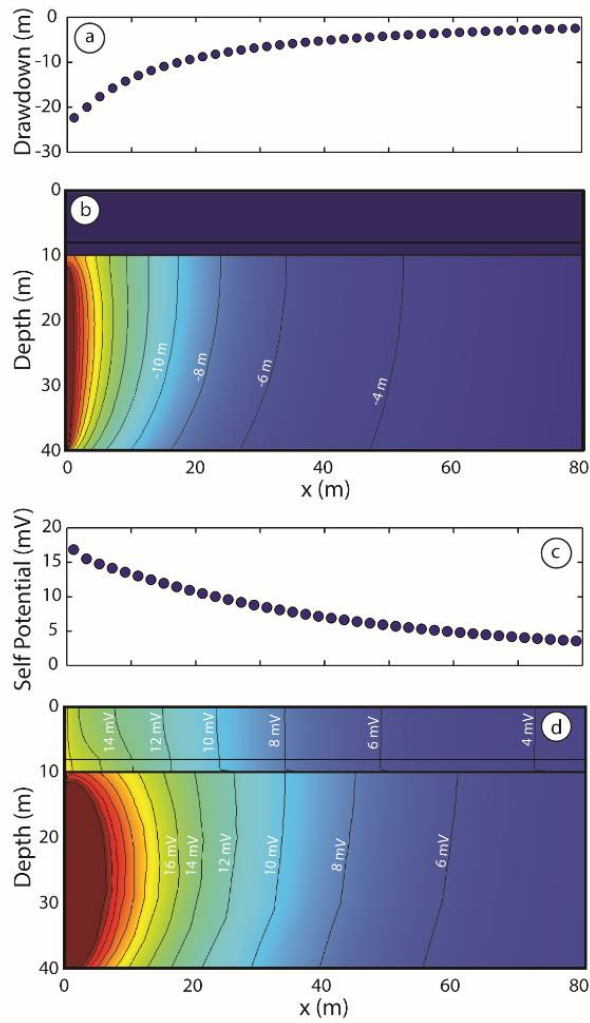


Figure 4.10. Modeled results showing the hydraulic head (a and b) and SP (c and d) distribution through a homogeneous confined aquifer generated after pumping for 300 min. Profiles (a) and (c) show the drawdown of hydraulic head and SP values as a function of distance away from the pumping well measured along the base of the confining layer and on the ground surface, respectively.

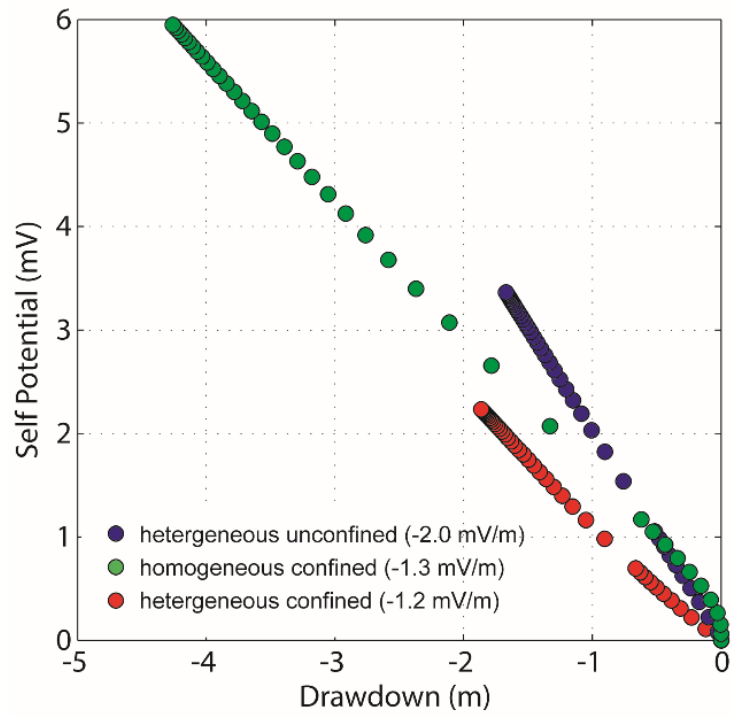


Figure 4.11. Linear correlations between surface SP and drawdown modelled at a distance of 25 m from the well during 300 min of pumping for three scenarios: heterogeneous confined aquifer (red); heterogeneous unconfined aquifer (blue); and homogeneous confined aquifer (green).

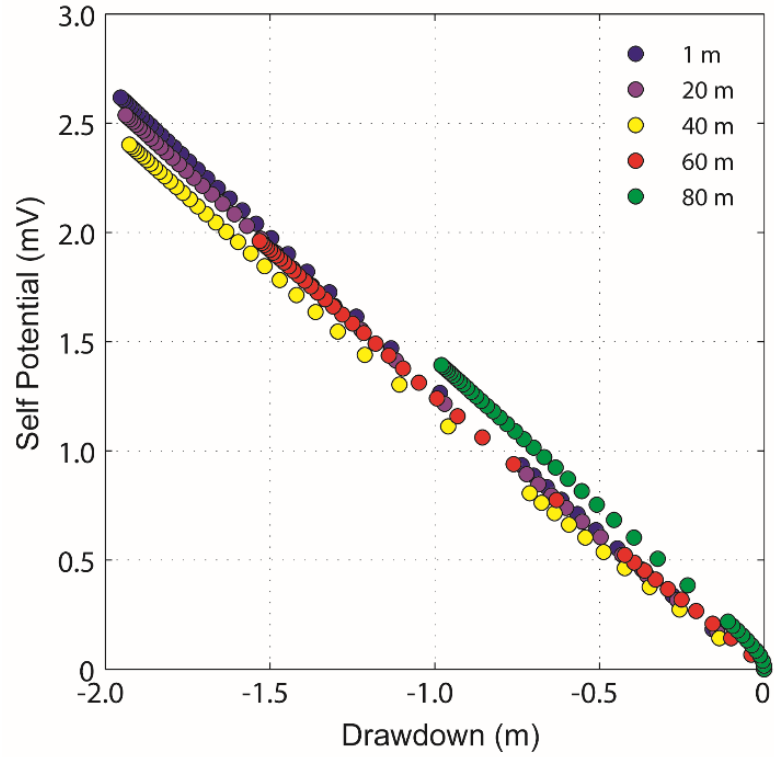


Figure 4.12. Linear correlations between surface SP and drawdown modelled for the heterogeneous confined aquifer case at five different distances from the pumping well (1m, 20m, 40m, 60m and 80m).

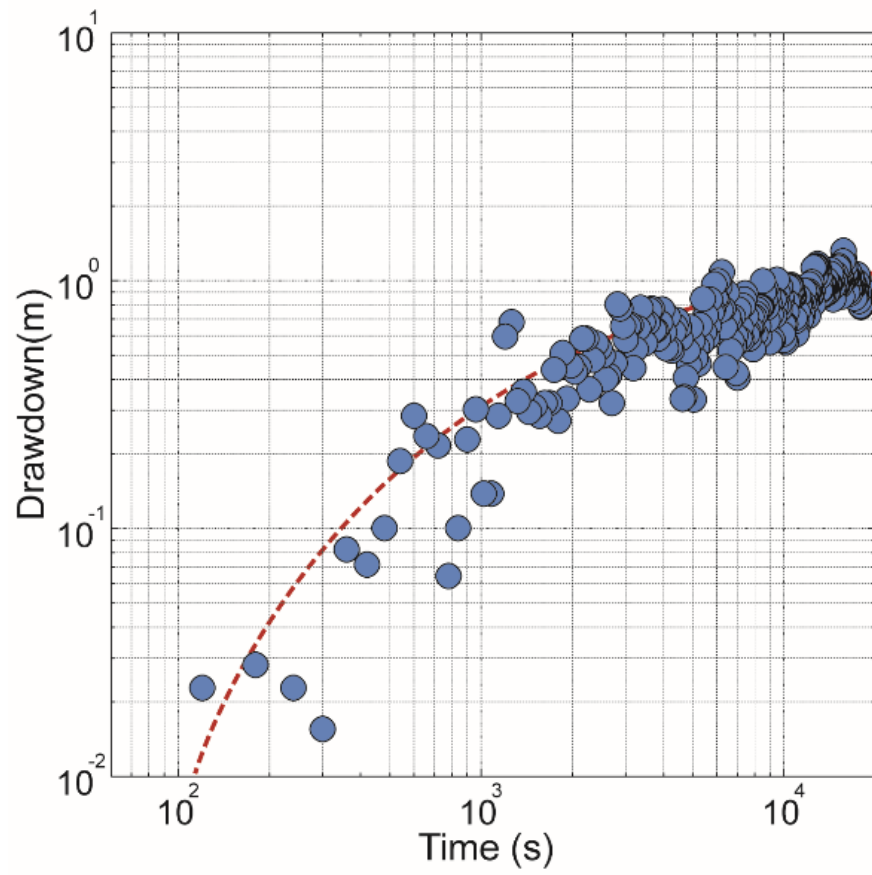


Figure 4.13. Determination of apparent transmissivity by best-fit of the drawdown calculated using self-potential data from electrode e1 with the Theis model (dashed line).

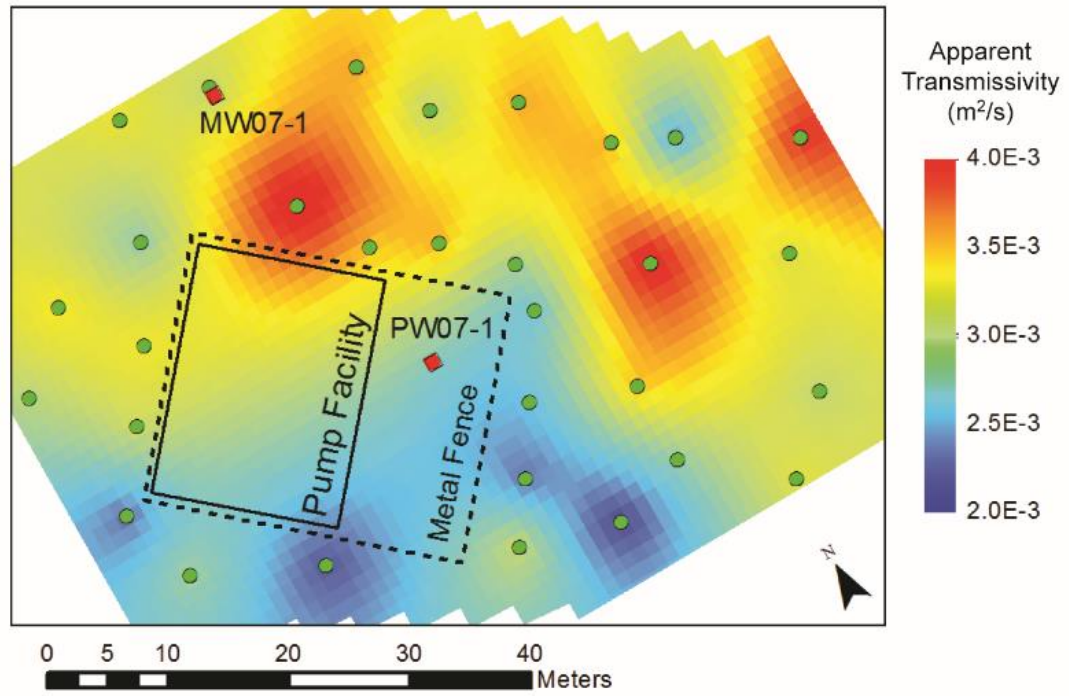


Figure 4.14. Interpolated distribution of apparent transmissivities calculated using the values determined at 31 electrodes (green points) and two well locations (red points).

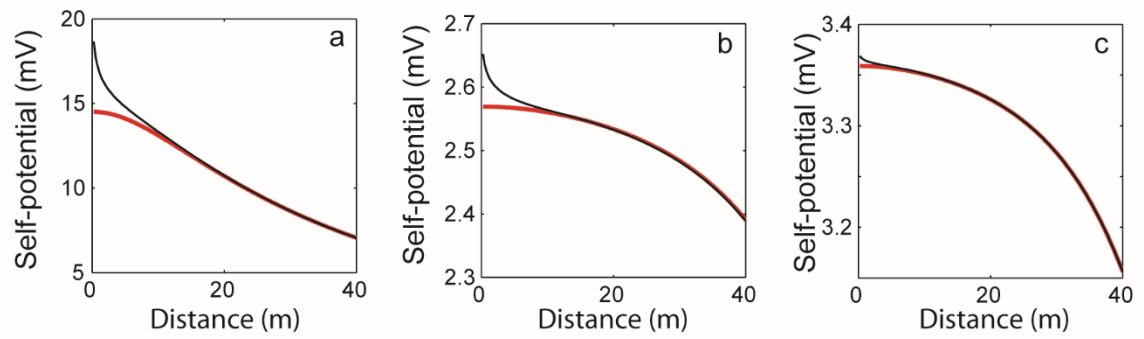


Figure 4.15. Modeled SP results along a profile away from the pumping well with an electrically conductive (black line) and insulating (red line) casing for the three different model scenarios; (a) homogeneous confined aquifer; (b) heterogeneous confined aquifer; and (c) heterogeneous unconfined aquifer.

4.12 Appendix A: Analog lowpass filtering for multichannel SP monitoring

The passive analog filtering scheme developed for the multi-channel SP array used in this study is shown in Figure A1. Each SP dipole is represented by an equivalent circuit consisting of a time-varying voltage source V_{si} galvanically connected across SP electrode e_i and reference electrode e_{ref} , having contact resistances R_{ci} and R_{cref} respectively. The two leads of each SP dipole were terminated at the recording site with a dedicated passive RC filter. The positive input of each filter was connected to an insulated lead, 20 – 50 m long, running from one of the SP electrodes. The negative input of each filter was connected to the ~200 m long insulated lead running from the reference electrode. The filtered outputs were sampled sequentially by a multiplexer connected to one differential channel of a data logger. For faster sampling of fewer channels (as in Figure 4.4), we could alternatively connect the filtered outputs directly to separate differential inputs on the 8-channel data logger.

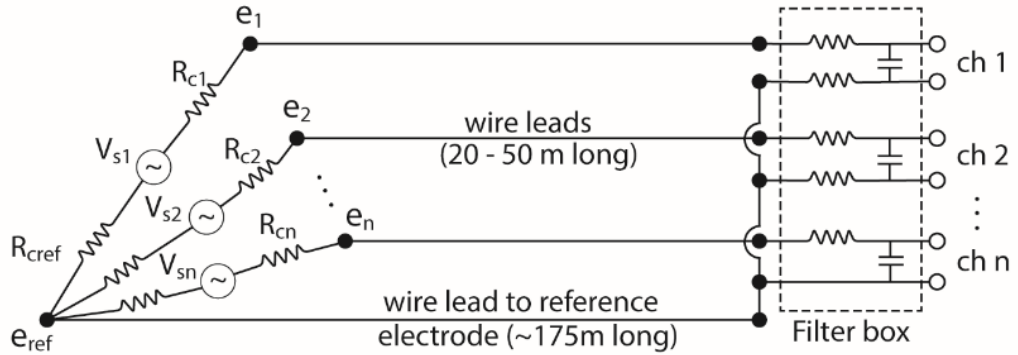


Figure A1. Schematic showing how an analog RC lowpass filter was applied across the leads of each SP dipole using a filter box at the recording site. The filtered outputs from 31 dipoles were sampled sequentially by a multiplexer connected to one differential channel of a data logger.

Figure A2 shows the equivalent circuit for a single SP dipole and filter. The contact resistances of the two electrodes are here replaced by their sum, the source impedance R_s , which represents the resistance between the two electrodes. Source impedances on the order of several $k\Omega$ or less should be achievable at most sites using non-polarizing electrodes set in wet bentonite clay at the base of short holes; R_s values were 3 – 5 $k\Omega$ for the field experiment at Springdale. Through basic circuit analysis it can be shown that the filter is a simple single pole RC lowpass rolling off gently at a rate of 6 dB/octave above a corner frequency F_c determined by the filter capacitor C_f and resistors R_f , as well as the source impedance R_s :

$$F_c = \frac{1}{2\pi(2R_f + R_s)C_f} . \quad (A4.1)$$

From equation (A4.1) it is clear that the filter resistors R_f should be larger than expected source impedances to prevent the corner frequency from being excessively sensitive to variations in R_s between one SP dipole and another. On the other hand, R_f should be much smaller than the input impedance R_i of any amplifier or data logger that is to be connected to the filtered output in order to prevent a voltage divided effect.

For SP monitoring, we decided to target a corner frequency near 1 Hz. With expected values of R_s being $\sim 3 k\Omega$ and typical amplifier/data logger input impedances expected to exceed 1 $M\Omega$, we settled on 10 $k\Omega$ resistors for R_f , and a 10 μF capacitor for C_f , thereby yielding a cut-off frequency of approximately 0.7 Hz. Variations in R_s from 0 to 8 $k\Omega$ would result in F_c varying from 0.8 to 0.6 Hz.

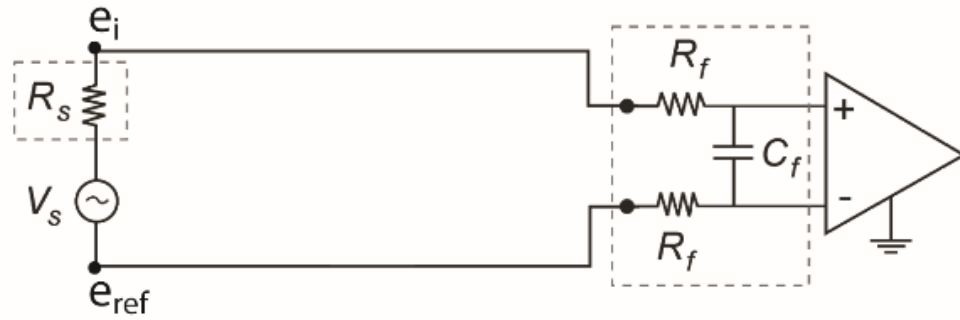


Figure A2. Schematic of the equivalent circuit for one SP dipole, and of the passive RC filter (within dashed box) that was applied across its wire leads at the input to the AM32B multiplexer that fed signals to a differential input amplifier on the CR1000 data logger. V_s and R_s represent the source voltage and resistance between a pair of monitoring (e_i) and reference (e_{ref}) electrodes.

Two points merit special mention. First, given that we were using a multiplexer, we could have used a single lowpass filter between the multiplexer and the data logger, rather than a dedicated filter on each dipole. The problem with that approach is that the filter would be exposed to abrupt changes in voltage (potentially 100's of mV of powerline noise) every time that the multiplexer switched from one channel to another. It would therefore be necessary to wait for the filter transient/step response to settle before the data logger acquired a sample.

Second, the data logger used for this monitoring experiment (a Campbell Scientific model CR1000) had an exceptionally high input impedance specification of $20\text{ G}\Omega$. We therefore could have increased R_f substantially without causing a significant voltage divider effect. Increasing R_f by a factor of 5 or 10 would have made the corner frequency practically immune to expected variations in R_s and would have allowed us to use more compact capacitors to obtain the same F_c . However, we ultimately settled on the lower resistance and higher capacitance (despite the inconvenience of bulky 10 μF capacitors measuring 2.5 cm

long by 0.8 cm in diameter), so that the filter box would remain suitable for use with other recording systems having lower input impedances.

Chapter 5

Surface self-potential patterns related to transmissive fracture trends during a water injection test⁴

5.1 Abstract

Variations in self-potential (SP) signals were recorded over an electrode array during a constant head injection test in a fractured bedrock aquifer. Water was injected into a 2.2 m interval isolated between two inflatable packers at 44 m depth in a vertical well. Negative SP responses were recorded on surface corresponding to the start of the injection period with strongest magnitudes recorded in electrodes nearest the well. SP response decreased in magnitude at electrodes further from the well. Deflation of the packer system resulted in a strong reversal in the SP signal. Anomalous SP patterns observed at surface at steady state were found to be aligned with dominant fracture strike orientations found within the test interval. Numerical modelling of fluid and current flow within a simplified fracture network showed that azimuthal patterns in SP are mainly controlled by transmissive fracture orientations. The strongest SP gradients occur parallel to hydraulic gradients associated with

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water flowing out of the transmissive fractures into the tighter matrix and other less permeable cross-cutting fractures. Sensitivity studies indicate that increasing fracture frequency near the well increases the SP magnitude and enhances the SP anomaly parallel to the transmissive set. Decreasing the length of the transmissive fractures leads to more fluid flow into the matrix and into cross-cutting fractures proximal to the well, resulting in a more circular and higher magnitude SP anomaly. Results from the field experiment and modelling provide evidence that surface-based SP monitoring during constant head injection tests has the ability to identify groundwater flow pathways within a fractured bedrock aquifer.

5.2 Introduction

Self-potential (SP) monitoring has an advantage over other geophysical methods in that measurements are directly sensitive to flow through porous and fractured rock, and can respond to weak fluid movements under an applied or natural hydraulic gradient (e.g. Sill 1983; Fournier, 1989). Electrical signals arise due to drag of excess electrical charge contained in the mobile portion of the electrical double layer at the rock-water interface (e.g., Davis et al. 1978; Ishido and Mizutani 1981; Revil et al. 1999; Jouniaux and Ishido 2012). Hydraulic gradients induced by pumping or injection transport mobile ions in the direction of fluid flow resulting in generation of convective currents. As a consequence, electrical potential differences occur causing return currents to flow by conduction (Revil and Leroy 2001).

Previous studies indicate that water injections have been responsible for generating detectable SP at different scales and varying injection rates. Crespy et al. (2008) observed a negative SP anomaly with maximum amplitude of $-6\ \mu\text{V}$ generated directly above a fluid

injection source located 15 cm below the surface of a laboratory sandbox. An overpressure of 17 Pa was attained by injecting 0.5 mL of fluid through a syringe by hammer stroke. The negative anomaly was immediately followed by a rapid reversal in the SP polarity, which they relate to a change in the polarity of the excess pore fluid pressure. Within a shallow bedrock aquifer, Hotzl and Merkler (1989) measured SP around a well while performing an injection/grouting experiment in the upper 10 m. By comparing their results with joint orientations they were able to show that SP measurements can be used to detect preferential flow within more permeable fracture sets. Deep geothermal experiments performed by both Ishido et al. (1983) and Ushijima et al. (1999) showed variations of the order of 3 to 5 mV during fluid flow at depths greater than 700 m. Ishido et al. (1983) showed that positive SP signals were correlated to pressure changes by opening the head valve of the geothermal well. Murakami et al. (2001) noted SP variations on the surface up to ~30 mV related to water injected at pressure of 3500-3800 kPa in a near vertical fracture zone at ~1600 m depth. SP magnitude decreased with increasing distance from the injection well. During a shut-in pressure test of an operation well in a geothermal reservoir, Yasukawa et al. (2005) reported SP anomalies of tens of millivolts caused by changes to the pressure field. During a 10-day SP monitoring experiment, Marquis et al. (2002) recorded amplitudes up to -5 mV on surface during the injection of 23,000 m³ of water into granite at 5 km depth. Murakami et al. (2001) and Marquis et al. (2002) concluded that the presence of a continuous electrically conductive casing was responsible for the electrical signal reaching the surface. Darnet et al. (2004) showed that the SP values recorded by Marquis et al. (2002) were mainly caused by electrokinetic processes as water circulated through fractures at depth, although electrothermal and electrochemical processes may also have

contributed to some extent. Troiano et al. (2017) modelled a complex water injection, incorporating coupling between fluid and heat transport, to evaluate SP anomalies associated with overpressure in the reservoir, as well as the influence of hydraulic stimulation on induced seismicity. Their results suggest that fluid injections at a depth of 5 km were the main source of SP measured at ground surface.

Extraction of water from an aquifer through pumping can also produce measureable SP signals at the ground surface (e.g. Bogoslovsky and Ogilvy 1973; Fagerlund and Heinson 2003; Revil et al. 2003; Rizzo et al. 2004; DesRoches and Butler 2016). Rizzo et al. (2004) showed that surface SP signals from both pumping and relaxation phases could be used to infer drawdown and estimate transmissivity of a porous aquifer. Using a sandbox experiment, Suski et al. (2004) were able to validate the approach in a controlled laboratory study.

Despite the breadth of modern SP studies (e.g. see Revil and Jardani 2013), relatively few have considered electrokinetic sources related to fluid flow through discrete fractures or fracture zones in a low permeability matrix (e.g. Bogoslovsky and Ogilvy 1973; Hotzl and Merkle 1989; Fagerlund and Heinson, 2003; Ishido and Pritchett 2003; Wishart et al. 2006; Suski et al. 2008; Robert et al. 2011; Nishi and Ishido 2012). Fagerlund and Heinson (2003) used SP signals, combined with borehole water levels, to approximate the drawdown pattern generated by pumping in a fractured rock aquifer. Their results show clear patterns in the drawdown contours, based on SP measurements, which were used to infer flow anisotropy related to the geometry of a fracture network. Wishart et al. (2006) showed that fluid flow through a network of highly transmissive fractures generated azimuthal variations in SP signals related to preferential flow directions and flow rates in a laboratory fractured

block model. Their analysis revealed that changes in fracture frequency or in hydraulic gradients driving flow were responsible for changes in the measured SP magnitudes. Using the same methodology, Wishart et al. (2008) investigated four fractured rock aquifers to demonstrate the linear relationship between SP magnitude and frequency of fractures along the same trend. Suski et al. (2008) used borehole SP measurements to detect and characterize highly transmissive fractures along a deep borehole in a fractured carbonate aquifer. Their analysis incorporated geochemical and borehole geophysical data to relate electrokinetic SP signals to flowing zones within the fractured network. Robert et al. (2011) acquired electrical resistivity tomograms and SP data to assess natural groundwater flow pathways through carbonate fracture zones. They found that negative SP magnitudes and low resistivity anomalies were associated with high yielding fracture zones confirmed through drilling and pumping. In recent numerical modelling using a 2D discrete-dual-porosity approach, Roubinet et al. (2016) simulated SP responses to flow within discrete fractures and across fracture walls into the less permeable matrix. Their results show that highly transmissive fractures well-connected to the hydraulic source (e.g. pumping well) tend to generate the strongest SP magnitudes. They also concluded that fluid flow between the rock matrix and fractures is necessary to generate SP signals associated with pumping such that modelling fluid flow through fractures alone, ignoring matrix fluid flow, will lead to SP responses that are orders of magnitude smaller.

In this study we evaluate SP signals recorded on the surface above a fractured and layered aquifer during a constant head injection test. Water was injected into a 2.2 m interval isolated by two inflatable packers set in a borehole 44 m below surface. Water injections generated clear negative SP anomalies as strong as -5 mV that were oriented

parallel to two major sub-vertical fracture sets. The SP anomalies can be interpreted as evidence of fluid flow from the pressure pulse away from the injection source along a hydraulically transmissive pathway. These observations are in agreement with previous results from Hotzl and Merkler (1989) in which SP anomaly azimuths matched fracture azimuths, as well as the numerical modelling results of Roubinet et al. (2016).

A 3D numerical model is used to investigate the transport of fluid into a simplified fractured network and bedrock matrix, and associated electrokinetic coupling during a constant head injection test. Although recent modelling by Roubinet et al. (2016) showed pumping-induced SP signals are sensitive to fluid flow through transmissive fractures, their model only accounts for flow in a 2D plane containing a fracture network with infinite vertical extent. They did not attempt to account for the limited vertical extent of a packed-off injection interval or for changes in resistivity with depth (e.g. presence of unsaturated bedrock or overburden, varying moisture content, metallic well casing). By incorporating the third dimension we are able to show how SP patterns vary with height above the injection zone and with vertical variations in resistivity. To our knowledge, this study is also the first to monitor transient surface SP signals during fluid injection at depth in discrete packer intervals over which fracture orientations have been well defined by televIEWER logging.

5.3 Description of the experiment

The SP monitoring experiment was carried out in June 2009 in the Black Brook Watershed (BBW), approximately 6 km north of the town of Grand Falls, in northwestern New Brunswick (Figure 5.1a). This watershed is part of an economically important agricultural region of the upper St. John River Valley where potatoes are the dominant crop.

Intensive agricultural practices have prompted several studies concerning impacts on surface and groundwater quality (e.g. Milburn 1996; Gallagher 1997; Valentin, 2002; Yang et al. 2010; Chow et al. 2011; DesRoches et al. 2014).

The injection well, BH10, was initially drilled, along with four others, as a part of an initiative to investigate the impact of nitrate application on the underlying groundwater supply (Figure 5.1b). The upper 6 m is lined with a steel casing extending through the upper 3 m of overburden sediments into the unsaturated bedrock. The remaining length of the well is open through both unsaturated and saturated bedrock to a depth of 70 m. The well has a diameter of 15.2 cm (6 inches) and intersected the water table 15 m below ground surface at the time of the experiment. Topography of the site is gradually sloping downward towards the northeast with <1 m of relief in the vicinity of the electrodes. However, the well itself was located on land elevated approximately 2 m above the nearest row of electrodes ~5 m away.

Geology at the site consists of fine-grained Ordovician and Silurian calcareous and argillaceous turbidite sequences deposited within the Matapédia Basin (Wilson et al, 2004). Regionally, the bedrock is weakly deformed forming broad NE-trending open to isoclinal folds. Within each of the wells, bedding is shallowly dipping (~31 degrees) in a southeasterly direction (DesRoches et al. 2014). Overburden sediments blanket the area with ~0.5 to 3 m of loamy lodgement till and minor amounts of ablation till, silt, sand, gravel and rubble (Langmaid et al. 1980; Rampton et al. 1984).

Groundwater flow in the aquifer occurs within the fracture network (see Figure 2 in DesRoches et al. 2014 or Figure 3.2 in Chapter 3). Within BH10, fracture depths, dips and azimuths, and apparent apertures were determined using continuous, oriented images from

optical and acoustic televiewer logs. Steeply dipping open and calcite-filled fractures and moderately dipping bedding plane fractures were identified with a geometric mean fracture spacing of 0.26 meters. The steeply dipping fractures were sub-divided into two distinct sets striking NNE and WNW, while the bedding plane fractures strike NE with a mean southeasterly dip of 31°. Calcite filled fractures are less pervasive with a NW strike. Because these fractures are filled, they are unlikely to contribute significantly to groundwater flow.

Hydraulic conductivities in BH10 were determined through 22 constant head injection tests performed using a set of inflatable packers to isolate 2.2 m intervals over the length of the well (DesRoches et al. 2014). Hydraulic conductivity ranged from 1.3×10^{-3} to 7.2×10^{-9} m/s with a geometric mean of 5.9×10^{-6} m/s. DesRoches et al. (2014) observed a correlation between the frequency of fractures showing oxidation and weathering and the hydraulic conductivity profile in the borehole. Fractures with significant weathering and oxidation were inferred to be most important to the groundwater flow system. Using the fluid saturation method, bedrock porosities were calculated based on 89 hand samples from nearby outcrops and show a mean value of just 1%.

Self-potentials measured on the surface are influenced by the distribution of electrical resistivity of the bedrock and overlying units. Overburden and bedrock resistivity at the site were determined through four dipole-dipole resistivity imaging profiles that passed through the area of the SP electrode array (Figure 5.3a - d). With an array of 24 electrodes, spaced 2 m apart, the survey reached a depth of approximately 10 m, with a sharp electrical contact at depths of ~1-4 m separating the overburden from the unsaturated bedrock. The resistivity model consists of overburden material with a resistivity of approximately 300 Ωm (0.0033 S/m) on the western side of the profiles increasing to approximately 800 Ωm

(0.0013 S/m) towards the east. The underlying unsaturated bedrock unit has a resistivity of 1500-2500 Ωm (0.00067 – 0.0004 S/m). Overburden resistivity was also consistent with values determined from an electromagnetic survey (using Geonics EM31) at the site (not presented here). The resistivity of the deeper bedrock below the water table was measured by logging with a normal resistivity tool in BH10. The vertical resistivity profile (Figure 5.3e) displays variability with a wavelength of 10-12 m, ranging from 1600 to 2800 Ωm (0.00063 – 0.00038 S/m) with an average of approximately 2100 Ωm (0.00048 S/m). The variability is thought to reflect subtle lithological changes in the bedrock, which are also evident, to a lesser extent, in the gamma ray log (Figure 5.3f). A fluid conductivity probe recorded a mean value of approximately 0.05 S/m (20 Ωm) in BH10.

The SP data were continuously recorded on the surface over the duration of the injection test at 3 second intervals using a Campbell Scientific AM32B multiplexer and a CR1000 data logger (input impedance of 20 G Ω) connected an array of 32 non-polarizing Pb-PbCl₂ electrodes (Petiau, 2000). Noise was reduced by operating the data logger in the sampling mode it provides to cancel stable 60 Hz powerline noise. The electrodes were arranged immediately northeast of well BH10 in a regular grid (Figure 5.1 and Figure 5.2) and installed in shallow holes (~20 cm deep), seated in bentonite mud, and covered with soil for the duration of the test to promote uniform surface contact, lower contact resistance, and minimize temperature and moisture variations. Soil temperature recorded during the test, near electrode e14 in the center of the array, was uniform at approximately 13°C. A mean contact impedance of ~1.5 k Ω was measured for all electrodes in the array. The reference electrode was positioned 250 m away from the well – sufficiently distant that any injection-related SP was expected to be negligible. Two additional electrodes were

deployed ~20 m south and west of the reference electrode in order provide two “remote dipole” SP measurements. Data from the remote dipoles were analyzed to identify background telluric fields which could overprint SP responses in the array. SP signals were reduced by a mean temporal reference value calculated over a 2-minute window 10 minutes prior to the start of injection, which corresponds to a stable period shortly after inflating the packer system. Rizzo et al. (2004) used the same approach to reduce their time-series SP data to isolate the electrokinetic contribution related to pumping. High-frequency noise of approximately ± 1.5 mV recorded during the test was smoothed using a second order polynomial Savitzky-Golay filter with a 2 minute moving window.

While SP data were being recorded, water was injected into a 2.2 m interval isolated between 43.2 and 45.4 m depth using a pair of inflatable packers. The injection was initiated several minutes after the packers were inflated, allowing time for the test interval to stabilize. Pressure in the test interval was monitored using an electronic pressure transducer. The initial injection resulted in an increase in hydraulic pressure of 79.0 kPa for the first 4 minutes, which stabilized at 76.8 kPa for the remainder of the test. The duration of the injection test was 24 mins, ending once the pressure and in-flow rate in the test interval showed no change with time. A flow rate of $0.178 \text{ Kg}\cdot\text{s}^{-1}$ ($1.78 \times 10^{-4} \text{ m}^3\cdot\text{s}^{-1}$) was measured following stabilization. Within the packer interval, three bedding-parallel fractures and two steeply dipping fractures show evidence of oxidation weathering and apparent apertures ranging between 3 – 7 mm, determined from borehole televiewer images (DesRoches et al. 2014). The steeply dipping fractures (from two populations, both exhibiting mean dips of 75°) are oriented NNE and WNW while the bedding plane fractures strike NE and dip $\sim 31^\circ$ to the SE. These fracture orientations are consistent with the three populations of

open fractures measured throughout the rest of the well and nearby wells. Constant head test results yielded a hydraulic conductivity of 7.2×10^{-6} m/s for the interval.

Figure 5.4 shows transient SP data recorded by six electrodes at distances of 6 to 41 m from the well. The SP data can be separated into three phases that correspond to the different stages of the constant head injection test. The first phase occurs following the inflation of the packers. During this phase, SP data was measured for a 10-minute period to produce a base-level response prior to the injection test. Prior to inflating the packers, vertical flow in the well circulates water from fractured zones of higher pressure to lower pressure (Quinn et al. 2012). Once the packers are inflated, the test interval becomes isolated as a closed system in which the pressure in the interval stabilizes prior to the start of the injection.

The second phase occurs following the initiation of fluid injection into the packer interval (time = 0 min). Fluid injection occurred until the pressure and in-flow rate in the test interval had stabilized. This phase resulted in a negative SP signal recorded on the surface. Data from Figure 5.4 illustrate a well-defined SP anomaly with a maximum amplitude of -5 mV at the electrode closest to the well. This anomaly is related to the start of fluid injection. The most obvious SP response is measured during the initial 6 minutes of the injection phase, and levels-off for the remainder of the test. Although the transient pressure curve was not digitally recorded during the experiment, the pressure was monitored during the experiment to know when steady state was achieved. It was observed that the initial pulse of water into the packer interval resulted in the largest pressure generated during the test. Although injection rates are low (0.178 kg/s), SP magnitudes

generated in the packer interval, centered at a depth of 44.3 m were strong enough to be measured on the surface.

Following the termination of the injection phase and deflation of the packers (at time = 24 min) we observed an abrupt reversal in SP polarity (+30 mV) in the electrode closest to the well. Thereafter, oscillations in SP were observed at electrodes located closest to the well (electrodes e1 through e18). Data recorded at the remaining electrodes at greater distances did not show the same reversal in polarity. The magnitude of the polarity reversal decreased with increasing distance from the well, ranging from +30 mV at e3 to +4 mV at e15, respectively positioned 6 m and 24 m from the well. The SP anomaly appears to result from the rapid release of pressure induced in the packer interval relative to the overlying and underlying columns of water. Although not recorded, while the packer interval is under pressure, there will be a pressure gradient between the closed packer interval and the column of water above and below the packers. Once the packer seal is released, the pressure differences in the well are expected to quickly stabilize and return to hydrostatic pressure. The response of higher pressures generated in the fractured bedrock, as a result of the injection, are expected to drive fluids back into the well since the fluid pressures in the fractures are well connected to the injection interval. The polarity reversal we observed is also consistent with results of Crespy et al. (2008) who measured SP during an injection test in a sand box experiment. They attribute the SP reversal to a change in the polarity of excess fluid pressure in the capillary with respect to the hydrostatic pressure in the sand at the injection depth. We also considered the possibility of fluid injection pressures resulting in fracture dilation, but any such effect is expected to be negligible due to the relatively small applied pressure (National Research Council 1996). During the recovery phase, the SP

response began to return towards pre-injection conditions, but recovery remained incomplete after 27 minutes, when the next injection was initiated.

Using the time-dependent data from the electrodes, a minimum curvature grid was calculated to show the difference between the SP responses averaged for 1 minute towards the end of the injection phase (23 minutes after injection started) and those measured prior to injection (Figure 5.5). Changes in SP between the ambient and injection phases are greatest near the well and tend to decrease with increasing distance from the well. However, the SP anomaly around the well is asymmetric, suggestive of hydraulic anisotropy and/or heterogeneity in the underlying aquifer (although electrical anisotropy and heterogeneity could also play a part). The SP anomaly exhibits two negative lobes trending roughly 010° and 065° degrees away from the well in agreement with the mean fracture strike of one steeply dipping fracture set and the moderately dipping bedding fractures measured in the well by televiwer logging (Figure 5.5; DesRoches et al. 2014). All fracture sets within the aquifer tend to be continuous and through-going, which is consistent with a non-stratabound fracture system (e.g. Odling et al. 1999) working to enhance vertical connectivity in the aquifer, between bedding plane and steeply dipping fractures at different depths. Empirically, it would appear that the surface SP anomaly exhibits asymmetry consistent with expected fracture-related azimuthal asymmetry in groundwater flow.

5.4 Numerical modelling

Numerical modelling following the approach of Sill (1983) was carried out using COMSOL Multiphysics (COMSOL, 2014) to better understand the SP distribution expected in a simple fractured rock aquifer during a constant head injection test. Several

sensitivity analyses are performed to explore (a) increased frequency of hydraulically conductive fractures near the well, (b) decreased length of hydraulically conductive fractures, and (c) increased hydraulic conductivity of the matrix rock. The conduction current and SP patterns are plotted to illustrate conditions in which these fracture scenarios are expected to generate electrical SP anomalies that align with fracture geometry.

5.4.1 SP theory

Self-potential (SP) related to fluid flow in a porous or fractured media is caused by coupling between flows of water and electrical charge. The coupling arises due to drag of excess electrical charge contained in the mobile portion of the electrical double layer at the rock-water interface (e.g., Davis et al. 1978; Ishido and Mizutani 1981; Revil et al. 1999; Jouniaux and Ishido 2012). For typical freshwater aquifers, the solid surfaces hold an excess negative charge, which is balanced by an excess positive charge in the pore fluid adjacent to the pore wall (e.g., Revil et al. 2003). Some of the positively charged ions in this electrical double layer are free to move with the pore fluid resulting in the establishment of streaming currents, leading to the development of a potential gradient driving conduction currents in the opposite direction (Revil and Leroy 2001). Fluid and electrical flows are described by transport equations operating at the scale of a representative elementary volume (e.g. Titov et al. 2005; Straface et al. 2010):

$$\mathbf{j} = -\sigma \nabla \varphi - L \nabla h \quad (5.1)$$

$$\mathbf{u} = -L \nabla \varphi - K \nabla h \quad (5.2)$$

where $\mathbf{u}$ is the volumetric fluid flux ($\text{m}\cdot\text{s}^{-1}$), and $\mathbf{j}$ is the total electrical current density ($\text{A}\cdot\text{m}^{-2}$) represented by the sum of conduction currents given by Ohm's Law $\mathbf{j}_c = -\sigma \nabla \varphi$ and

streaming or convection currents $\mathbf{j}_s = -L\nabla h$ representing the electrokinetic contribution. Here, σ is the bulk electrical conductivity ($\text{S}\cdot\text{m}^{-1}$) of the rock, φ is the electric potential (V), and L is the electrokinetic current coupling coefficient ($\text{A}\cdot\text{m}^{-2}$). Hydraulic head (m) is given as $h = \psi + z$, where $\psi = p/\rho g$ is the pressure head (m), p is the pore fluid pressure (Pa), ρ is the fluid density ($\text{kg}\cdot\text{m}^{-3}$), g is gravitational acceleration ($\text{m}\cdot\text{s}^{-2}$), and z is the elevation head (m).

The transport eq.(5.1) and (5.2) are linked through the electrokinetic current coupling coefficient (Revil et al. 2003; Darnet et al. 2003; Straface et al. 2010):

$$L = -\sigma C \quad (5.3)$$

where C is the electrokinetic voltage coupling coefficient (V m^{-1}) representing the voltage that would develop across a sample of the material in response to fluid flow driven by a unit change in hydraulic head across the sample.

Eq.(5.2) can be simplified under the assumption that the electro-osmotic contribution to fluid flow $-L\nabla\varphi$ is several orders of magnitude smaller than that produced by the hydraulic gradient associated with fluid injection (Sill 1983; Ishido 1989). This assumption leads to the classical Darcy's equation:

$$\mathbf{u} = -K\nabla h \quad (5.4)$$

where K is hydraulic conductivity (m s^{-1}). At steady-state, assuming fluid is incompressible, eq.(5.4) can be combined with the continuity equation ($\nabla \cdot \mathbf{u} = 0$) to form the governing equation for flow in porous media:

$$\mathbf{Q} = -\nabla \cdot (K\nabla h) \quad (5.5)$$

where $\mathbf{Q}$ is the hydraulic source or sink (s^{-1}). Using eq.(5.5), hydraulic head distribution for the bedrock matrix can be solved by specifying hydraulic conductivities and appropriate boundary conditions (i.e. hydraulic head or flux).

By representing fracture planes as 2D interior boundaries, rather than thin 3D domains, fluid flow along a fracture can be simulated using a modified form of eq.(5.4) combined with the continuity equation to obtain (e.g. Blessent et al. 2009):

$$d_f \mathbf{Q} = -\nabla_T \cdot (K d_f \nabla_T h) \quad (5.6)$$

where d_f is the fracture aperture, and ∇_T denotes the gradient operator restricted to the fractures tangential plane. Representing fractures as 2D interior boundaries is computationally efficient, compared to modelling each fracture as very long and narrow domains requiring an extremely dense mesh. This modelling approach is used in COMSOL Multiphysics subsurface flow module (COMSOL v5.0) to account for fluid transport within the matrix and fracture planes.

For the electrical problem we appeal to the continuity equation for steady currents ($\nabla \cdot \mathbf{j} = 0$) which can be used to identify sources (non-zero divergence) of conduction currents arising from electrokinetic coupling leading to the governing equation solved in COMSOL:

$$\nabla \cdot \mathbf{j}_c = \nabla \cdot (\sigma \nabla \varphi) = -\nabla \cdot (L \nabla h). \quad (5.7)$$

By substituting eq.(5.3) for the current coupling coefficient L , the right hand side of eq.(5.7) can be expanded, as shown by DesRoches and Butler (2016) (following Sill 1983) to reveal four distinct sources of conduction currents:

$$\nabla \cdot \mathbf{j}_c = \sigma(\nabla \mathcal{C} \cdot \nabla h) + \mathcal{C}(\nabla \sigma \cdot \nabla h) - \left(\frac{\sigma \mathcal{C}}{K}\right)(\nabla K \cdot \nabla h) - \left(\frac{\sigma \mathcal{C}}{K}\right)(\nabla \cdot \mathbf{u}). \quad (5.8)$$

The right hand side of eq.(5.8) shows that SP sources can be generated by flow (hydraulic gradients) directed perpendicular to boundaries (or parallel to gradients) in electrokinetic current coupling coefficient L , electrical conductivity σ , or hydraulic conductivity K or arising from any divergence in fluid flow (e.g., injection or pumping wells).

5.4.2 Model description

SP distribution was modeled based on a rigorous modelling approach proposed by Sill (1983). Eq.(5.5) and (5.6) were first solved to determine hydraulic head distribution through the matrix and fracture network. The distribution of the hydraulic head in steady state was determined by the hydraulic conductivity of the domain units (i.e. fractures and matrix), the hydraulic boundary conditions and the injection flux which are all discussed below. Given the hydraulic head distribution, sources of conduction currents on the right-hand side of eq.(5.8) were calculated using a uniform voltage coupling coefficient ($\mathcal{C}$), and electrical conductivity (σ) for the water-saturated matrix, and specific hydraulic conductivities (K) for matrix and fracture sets, in addition to the source term ($\mathbf{Q}$) resulting in divergence in fluid flow. Lastly, the SP distribution (φ) through the model was found by solving eq.(5.8).

The model in 3-dimensions consists of an x and y lengths of 200 m and a depth extent of 100 m centered around an injection well. The model is subdivided into three horizontally layered domains consisting of two bedrock layers and a thin overburden layer. An overburden layer extends from surface to 3 m depth, and the upper 12 m of the

bedrock layer is unsaturated. The water table occurs at 15 m depth and the remaining thickness of the model is fully saturated. The bedrock layers are cross-cut by two vertical fracture sets. Set 1 has an azimuth of -43.3° (316.7°) and Set 2 has an angle of 36.6° , intersecting with an inner angle of $\sim 80^\circ$. Both fracture sets terminate at the external boundary of the model. Figure 5.6 shows a horizontal slice and vertical section through a smaller subdomain of the 3-dimensional model with an x and y length of 50 m and depth extent of 70 m. This subdomain is used for presenting results.

A vertical well, representing BH10, is incorporated in the model with a 6-inch diameter and a total depth of 70 m. The upper 6 m of the well through the overburden and upper unsaturated layer is lined with steel casing. The packer interval is isolated between 43.2 and 45.4 m depth and represented as a highly transmissive domain with a hydraulic conductivity of $1 \text{ m}\cdot\text{s}^{-1}$. In the middle of the packer interval, the injection is represented as a point source volumetric flux condition applied at a constant injection rate (Q) of $1.78 \times 10^{-4} \text{ m}^3\cdot\text{s}^{-1}$, equal to the injection rate used in the field to achieve steady state. This approach does not require prior allocation of total discharge into fractures and matrix, as that is determined as part of the model solution (e.g Neville and Tonkin 2004). The water table surface and the base of the model domain were defined as no-flow boundary conditions. The upper and lower bounds of the packer test interval in the well representing the packer seal were also set as no-flow. To isolate the influence of the injection, the initial hydraulic head conditions throughout the entire saturated domain were set to $h_0 = 0 \text{ m}$. Gravitational forces influencing groundwater flow through the domain were neglected. The model ground surface is flat and arbitrarily assigned a reference elevation of 0 m. The vertical boundaries located on all sides of the domain were assigned a constant hydraulic

head of 0 m ($h_0 = 0$ m), such that under ambient conditions no hydraulic gradients exist in the domain.

For the electrical problem, an insulating boundary condition was assigned on all boundaries of the model domain. A SP reference point is located on the ground surface 100 m from the well ($\phi_0 = 0$). The distance to the model boundaries is considered large enough to eliminate boundary influences on hydraulic and electrical current flow. To show this, base case modelling was repeated within a large domain and results were negligibly different.

Model layers and fracture sets were each assigned uniform hydraulic conductivities. The saturated bedrock matrix was represented as an equivalent porous medium with a hydraulic conductivity of $1 \times 10^{-7} \text{ m} \cdot \text{s}^{-1}$. This value was approximated based on the geometric mean of the borehole packer testing, representing both fractures and matrix within BH10. Bedrock matrix was then overprinted by two sets of widely spaced hydraulically conductive fracture planes. Fractures were represented as thin vertical interior boundaries within the equivalent porous media. The fracture planes were considered to have semi-infinite extent, truncating at the upper and lower bounds of the bedrock and outer boundaries of the model domain. Hydraulic conductivity of the fracture sets was assigned to be several orders of magnitude greater than that of the matrix. Using the cubic law derived by Snow (1968) with inputs of model fracture spacing of approximately 10 m and aperture of 3 mm, we approximated an equivalent hydraulic conductivity for fracture Set 1 to be $5 \times 10^{-3} \text{ m/s}$. For the purpose of modelling anisotropic flow along the transmissive fracture set, hydraulic conductivity of fracture Set 2 was lowered by two orders of magnitude to $5 \times 10^{-5} \text{ m/s}$.

Electrical resistivities of the three model layers were based on the resistivity sections and the vertical resistivity profile shown in Figure 5.2. The assigned resistivity of the overburden in the model is 600 Ωm , which is considered to be representative of the variable uppermost layer in the sections, with the assigned resistivity of the unsaturated bedrock increasing to 2500 Ωm . Based on resistivity measurements taken from BH10, an average resistivity of 2100 Ωm is assigned to the saturated bedrock. A uniform resistivity distribution is assumed for each of the model layers. Therefore, within the saturated bedrock the matrix and the fractures are assigned the same value. The metallic casing lining the well in the upper 6 m of the model is assigned an electrical conductivity of 100,000 $\text{S}\cdot\text{m}^{-1}$.

The intrinsic electrokinetic voltage coupling coefficient (C) assigned to the saturated portion of the model was determined by lab measurements carried out on four rock samples acquired from BH10, following procedures outlined in Morgan et al. (1989). The rock samples were acquired as chip samples during drilling and were air-lifted to surface from bedrock at approximately the depth of the packer interval. We determined the voltage coupling coefficient from the ratio between the electrical potential difference measured at two ends of a cylindrical sample and an applied hydraulic head difference across the sample. Lab measurements provided an average value of $-7.8 \text{ mV}\cdot\text{m}^{-1}$ ($-8 \times 10^{-7} \text{ V}\cdot\text{Pa}^{-1}$). Given knowledge of electrical conductivity of the fluid in BH10 at the time of the packer testing ($\sigma_f = 0.05 \text{ S}\cdot\text{m}^{-1}$) the voltage coupling coefficient can also be approximated using two empirical correlations: $\log_{10} C = -0.921 - 1.091 \log_{10} \sigma_f$ where C is given in $\text{mV}\cdot\text{m}^{-1}$ (Revil et al. 2003), and $C = -1.2 \times 10^{-8} \sigma_f^{-1}$, with C given in $\text{V}\cdot\text{Pa}^{-1}$ (Jouniaux and Ishido 2012). Using both approaches with the observed borehole fluid conductivity yields

voltage coupling coefficients of -3.2 and $-2.4 \text{ mV}\cdot\text{m}^{-1}$ ($-3.3 \times 10^{-7} \text{ V}\cdot\text{Pa}^{-1}$ and $-2.4 \times 10^{-7} \text{ V}\cdot\text{Pa}^{-1}$) which are two to three times smaller than values determined in the lab. The electrokinetic voltage coupling coefficient from the lab was assumed to better reflect in situ rock-fluid interactions and was used in the model.

The finite element mesh was refined in locations where hydraulic and SP gradients were expected to be high (e.g. near the injection well and steel casing). The mesh was discretized with a minimum element dimension of 0.025 m near the injection well, a maximum dimension of 38 m at the outer edge, and a maximum growth rate of 1.5 , with the final mesh comprising 1490220 tetrahedral, and 319231 triangular elements.

5.4.3 Model results

Using the boundary conditions described above, we calculated a steady-state solution for hydraulic head distribution through the 3-dimensional fracture network and bedrock matrix. The distribution of hydraulic head derived from the injection within the packer interval is shown in Figure 5.7(a – d) as a series of horizontal sections through the model at different depths. At all depths, the shape of the hydraulic head distribution is elongate parallel to fracture Set 1, which has a higher hydraulic conductivity compared to fracture Set 2. Fluid flow within Set 1 results in a strong hydraulic gradient generated perpendicular to flow direction associated with water flowing out of the transmissive fractures into the tighter matrix and cross-cutting fractures.

Using the hydraulic head distribution, sources of conduction currents were calculated from hydraulic head gradients using the specified distribution of relevant physical properties (i.e., electrokinetic coupling coefficient C , electrical conductivity σ , and hydraulic conductivity K). Resulting conduction currents occur at the injection source due to

divergence in fluid flow and where flow is perpendicular to any hydraulic conductivity K contrasts, such as between the borehole wall, fractures and matrix. Within the saturated bedrock the electrokinetic coupling coefficient ($\mathcal{C}$), and electrical conductivity (σ) are assumed to be uniform and therefore do not generate conduction currents.

In Figure 5.8(a – b), conduction currents are distributed around the packer interval due to fluid injection into the fracture network and matrix. The conduction currents are not directed purely radially towards from the injection point as would be expected in homogeneous, isotropic media but are instead affected by the orientations of the fractures over which the source terms are distributed. The strongest conduction currents are generated near the injection source within the rock matrix with magnitudes dropping across fractures from Set 1 (Figure 5.8a). In the vertical section, conduction currents are similarly strongest at the injection interval, and elongated parallel to the fracture planes (Figure 5.8b). Weaker conduction currents cross the boundary into the overburden layer and around the metallic well casing. Conduction currents generally flow towards the injection point, with the exception of the overburden layer where current is forced to flow horizontally at the surface and is channeled down the conductive well casing.

The conduction current sources developed in response to fluid flow were used to solve for SP distribution (ϕ) through the model accounting for the electrical resistivity distribution associated with the three layers and the metallic well casing. In Figure 5.9(a – f) the modeled SP is shown at different depths from ground surface down to the level of the packer interval. Similar to the hydraulic head distribution, the SP response at all depths is elongate parallel to the orientation of fracture Set 1; the strongest SP gradients are parallel to the hydraulic gradients associated with water flowing out of the conductive fractures (Set

1) into the tighter matrix and cross-cutting fractures (Set 2). Within the horizontal plane through the packer interval (Figure 5.9f), the SP magnitude rapidly diminishes perpendicular to fracture Set 1, from -244 mV at the packer interval to -60 mV over a distance of 5 m. The SP contribution from fracture Set 2 is minimal despite the hydraulic conductivity being two orders of magnitude greater than the matrix as the bulk of the flow is accounted for in the more transmissive fractures (Set 1). Although not explicitly modelled, SP effects of fluid flow within the fractures themselves are considered to be negligible for two reasons. First, the high permeability of the fractures results in a significantly lower hydraulic gradient, compared to the flow from the fracture into the matrix. Second, fractures have a much smaller grain surface area to volume ratio compared to porous matrix, resulting in reduced electrokinetic coupling at the rock-water interface. This is consistent with model findings from Roubinet et al. (2016) who showed that pump-induced fluid flow strictly occurring within a transmissive fracture network, in the absence of a permeable rock matrix, resulted in a SP response approximately 3 orders of magnitude smaller.

The magnitude and elongation of the SP anomaly shown in Figures 5.9(a - f) decreased as a consequence of both increasing distance from the injection source and presence of resistivity layering in the model. By making the electrical conductivity of the 3 m thick overburden layer equal to the unsaturated layer ($2500 \Omega\text{m}$) the magnitude of SP on the surface at the well increased from 12.2 mV to 16.3 mV. For a given current density associated with the injection source, the presence of a more electrically conductive overburden layer results in smaller voltage drops (SP) observed at surface. Likewise, modelling by Titov et al. (2005) showed that SP magnitudes decreased roughly linearly with

the logarithm of the conductance (electrical conductivity-thickness product) of an overlying unit.

The presence of a metallic well casing in the upper part of the model was expected to shunt electrical currents and thereby increase SP response near the well (Marquis et al. 2002; Titov et al. 2005; Darnet et al. 2006). However, since the well casing penetrates only the upper 6 m of the model, that is within the unsaturated part of the aquifer, its influence on the SP calculated over the surface is less than 0.2 mV at the well casing. To assess this, we also modeled the SP distribution without the metallic well casing.

5.5 Discussion

5.5.1 Influence of fracture heterogeneity

We now alter the model to investigate the sensitivity of SP to three different fracture heterogeneities, while holding the injection rate constant. Figure 5.10a shows the base case simulation from Figure 5.9(f) using an expanded colour scale, while Figure 5.10(b – d) show the SP field in the same horizontal slice at the depth of the packer interval for three different fracture patterns. The first sensitivity involves doubling the number of transmissive fractures (from 3 to 6 over a 10 m interval) around the injection well (Figure 5.10b). Simulation results show that increasing the fracture density for the more transmissive set (Set 1) further elongates the SP pattern in response to hydraulic gradients generated perpendicular to the fractures, compared to result presented in Figure 5.10a. These fractures tend to be hydraulically better connected to the packer interval compared to the model results with lower density of fractures.

The second sensitivity analysis (Figure 5.10c) alters the base case model by changing the length of individual transmissive fractures within Set 1, making them considerably shorter and discontinuous, leading to shorter transmissive pathways for water to flow during injection. Simulation results show a broader SP distribution with an irregular shape attributed to the location and orientation of fractures that are hydraulically connected to the packer interval. Having shorter transmissive fractures will force more water into the matrix and into the cross-cutting fractures near the well. As shown in Figure 5.10c this leads to a less elongated SP anomaly. Two-dimensional SP modelling by Roubinet et al. (2016) showed similar results, where the highest SP magnitudes were generated along shorter fracture planes that were hydraulically connected to the well. They concluded that pumping-induced flow in fractured rock produces SP primarily in response to fluid flowing from the matrix into fractures, rather than in response to fluid flowing along the transmissive fractures themselves.

The final sensitivity analysis was performed by setting the hydraulic conductivity of both fracture sets equal in order to understand if the SP anomaly in Figure 5.9 is related to water flowing perpendicular to the hydraulically conductive fracture set (Set 1) into the matrix or along the other less conductive fracture set (Set 2). SP modelling results show a near circular SP anomaly centered on the packer interval (Figure 5.10c). Strong hydraulic and electrical gradients that were previously modelled perpendicular to fracture Set 1 are diminished as fluid now preferentially flows away from the packer interval along both fracture orientations. The SP anomaly is reduced in magnitude and spatial extent, and has become more equant compared to previous model results. This is as expected owing to the increase in effective hydraulic conductivity near the packer interval and fluid flowing along

two sets of hydraulically conductive fractures (Set 1 and 2). Increasing the hydraulic conductivity of the matrix instead by 2 orders of magnitude, from 10^{-7} m/s to 10^{-5} m/s, also results in a similar decrease in the hydraulic and SP gradients that are generated perpendicular to fracture Set 1.

The initial model and sensitivity analyses performed in this study demonstrate that fluid injected into an isolated packer interval in a fractured bedrock aquifer can generate a negative SP anomaly that parallels the trend of a highly transmissive fracture set. These modelling results help to inform our interpretation of SP measured during the constant head packer testing in the Black Brook Watershed. The negative SP anomaly simulated as radiating from the well is consistent with the elongated SP anomalies recorded on the surface to the north of BH10, shown in Figure 5.4 and Figure 5.5. Modelling has shown that fluid flowing preferentially along zones of tightly spaced fractures produces strong SP gradients perpendicular to the strike, and an elongated SP anomaly parallel to the strike of the transmissive fracture zone. Televiewer logging in BH10, as well as nearby wells, revealed several fracture zones with tight spacing and uniform strike orientations (DesRoches et al. 2014). These conditions are expected to produce SP anomalies that are elongated parallel to their strike orientation.

Modelling results also show that decreasing the length of the transmissive fracture set leads to more fluid flow into the matrix and into cross-cutting fractures, resulting in a broader and higher magnitude SP anomaly. Similar conclusions have been reached by Roubinet et al. (2016). Fracture mapping on large outcrop exposures in the Black Brook watershed (DesRoches et al. 2014) revealed that, although short fractures are present, most fractures are long and interpreted to be through-going. The nature of through-going

fracture sets present in Black Brook watershed would favour the establishment of SP anomalies elongated parallel to the fracture trend.

It is recognized that in order to produce a SP anomaly that parallels a fracture set there must exist a sufficient hydraulic conductivity contrast between the fractures and the matrix. In the case of our model, a contrast of three and a half orders of magnitude was sufficient to produce strong hydraulic gradients along the boundary of the highly transmissive fracture planes. Decreasing this contrast with the matrix results in the production of a more circular SP anomaly that is more consistent with expectations of a homogenous, isotropic medium. The hydraulic testing in BH10 (DesRoches et al. 2014) confirmed that the bedrock matrix is tight, with intervals of unfractured bedrock often measuring below the detection limit (i.e. $\sim < 10^{-10}$ m/s). This demonstrates that the hydraulic conductivity contrast between fractured and unfractured rock should be sufficient to generate azimuthal SP anomalies around wells in the Black Brook Watershed.

In considering the physical nature of SP generated by fracture flow, we note that, intuitively, transmissive fractures within a relatively impermeable matrix (i.e. $K_f \gg K_m$) can be expected to act in a similar manner to an extension of the well. Given their very large hydraulic conductivity contrast with the matrix, and their relatively low surface area to volume ratio, one could consider that the injection source term is effectively distributed along the transmissive fractures. Smaller SP effects are also expected to be produced from small hydraulic gradients along the fractures themselves. However, as mentioned above, modelling by Roubinet et al. (2016) led them to conclude that SP signals generated by hydraulic gradients within the fractures alone are orders of magnitude smaller than those produced when flow in the matrix is considered.

The modelling in this study assumed that the electrical conductivity of the fractures was no different from that of the matrix. In effect, the fractures were ‘invisible’ to the conduction currents. In some fractured rock aquifers, including the Black Brook Watershed, electrical conductivity contrasts between fractures and matrix may exist where fracture conductivity can be considerably higher, in particular where high fluid conductivities exist. Roubinet et al. (2016) assigned the fractures an electrical conductivity two orders of magnitude higher than the matrix, but did not elaborate on the sensitivity of conduction current paths or SP patterns to this contrast. Intuitively, elevated electrical conductivity along a fracture should limit the SP field generated by flow *within* the fracture more so than that generated by water forced out of the fracture into the matrix. Furthermore, the influence of elevated fracture electrical conductivities on conduction currents would be limited by their small cross-sectional area (limiting their conductance). However, further investigations are recommended to investigate what influence elevated electrical conductivities along fractures might have on conduction currents and resulting SP anomalies compared to the predictions obtained here using bulk (i.e. equivalent porous medium) isotropic electrical conductivity estimates.

5.6 Conclusions

Surface-based SP monitoring during a constant head injection test performed within a 2.2 m interval of a well at 44 m depth in a fractured aquifer has been investigated through field measurements and numerical modelling. Water injection generated a negative SP response at surface that was concurrent with the start of injection. Steady-state SP at the end of injection was strongest near the well and generally decreased in magnitude further from the well, but remained anomalously strong along two azimuths corresponding to the

strikes of two fracture sets measured within the test interval. These field data, supported by numerical modelling, suggest that SP measurements can be used to infer prominent flow directions within a fractured aquifer. Strong reversals in the SP signals that occurred on release of the packer seal were interpreted to respond to flow reversal from the aquifer back into the well although that process has yet to be modelled.

The field measurements and modelling results both indicate that SP monitoring during the injection test has the ability to identify azimuthal variations in subsurface groundwater flow that parallel mapped fracture set orientations. Numerical modelling of fluid and current flow within a simplified fracture network showed patterns in SP are mainly controlled by transmissive fracture orientations. The strongest SP gradients occur parallel to hydraulic gradients associated with water flowing out of the transmissive fractures into the tighter matrix and other less permeable cross-cutting fractures. Sensitivity studies also indicate that increasing fracture frequency near the well increases the SP magnitude and enhances the SP anomaly parallel to the transmissive set. Decreasing the length of the transmissive fractures leads to more fluid flow into the matrix and cross-cutting fractures proximal to the well, resulting in a broader and higher magnitude SP anomaly.

The results of this study provide a new approach for the assessment of azimuthal hydraulic anisotropy in fractured rock aquifer systems having steeply dipping transmissive fractures. A sensible next-step would be to model a realistic discrete fracture network (DFN), with network characteristics inspired by field measurements from the Black Brook Watershed, to investigate how controlled variations in fracture orientation, length, frequency, and hydraulic conductivity within the aquifer would affect the SP response at

surface. The effect of elevated fracture electrical conductivities on conduction current paths and resulting SP anomalies should also be considered.

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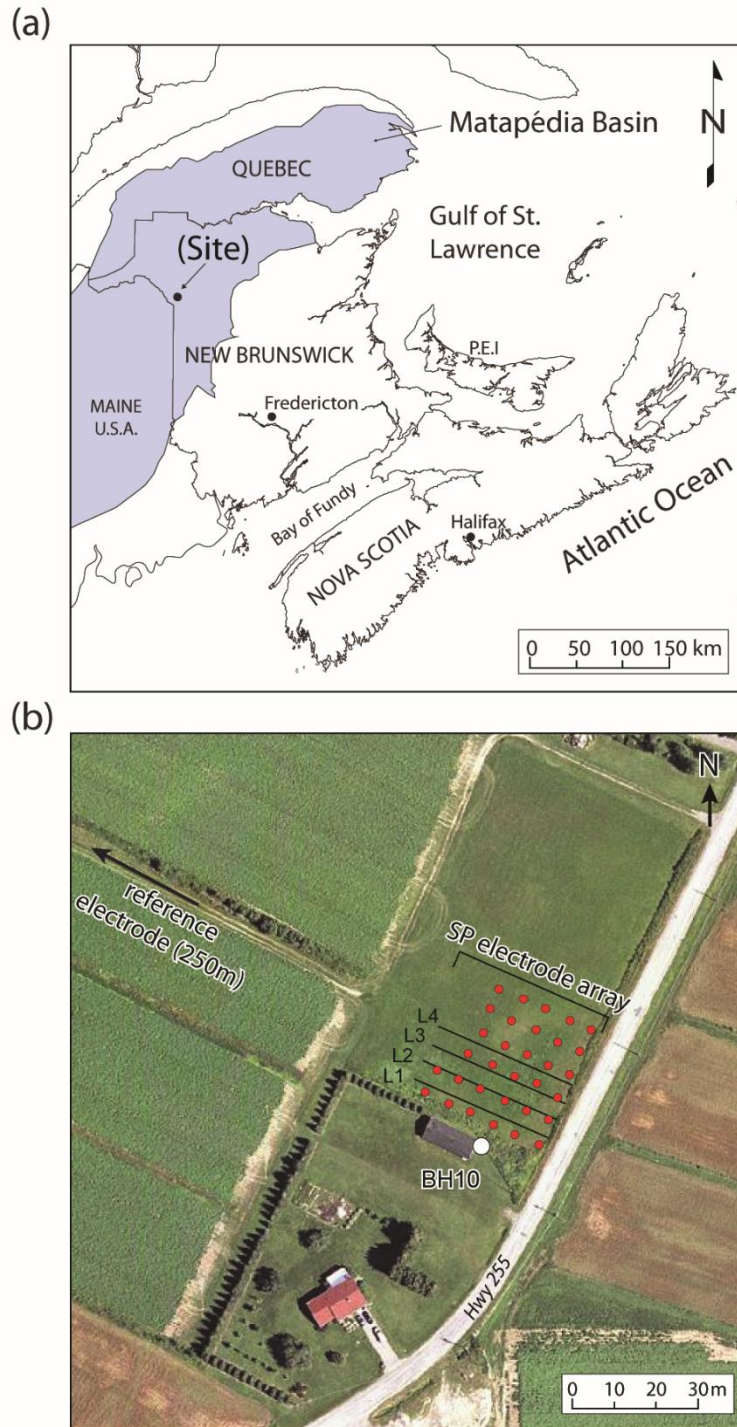


Figure 5.1. a) SP study area located in northwestern New Brunswick, Canada within the Matapédia Basin; b) location of the 32 electrodes (red dots) and four resistivity profiles (L1 – L4) near BH10. Reference electrode is located 250 m from BH10.

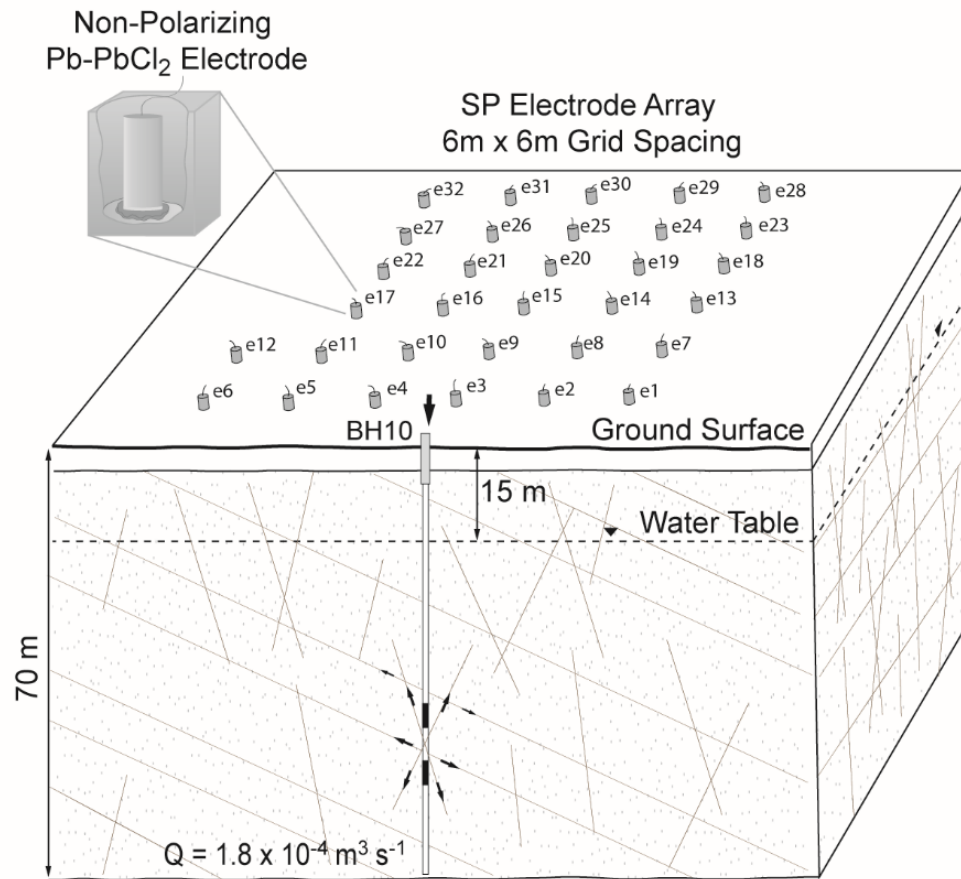


Figure 5.2. Schematic diagram of 2.2 m packer interval isolating a portion of fractured bedrock aquifer between 43.2 and 45.4 m depth. The aquifer is overlain by approximately 12 m of unsaturated bedrock and 3 m of overburden. Figure depicts position of SP electrode array on surface to the north of well BH10.

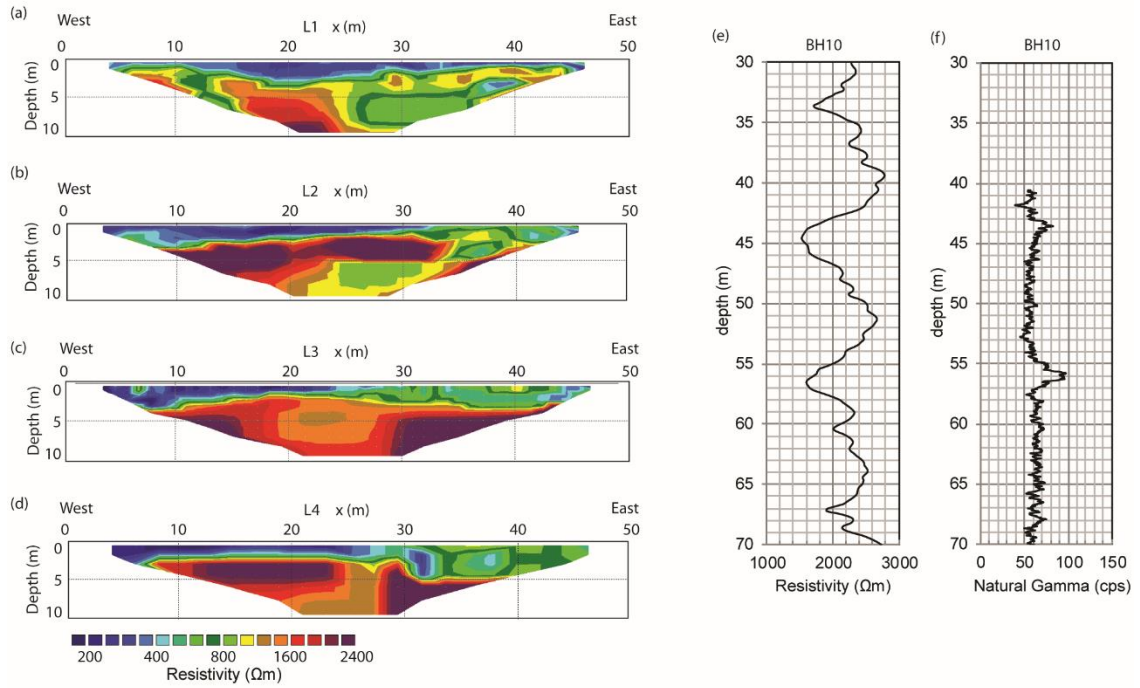


Figure 5.3. Four electrical resistivity tomography profiles (a – d) completed at the site with 2 m electrode spacing showing a sharp electrical contact between the overburden and the unsaturated bedrock. Overburden appears as low resistivity increasing into the unsaturated bedrock. The profiles image to a depth of approximately 10 m. Well logs of (e) normal resistivity with 0.2 m electrode spacing and (f) natural gamma ray emissions are shown for portions of BH10 that were successfully logged.

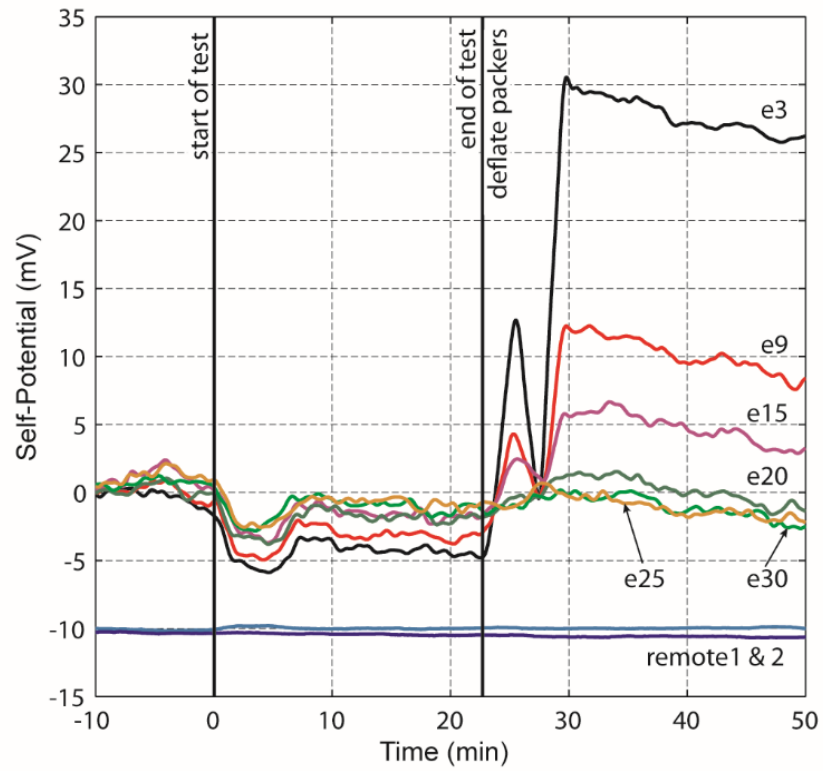


Figure 5.4. SP observed by six electrodes at distances of 6 to 41 m from the borehole collar during the injection of water into the borehole interval between 43.2 and 45.4 m depth. All dipoles were temporally referenced prior to the injection test, and a -10 mV DC shift was applied to the remote dipoles to remove overlap with electrode data from the array.

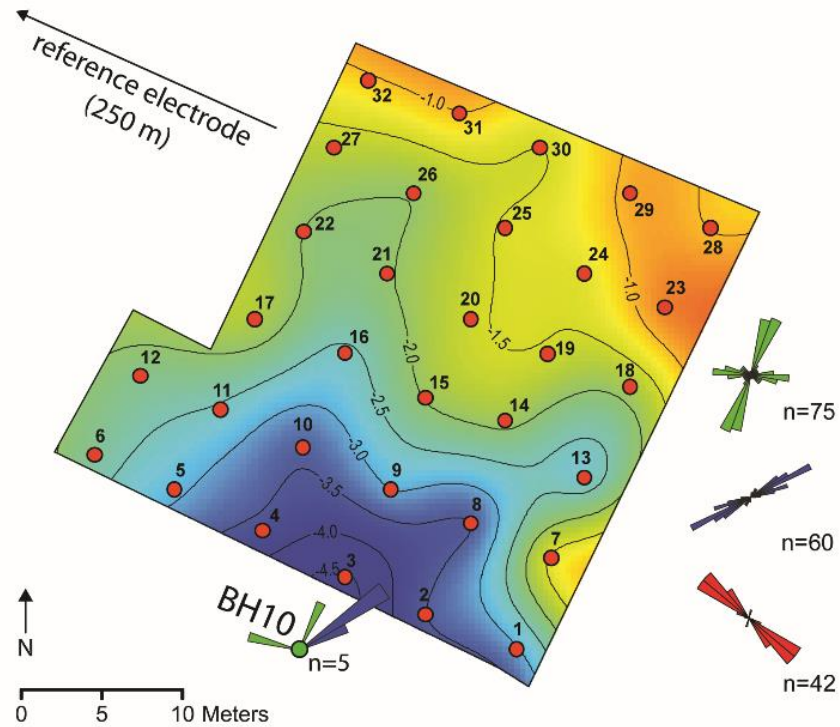


Figure 5.5. SP anomaly recorded at electrodes (red circles) on surface resulting from the constant head injection test in a packer interval between 43.2 and 45.4 m depth in BH10. The SP data represents the change in SP response recorded between ambient conditions (prior to injection), and conditions related to the end of water injection. The rose diagram located on the injection well (BH10) shows the strike of the 5 fracture planes that intersect the isolated injection interval. The rose diagrams on the right correspond to 177 fracture planes that intersect the entire length of the borehole (red = calcite-filled fractures, blue = bedding plane fractures, green = high angle fractures).

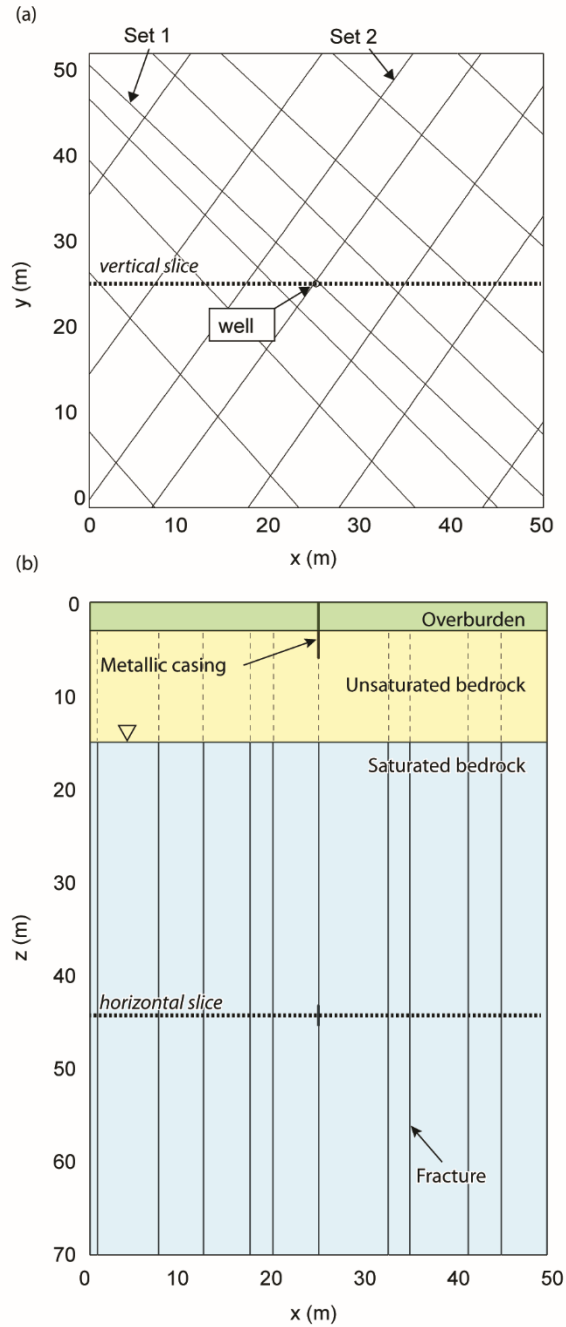


Figure 5.6. Model representing simplified fracture geometry observed in the Black Brook Watershed. (a) Plan view of the model subdomain showing the two vertical fracture sets incorporated into the numerical model. Fracture set 1 has an azimuth of -43.3° (316.7°), and fracture set 2 has an azimuth of 36.3° from north. A fracture from each set intersects at the well within the packer interval. (b) vertical section of the model subdomain showing the vertical fracture sets through the saturated and unsaturated portions of the model. Metallic casing is shown extending through the overburden unit into the upper unsaturated bedrock.

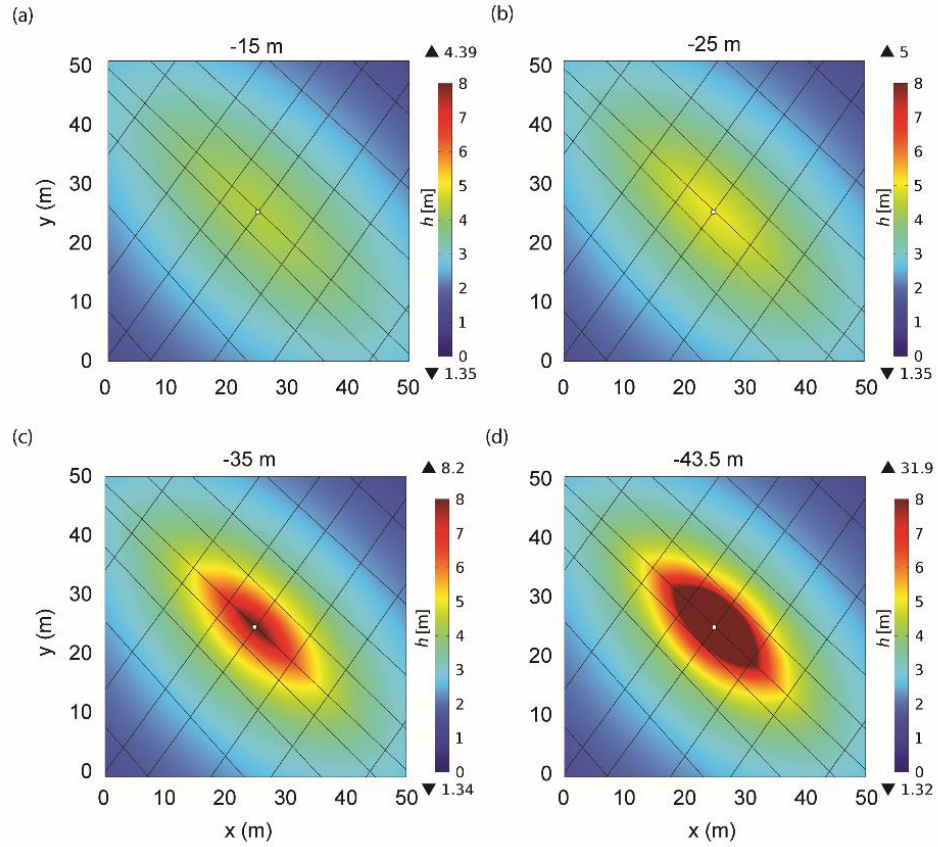


Figure 5.7. Horizontal slices of hydraulic head (m) through the 3D model at various depths that would be expected at steady state for water injection through a 2.2 m packer interval between 43.2 and 45.4 m depth. The shape of the hydraulic head distribution is elongate parallel to fracture Set 1 with higher hydraulic conductivity, which generates stronger hydraulic gradients perpendicular to these fracture planes.

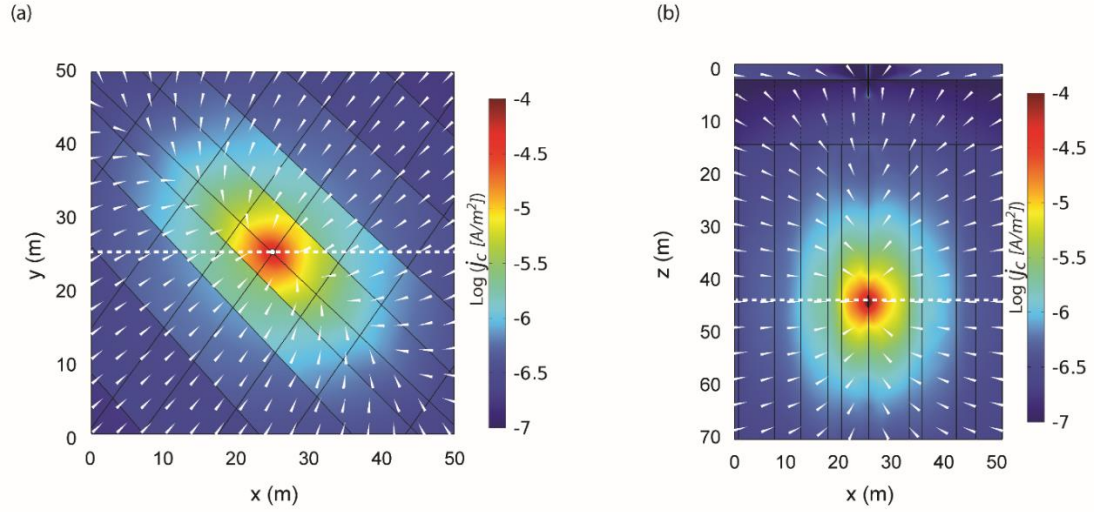


Figure 5.8. Conduction current magnitudes (colour) and directions (arrows) simulated for water injection into the fractured aquifer model. (a) horizontal slice and (b) a vertical section through the packer interval (-43.5 m). The location of the horizontal slice and vertical sections are represented as a white dashed lines. The distribution of conduction current density is presented as log-values ($\text{Log}(j_c)$) where j_c is in units of A/m^2 .

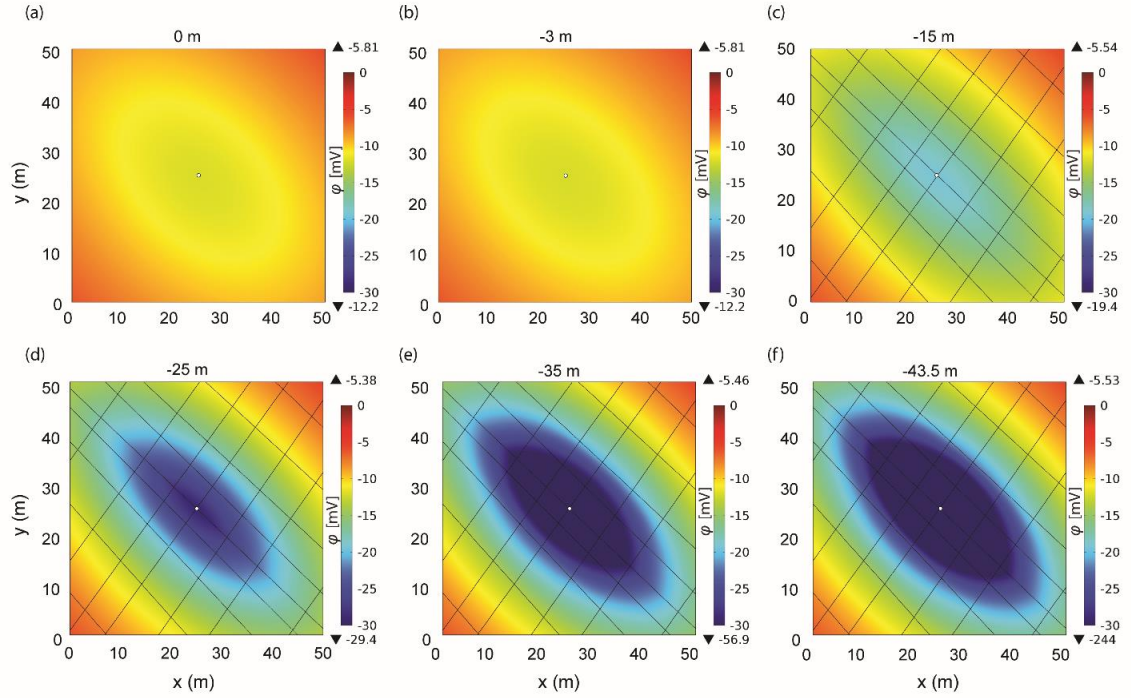


Figure 5.9. Horizontal slices of SP (in mV) through the 3D model at various depths that would be expected for water injection through a 2.2 m packer interval between 43.2 and 45.4 m depth. Slice depths are labelled above each panel; colour scale is the same for all panels, with minimum and maximum SP values labelled, saturating at depth near the injection interval. SP distribution at all depths shows a preferential elongation parallel to fracture set 1 associated with higher hydraulic conductivity. The SP anomaly weakens and becomes less elongated at shallower depths, in particular through the unsaturated bedrock and overburden units, despite the presence of shallow metallic well casing (upper 6 m of model)

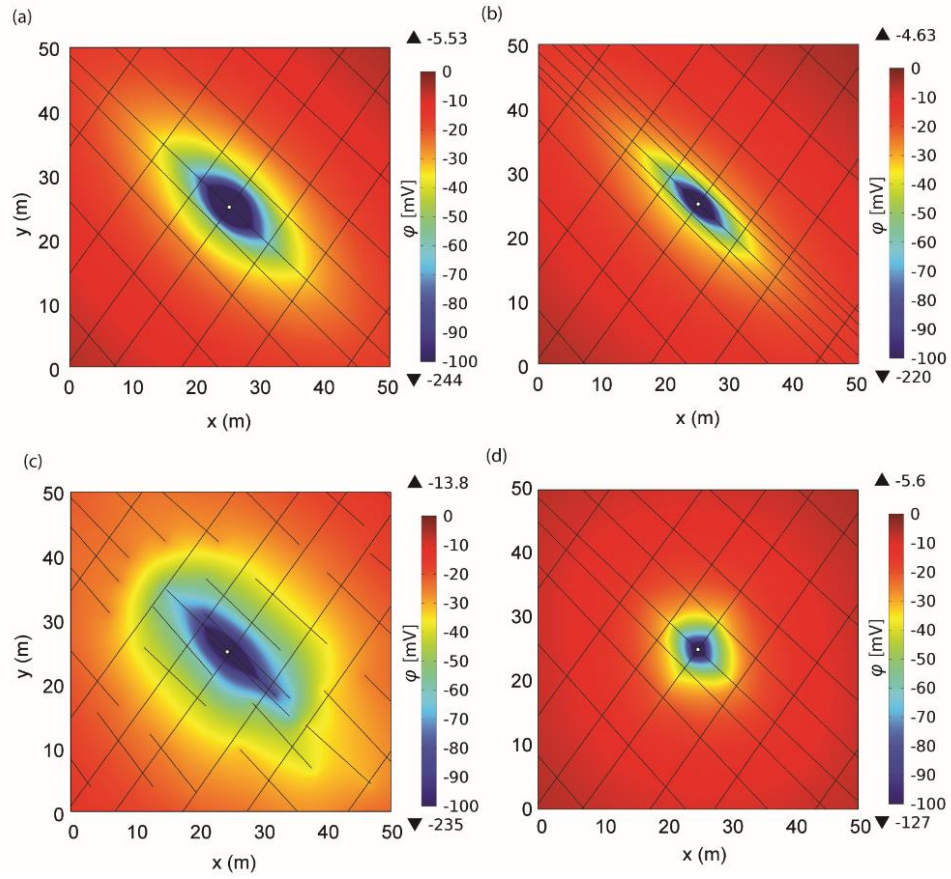


Figure 5.10. Self-potential distribution simulated at a depth slice through the packer interval (-43.5 m) for different scenarios adjusting the fracture geometry and hydraulic conductivity parameter. The SP maps show (a) the base case result from Figure 5.9f; (b) the effect of increasing the number of hydraulically conductive fractures; (c) the effect of hydraulically conductive fractures that are shorter and discontinuous; and (d) the effect of making the hydraulic conductivity of both fracture sets equal.

Chapter 6

Conclusions and recommendations

In chapters 2 and 3 of this dissertation, borehole and outcrop fracture measurements were integrated with structural, hydrogeological and borehole geophysical data to develop advanced conceptual models for fractured bedrock aquifers in two significantly different sedimentary basins of New Brunswick. Through field experiments and numerical modelling, chapters 4 and 5 demonstrated that self-potential fields measured at surface, during pumping and hydraulic testing, may serve as a relatively non-invasive and inexpensive tool to approximate the shape of the water table and estimate equivalent porous media hydraulic properties, or to infer fracture-related azimuthal anisotropy in fluid flow. With the objectives of the thesis in mind (Section 1.4), the following sections present the conclusions and main contributions of the thesis.

6.1 Aquifer conceptual models

The development of aquifer conceptual models and the estimation of hydraulic properties for the two fractured rock aquifers was an initial objective of this thesis. The results contributed to wellfield and watershed management goals external to this thesis, while also supporting the development of SP monitoring techniques in fractured rock.

The study presented in Chapter 2 characterized the fractured sandstone-shale aquifer of the Springdale Wellfield within the Carboniferous Moncton Subbasin, and showed the influence of fracture anisotropy and lithological heterogeneity at the local aquifer-scale. By incorporating hydrogeophysical borehole logging information with pump test results and knowledge of the regional geology, a conceptual model of the aquifer was established. The

aquifer consists of channelized sandstone deposits that are lenticular in nature, and adjacent laminated shale units. Fractures in the sandstone units are significantly more conductive than the fractures in the shale units. Flow logs combined with televIEWER images identified significant flowing intervals at bedding parallel fracture zones located at the upper and lower boundaries of the sandstone unit. Pump tests revealed that the aquifer is confined and identified the presence of one or more impervious boundaries limiting the lateral extent of the aquifer. Such an interpretation is consistent with the lenticular nature of channelized sandstone units, also given that the degree of fracturing and connectivity between fracture sets appears to decrease with the transition into the shale units.

Chapter 3 characterized fracture properties of an unconfined aquifer in folded turbidite deposits (calcareous to argillaceous mudstones to siltstones) underlying the Black Brook Watershed in the Silurian Matapédia Basin. A conceptual model for the fracture network controlling groundwater flow was established based on a combination of hydrogeophysical logging, outcrop mapping, and constant head hydraulic testing. Borehole logging and hydraulic packer tests indicated groundwater flow is dominantly controlled by bedding plane fractures with flow paths parallel to the regional northeasterly structural trend. Hydraulic test results suggested a general decrease in hydraulic conductivity with depth, which corresponded to a decrease in the frequency of fractures exhibiting oxidation and weathering. The conceptual model provided a valuable approach to predict azimuthal anisotropy in hydraulic conductivity in the remaining portions of the BBW, as well as adjacent watersheds elsewhere within the Matapédia Basin. Such insight is important within this agricultural region to develop watershed best management practices, and better-inform water well placement. Careful extrapolation of this approach can be helpful in similar

structural settings throughout the Matapédia Basin where boreholes and outcrop exposures are limited, whereby a simpler understanding of the regional structural trends may allow for the dominant fracture pathways to be predicted.

6.2 SP responses during pumping and hydraulic testing

Chapters 4 and 5 used findings from the previous two chapters as a basis for interpreting SP signals recorded on the surface during periods of pumping and hydraulic testing. Findings from these two investigations demonstrate that SP measurements can be used to evaluate flow in fractured rock settings, in conditions of either steady-state or transient flow. In addition, findings from one of the SP investigations demonstrated the use of a multi-channel SP recording system developed to efficiently remove noise from the electrical signal prior to recording the data.

The first case (Chapter 4) demonstrated the application of surface-based SP monitoring to infer drawdown during pumping, and estimate spatial variations in the apparent transmissivity of the heterogeneous and confined fractured sandstone/shale aquifer at Springdale, NB. Pumping-induced SP anomalies were measured at 31 locations around a municipal production well using a multi-channel SP recording system. With the aid of passive analog lowpass filters, the signal-to-noise ratios were improved by a factor of ten compared to those obtained using the data logger's powerline noise cancelling sampling mode alone. SP anomalies reached magnitudes of 2 - 9 mV at distances up to 36 m from the well during pumping at a rate of $7.8 \text{ L}\cdot\text{s}^{-1}$ for 300 minutes. Using an apparent electrokinetic coupling coefficient obtained from a single electrode-monitoring well pair, apparent drawdown curves were inferred below each electrode location using the SP signals over the duration of the pumping period. Inferred drawdown curves for all electrode

locations were interpreted using the Theis model to provide estimates of apparent transmissivity that were comparable to estimates obtained using conventional pump tests and direct drawdown measurements.

Numerical modelling using an equivalent porous medium approach demonstrated that, for three relatively shallow aquifer scenarios, the hydraulic conductivity contrast between the permeable sandstone channel and the surrounding shale unit gave rise to a prominent inflection in the modelled SP profiles directly above the edge of the channel. This confirmed the sensitivity of surface SP measurements to lateral hydraulic heterogeneities. It was also noted that under the same electrical conditions and pumping rate, the magnitude of SP simulated on the surface increased when the hydraulically conductive channel unit was removed, which is associated with a hydraulic gradient increase in the unit. A change in electrical conductivity structure, in particular removing the electrical conductivity of the clayey glacial till confining layer, also caused the SP magnitude to increase on the surface. SP modeled on the surface near the pumping well in each of the cases revealed linear correlations with modeled hydraulic head at the top of the aquifer, despite the aquifer being both electrically and hydraulically heterogeneous. The effect of a conductive well casing, which has been shown in previous studies to increase current flow near surface and thereby increase SP effects on surface, was shown to be modest to negligible. Differences in length of conductive casing and open well bore geometry between previous studies and this study likely explain the disparity in the casing effects on surface SP magnitudes.

Forward modelling and field monitoring showed that SP data measured on the surface can be used to complement a limited number of monitoring wells to infer the

evolution of a drawdown pattern during pumping and estimate hydraulic properties of an aquifer. The analyses suggest that SP monitoring should be effective where confined aquifers are overlain by overburden of uniform conductance. In addition, they show that SP monitoring during pumping can be a valuable way of obtaining a high density of hydraulic data needed for tomographic studies to determine hydraulic properties of a heterogeneous aquifer. This can be conducted in a way that is non-invasive and significantly less expensive than installing tens or hundreds of monitoring wells.

The second application of the SP method (Chapter 5) focused on surface-based SP signals monitored during a constant head injection test at 44 m depth in the fractured calcareous to argillaceous, mudstones to siltstones underlying the Black Brook Watershed in northwestern New Brunswick. To our knowledge, this study was the first of its kind to monitor induced fluid flow from injections in discrete packer intervals, and to show preferential flow paths that follow fracture sets previously identified within the interval. We illustrate that a measurable SP response was produced at the earth's surface where the occurrence of negative SP was concurrent with the start of injection. Steady-state SP at the end of injection was strongest near the well and decreased in magnitude further from the well but not in a radially symmetric fashion. Coincidence between the shape of surface SP anomaly and fracture strike orientations measured within the test interval suggest electrical measurements can be used to infer dominant flow directions within a fractured aquifer. A strong reversal in the SP signal that occurred upon release of the packer seal is speculated to be a consequence of flow reversal from the aquifer back into the well.

Numerical modelling of fluid and convection current flow within the simplified fracture network showed patterns in SP are mainly controlled by transmissive fractures

orientations. The strongest SP gradients occur parallel to hydraulic gradients associated with water flowing out of the transmissive fractures into the tighter matrix and other less permeable cross-cutting fractures. Sensitivity studies also indicate that increasing fracture frequency near the well increases the SP magnitude and enhances the SP anomaly parallel to the transmissive set. Decreasing the length of the transmissive fractures leads to more fluid flow into the matrix and into cross-cutting fractures proximal to the well, resulting in a broader and higher magnitude SP anomaly.

Overall, the results presented in this thesis clearly demonstrate that SP data recorded on surface during pumping and hydraulic testing can be used as a tool to infer fracture-related azimuthal anisotropy in fluid flow, approximate the shape of the water table and estimate equivalent porous media aquifer hydraulic properties. SP data recorded using a network of electrodes distributed over a large spatial extent can be used as a cost-effective method to augment the hydraulic data provided from traditional pumping or hydraulic testing projects. Hydraulic and SP data can be used together in numerical modelling to allow for better assessment of groundwater flow and contaminant transport predictions.

6.3 Recommendations

Despite the progress made towards understanding the influence of fracture-related anisotropy and heterogeneity on SP signals measured on ground surface, additional questions remain.

- It would be desirable to have a better physical understanding of the difference between an impervious and a leaky confining layer boundary in terms of their contribution to SP magnitude. Based on the rigorous modelling approach proposed by Sill (1983), the presence of a hydraulic gradient across a boundary in hydraulic conductivity results in

development of a current density source. This would be the case of a leaky confining layer boundary, where a very high hydraulic gradient would be defined across the boundary resulting in development of current sources, although fluid flow across the boundary would be negligible. Modelling by Wurmstich (1995) showed that leakage from a confining formation above a hydrocarbon reservoir resulted in weak conduction current (and hence, SP) sources. However, there is some uncertainty over how to handle a truly impervious layer where hydraulic head in an upper aquifer is completely unaffected by head in a lower aquifer, such that the hydraulic gradient between them is not defined.

- Although groundwater and current flow numerical models completed as part of this thesis were inspired by aquifer characteristics measured in the field, fractured rock domains were approximated as either equivalent porous media (EPM) or as a simplified fracture network with uniform assignment of properties. Through laboratory testing using a fractured block model, Wishart et al. (2006) revealed that changes in fracture frequency were responsible for changes in SP magnitudes measured on the surface. Similarly, Wishart et al. (2008) showed that SP magnitudes changed depending on fracture frequency at four fractured rock aquifers where flow was driven by a natural hydraulic gradient. While the modelling in chapter 5 clearly revealed effects of fracture network anisotropy, some uncertainty remains regarding the influence of fracture network heterogeneity (i.e., spatial variations in orientations, length, frequency, and hydraulic conductivity, etc.) on the SP response at surface. To address this question, a realistic distribution of fractures and fracture set parameters may be generated stochastically, incorporating measured field characteristics, to be used

as input into a coupled groundwater and current flow model. An example of a discrete fracture network (DFN) is shown in Figure 6.1, within a 50 x 50 x 20 m domain generated using FracSim3D (Xu and Dowd 2010). SP modelling results may be used to investigate the spatial relationship between fracture frequency and SP magnitudes on surface. Results may also allow for further testing of SP interpretation approaches specifically designed for dealing with heterogeneous aquifers to image spatial changes in hydraulic head and transmissivity during hydraulic testing investigations, such as geostatistical methods described by Straface et al. (2007; 2010) (and references within).

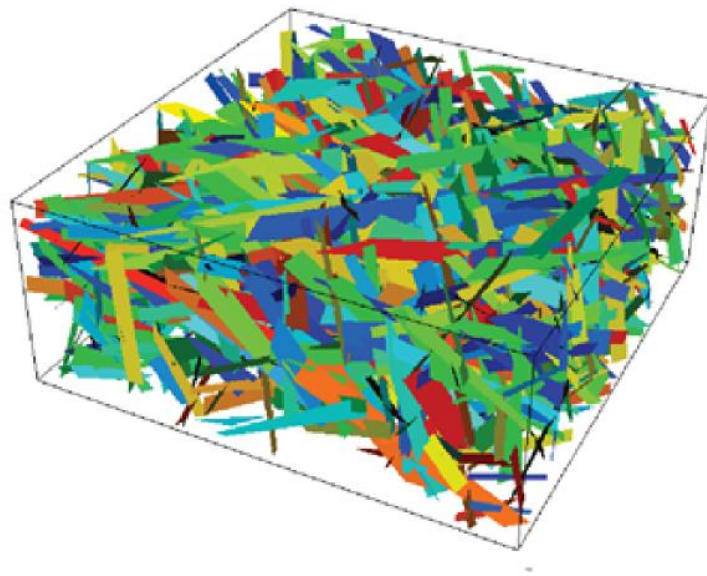


Figure 6.1. Stochastic discrete fracture network within a 50 x 50 x 20 m aquifer. Colour variations represent assigned hydraulic conductivities, based on parallel plate model using distribution of measured apertures.

- The SP modelling in this thesis largely assumed that the electrical resistivity distribution was predominantly horizontally layered. With all other parameters being equal, how strongly would lateral variations in electrical resistivity affect the SP patterns observed

at surface? In environments where surface SP can be used as a proxy for hydraulic head, understanding how lateral resistivity variation affect the SP distribution is important before using the high spatial density of SP measurements to help infer hydraulic head and subsurface variations in hydraulic properties. As an example, additional forward modelling would be useful to predict how surface SP patterns are affected by lateral variations in either the resistivity or the thickness of overburden.

- SP modelling presented in Chapter 5 assumed a uniform electrical conductivity through the fractures and matrix, which may be a reasonable to assume in a fractured aquifer with a clay component where saturated matrix electrical conductivity may be similar to that measured along fractures. However, in situations where the fluid to saturated-matrix electrical conductivity ratio (σ_f/σ_m) is higher, the effect of fractures on conduction current paths would likely increase. Although Roubinet et al. (2016) specified an electrical conductivity contrast between the matrix and fractures of 2 orders of magnitude, they did not investigate the sensitivity of their results to that contrast. Additional modelling or laboratory experiments could focus on determining how important it is to account for differences between the electrical conductivity of fractures versus the matrix.

6.4 References

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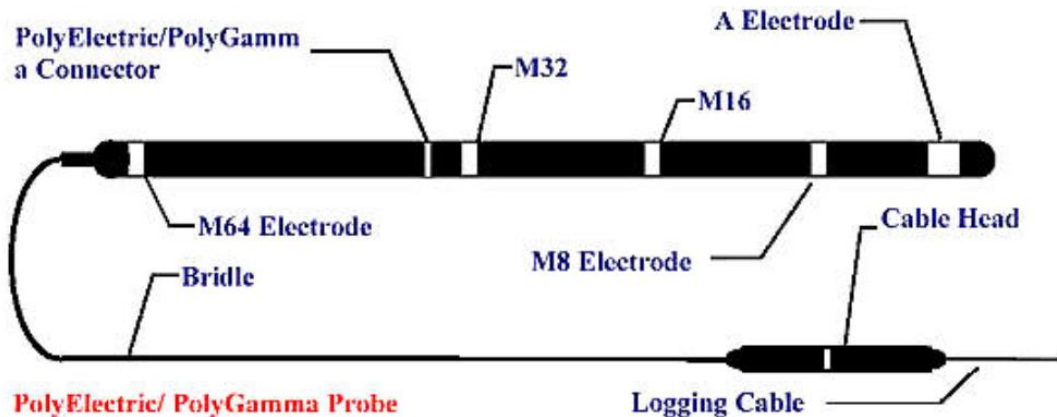
Appendix A:

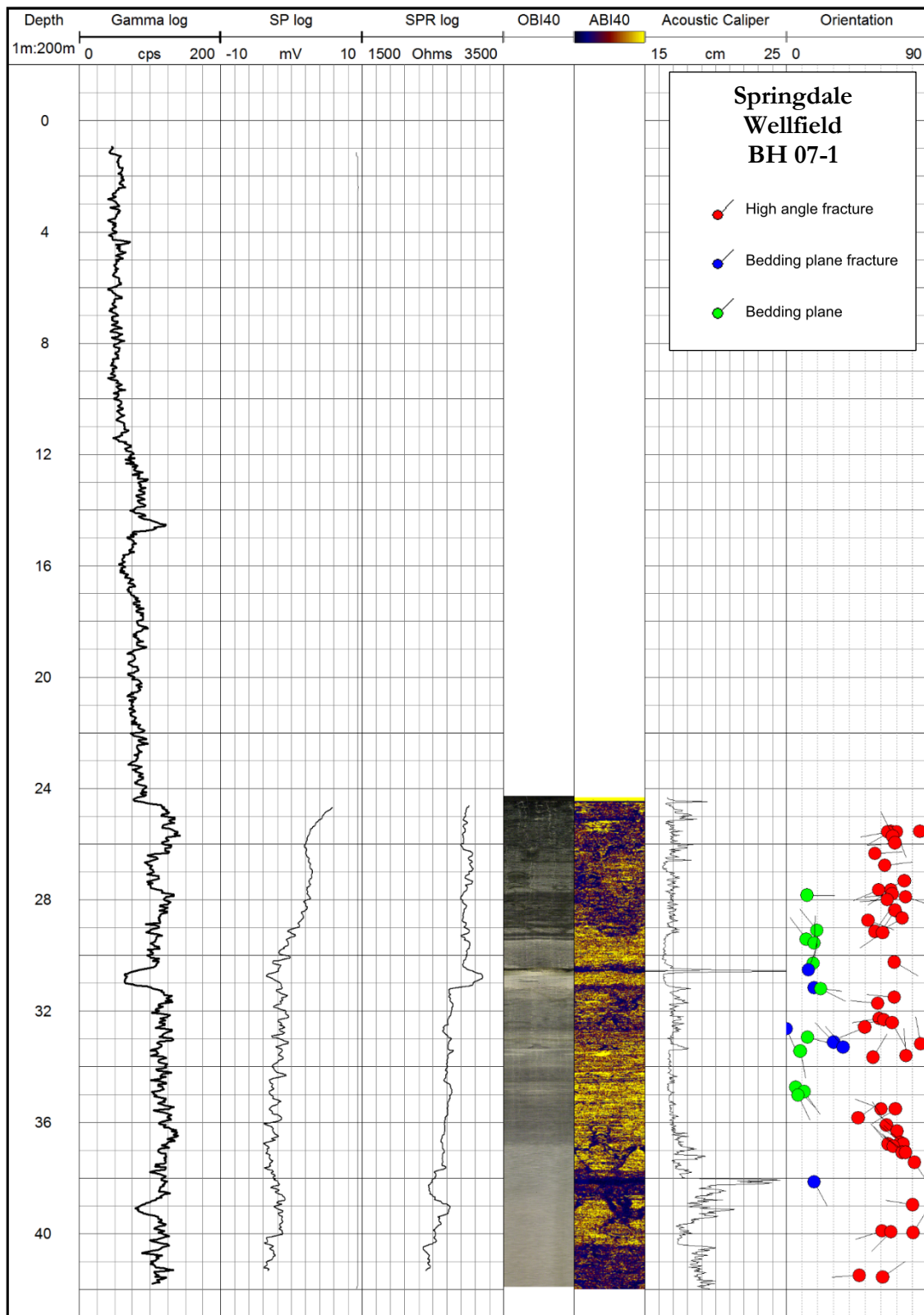
Compilation of borehole geophysical logs recorded within the Springdale Wellfield

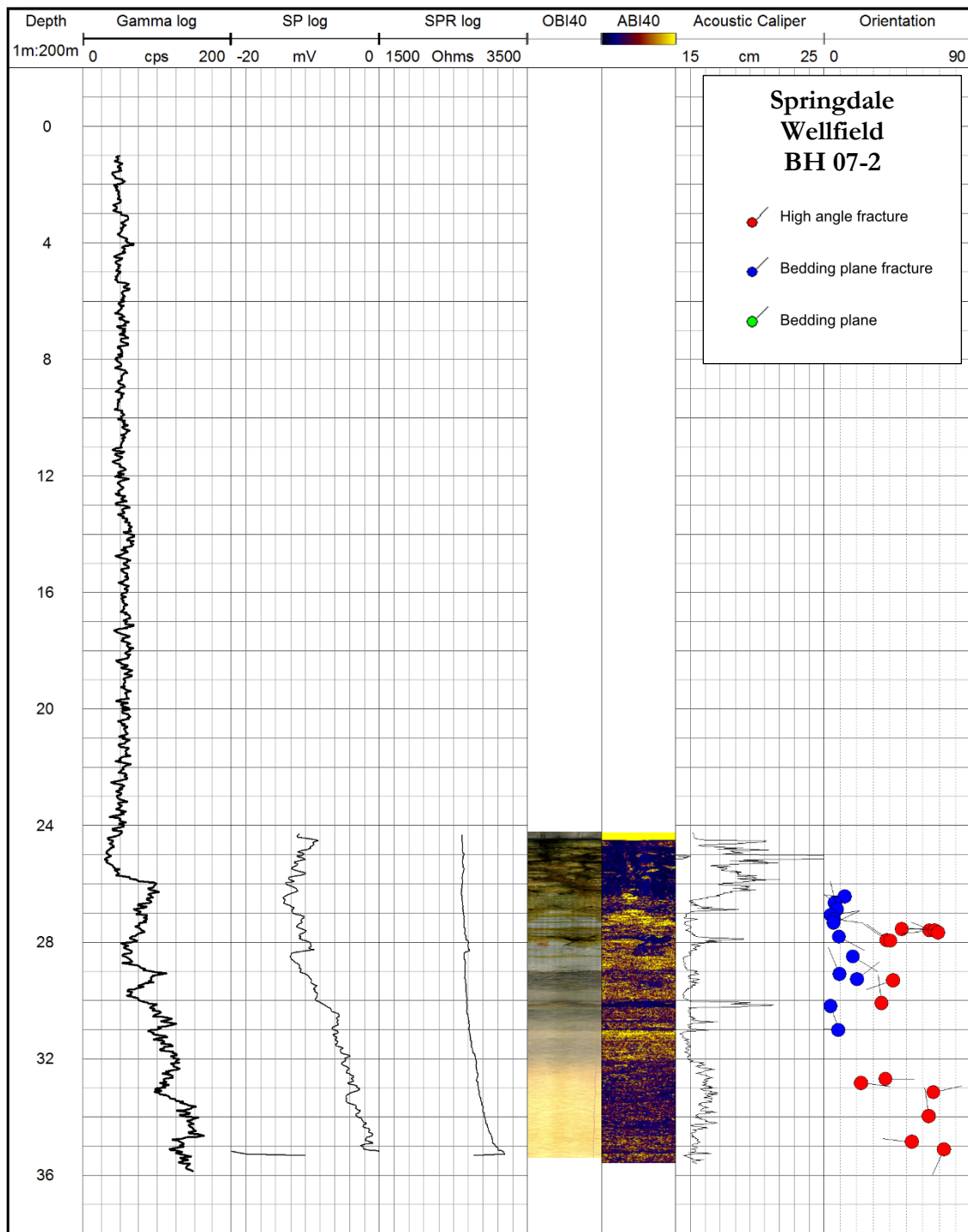
Borehole	X coordinate* (m)	Y coordinate* (m)	Depth to Bedrock (m)	Total Depth (m)	Top of Well Casing Elevation (m)
BH 07-1	2591773	7419825	10.7	42.7	71.7
BH 07-2	2591710	7419766	23.5	36.6	68.5
BH 08-1	2591864	7419598	12.2	58.8	74.6
BH 08-2	2591853	7419595	18.3	60.9	74.3
BH 09-2	2591856	7419602	4.6	91.4	115.5

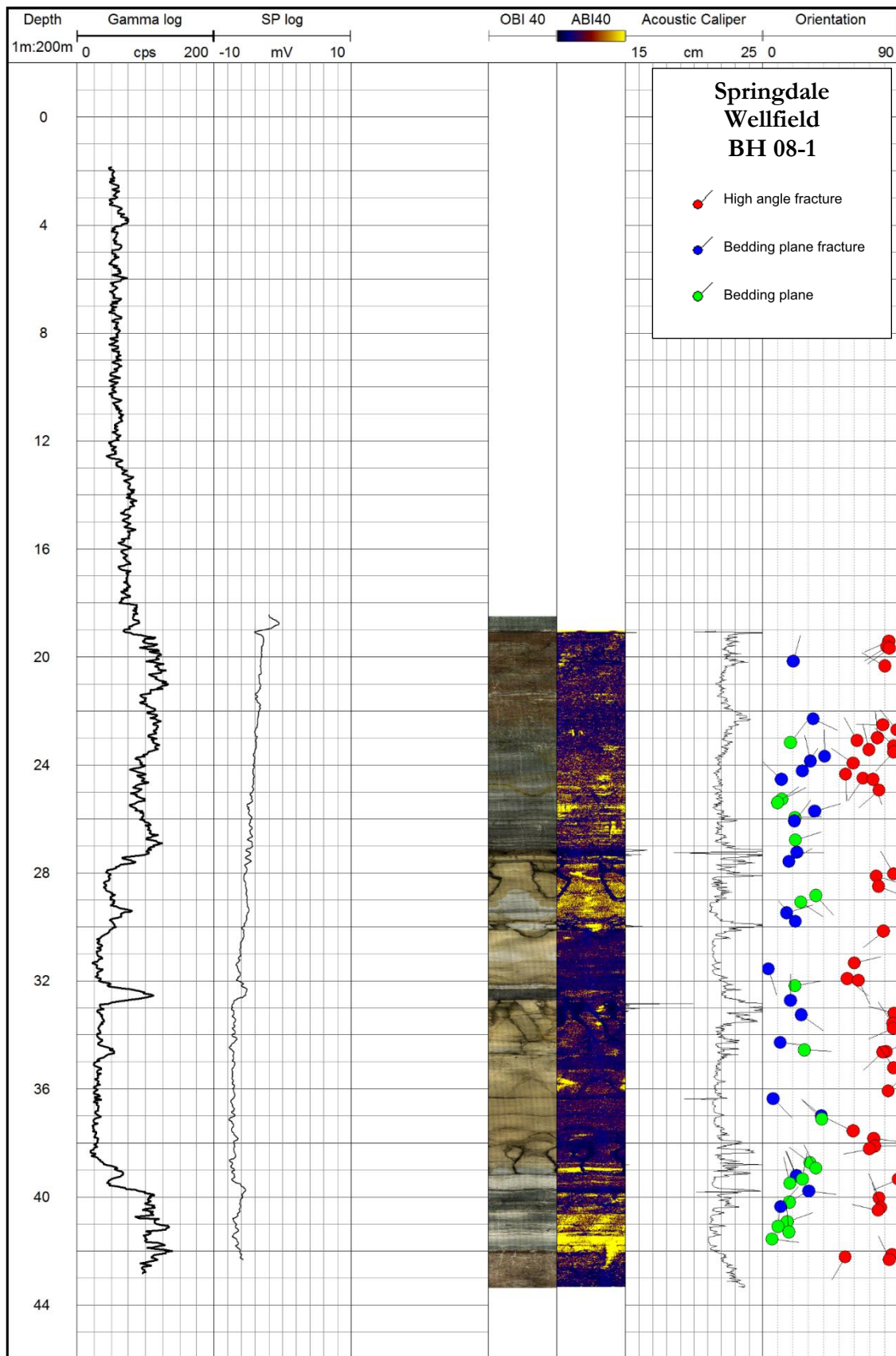
* NAD83 (CSRS) NB Double Stereographic projection

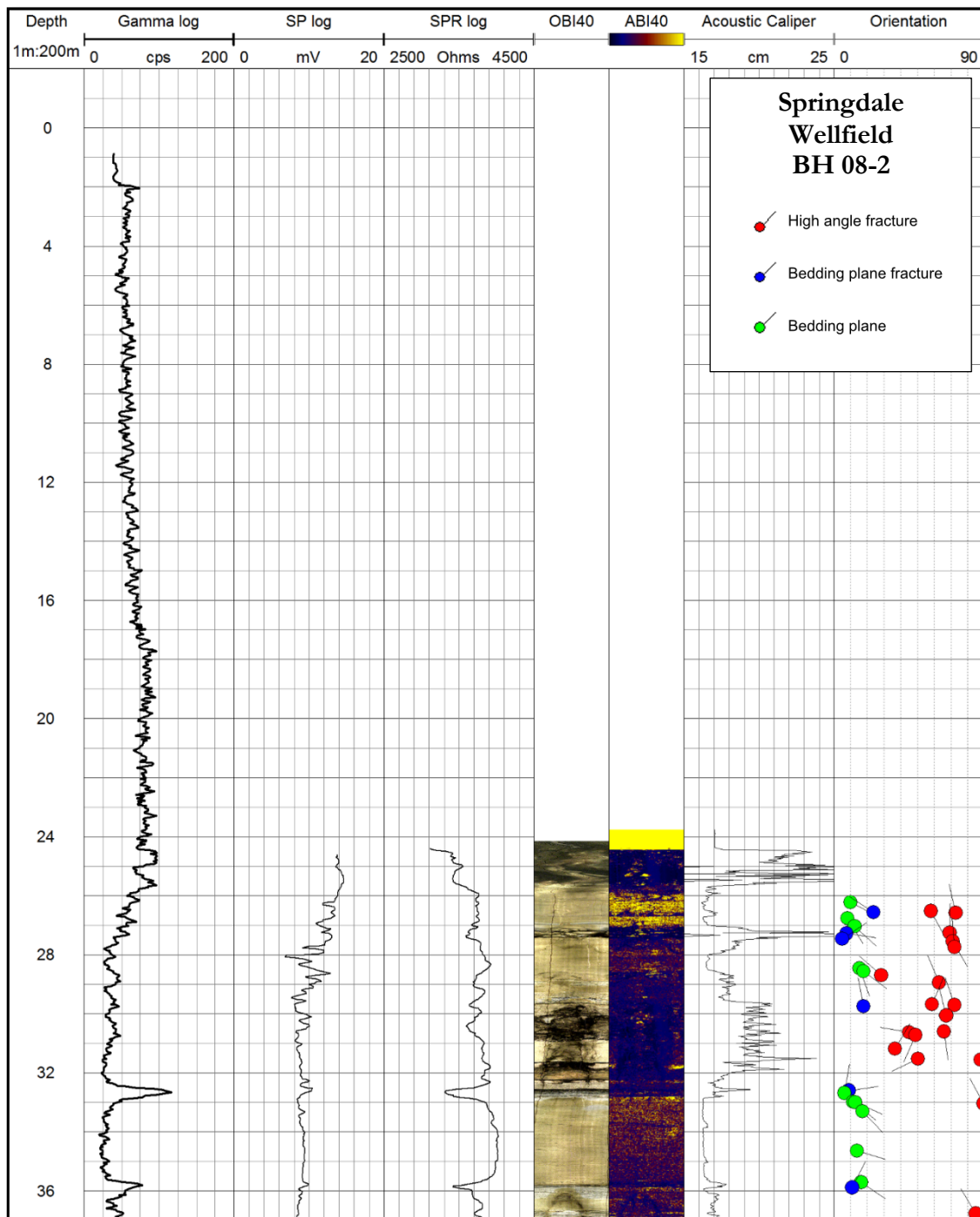
Borehole Geophysical Probe	Description
2PGA-1000	Measurement of gamma log, single point resistance log, and self-potential
ALT OBI-40	Optical televiewer imaging
ALT ABI-40	Acoustic televiewer imaging
2PFA-1000	Fluid Temperature and Conductivity
2PEA-1000	Multipoint Resistivity probe (see image below)
HFP-2293	Heat pulse flowmeter

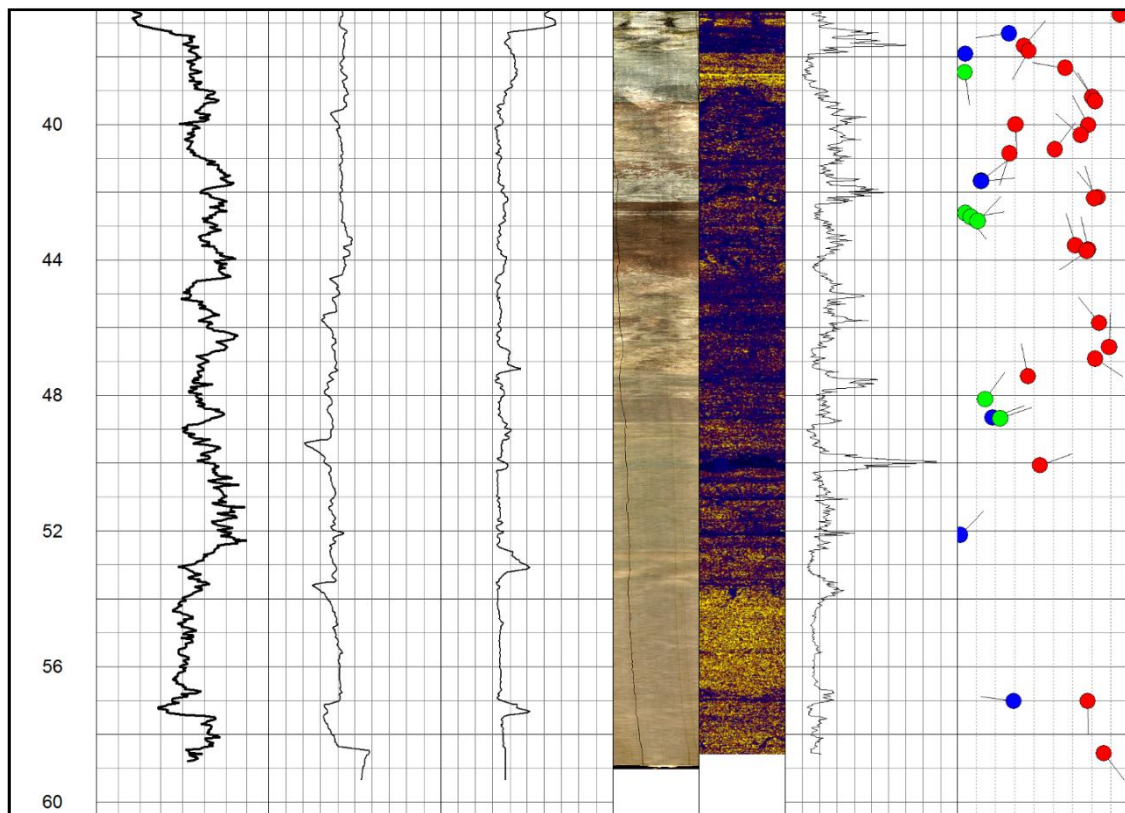


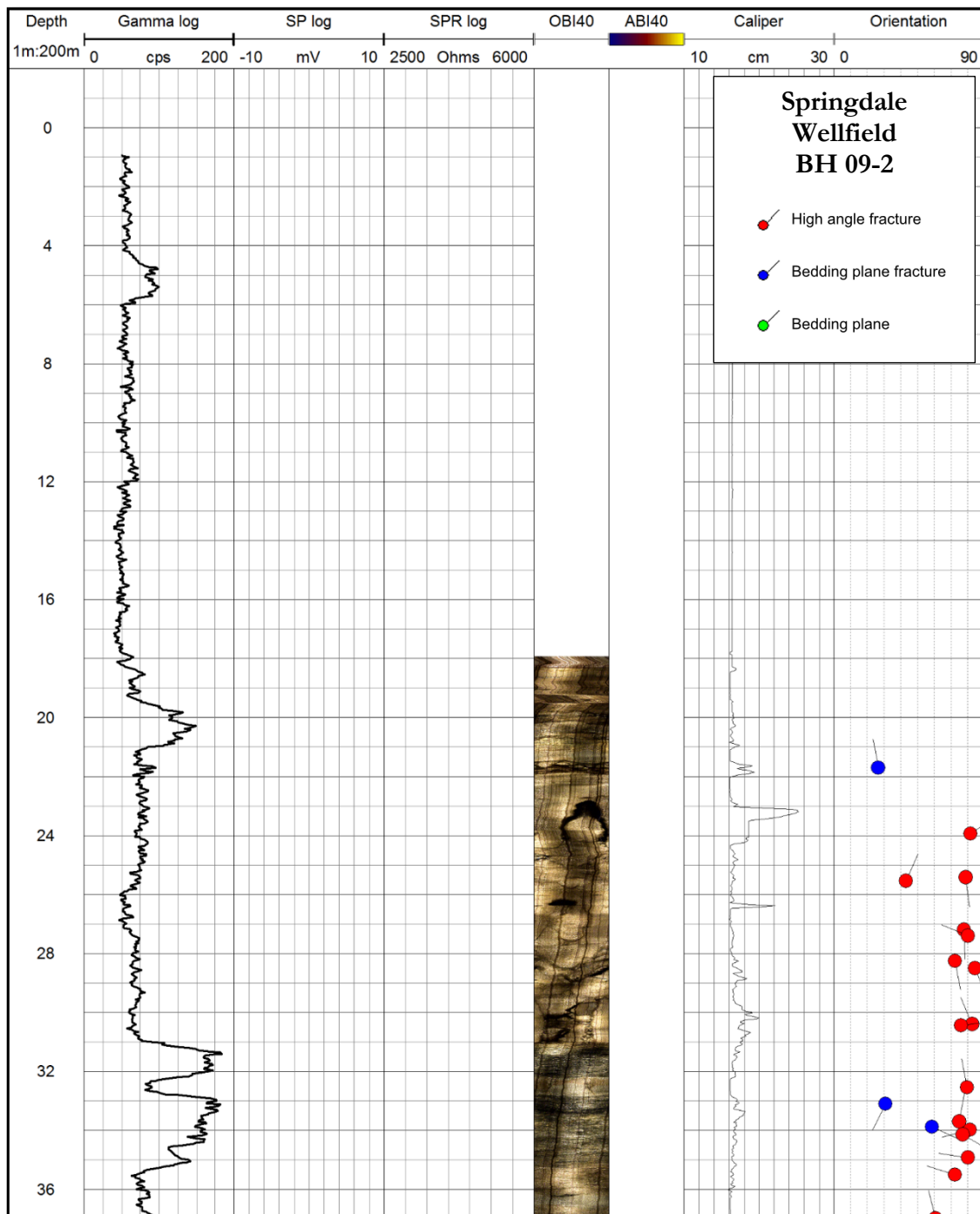


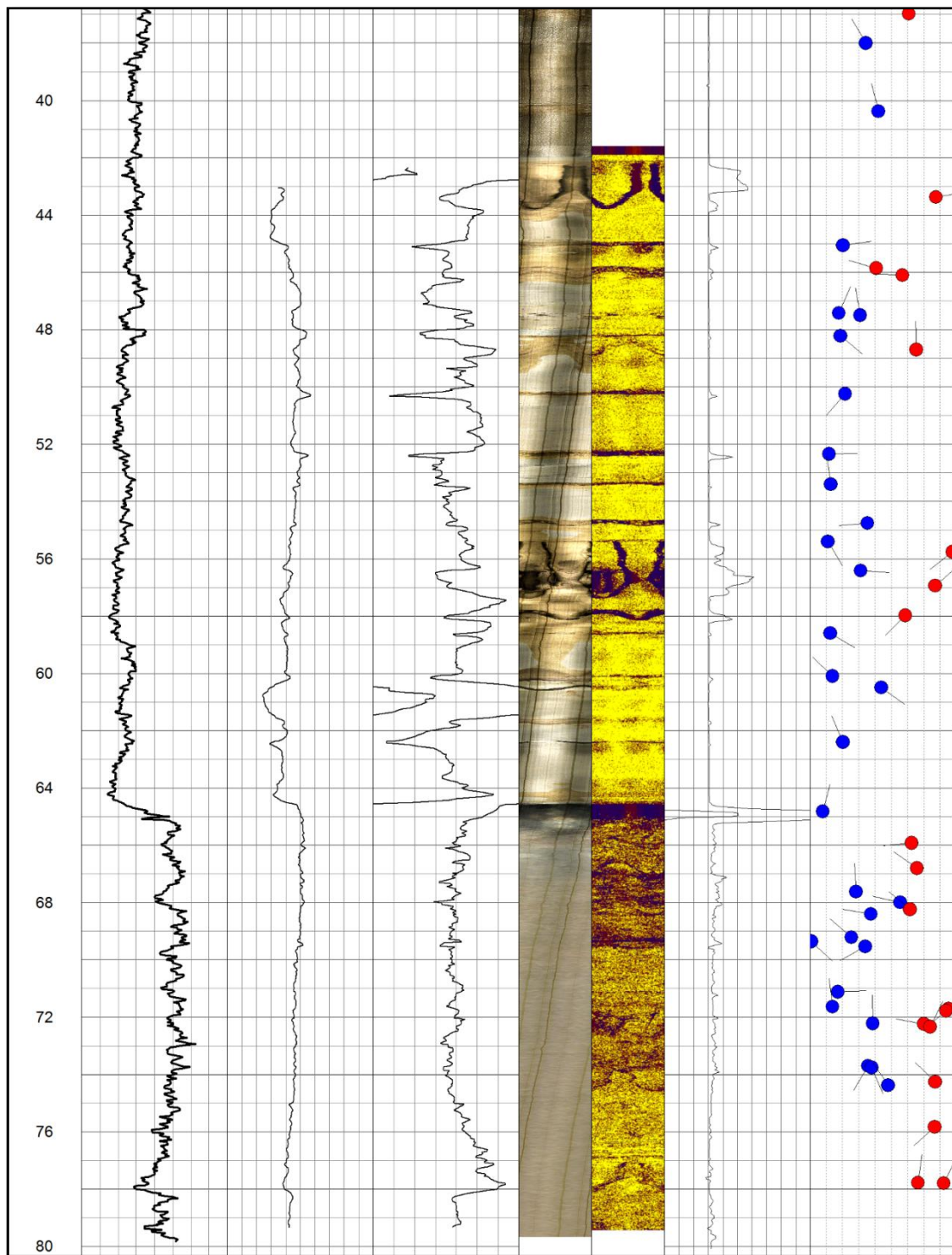


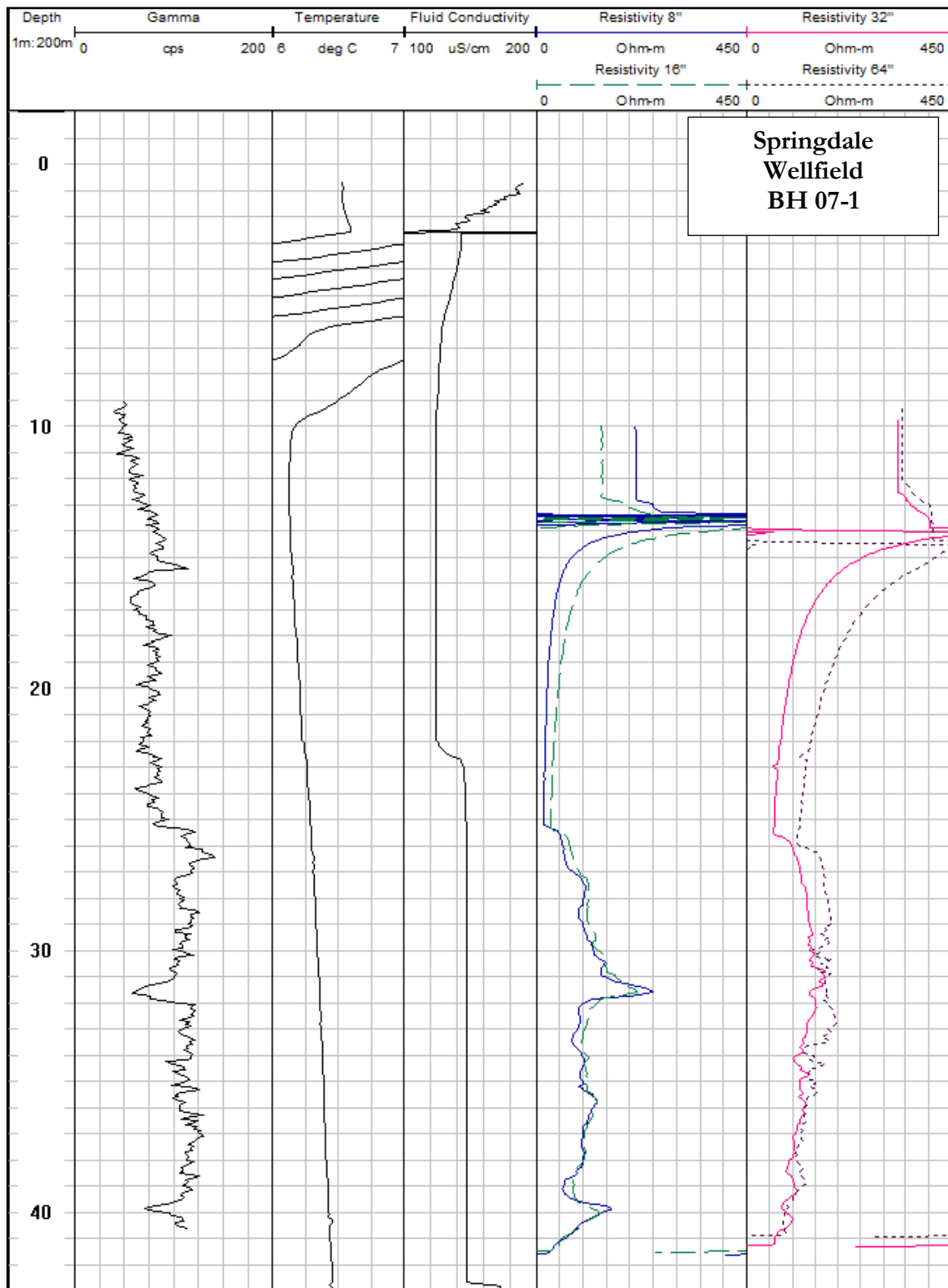


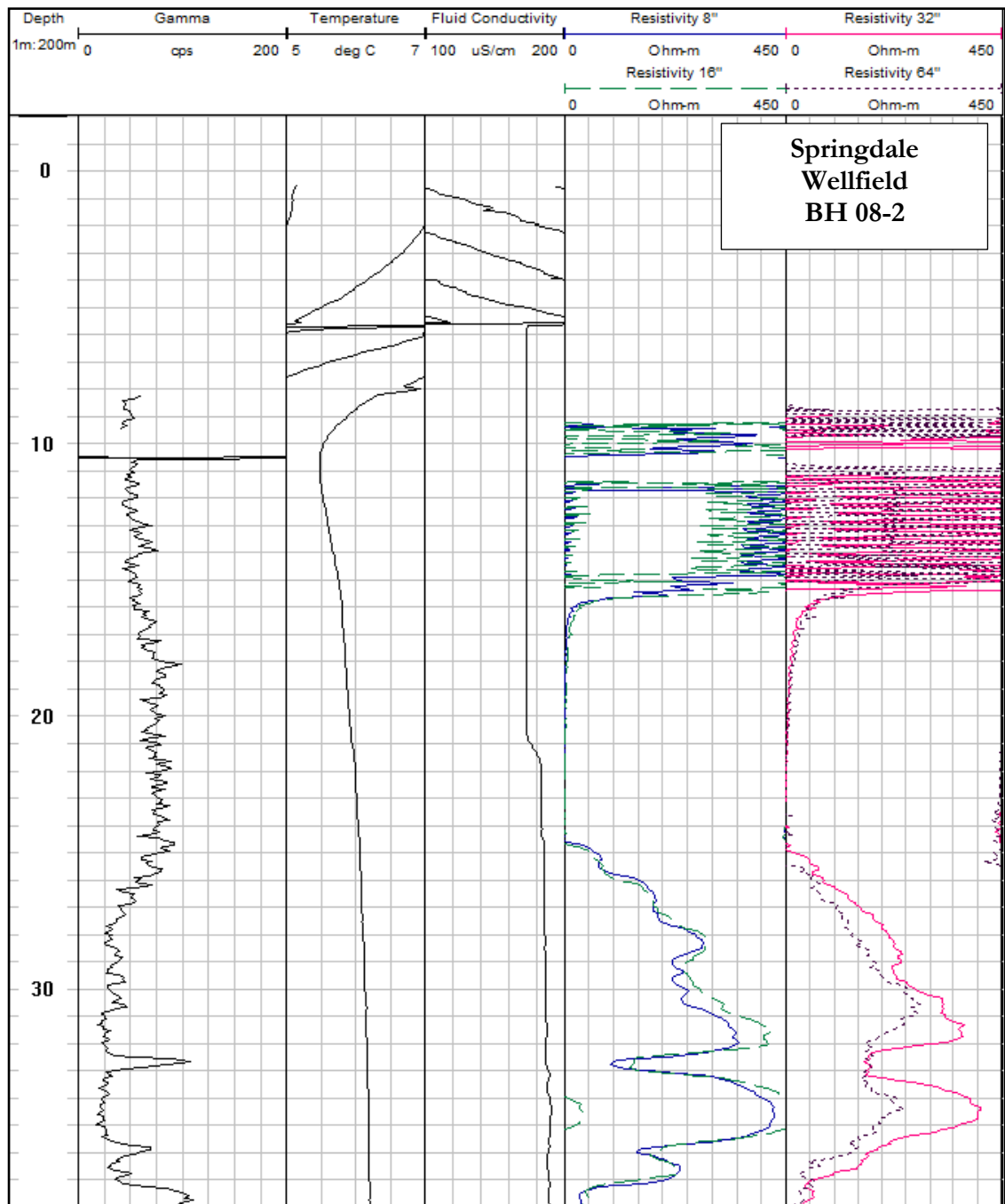


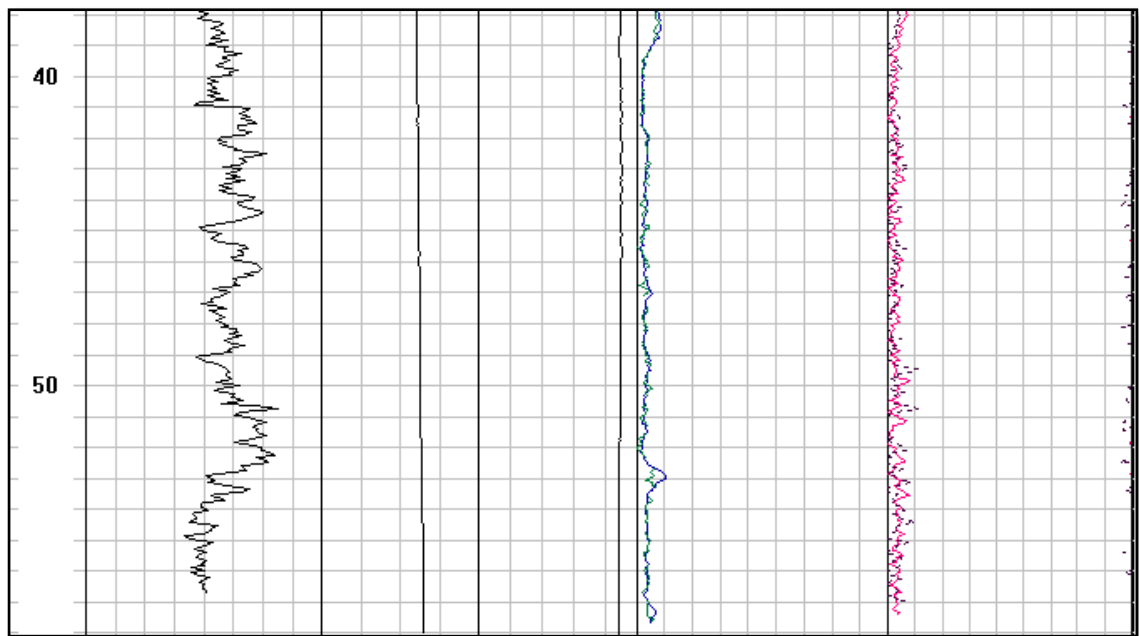


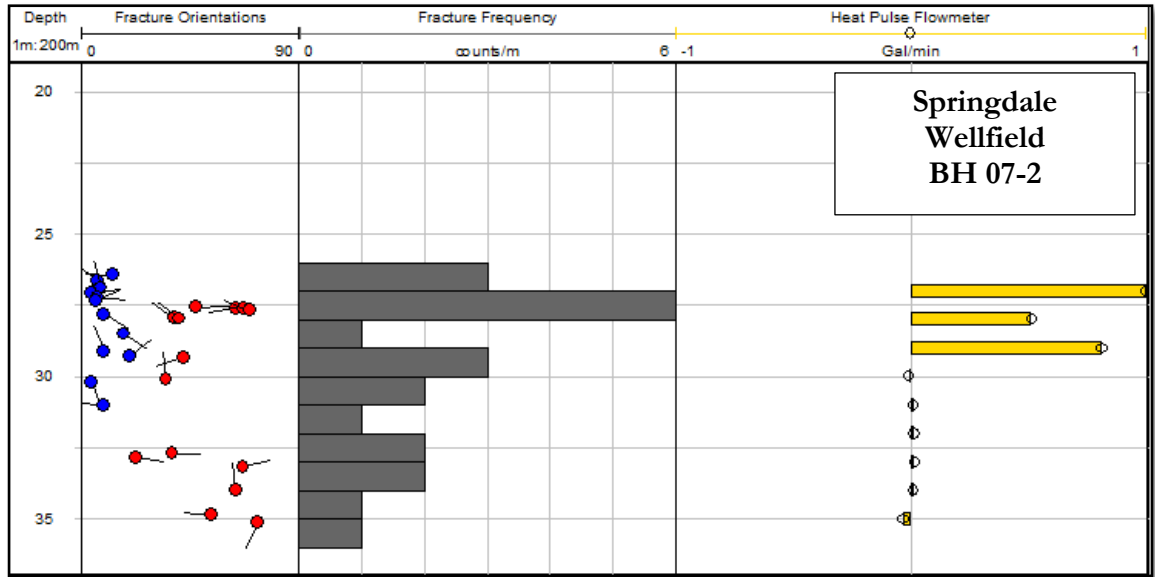


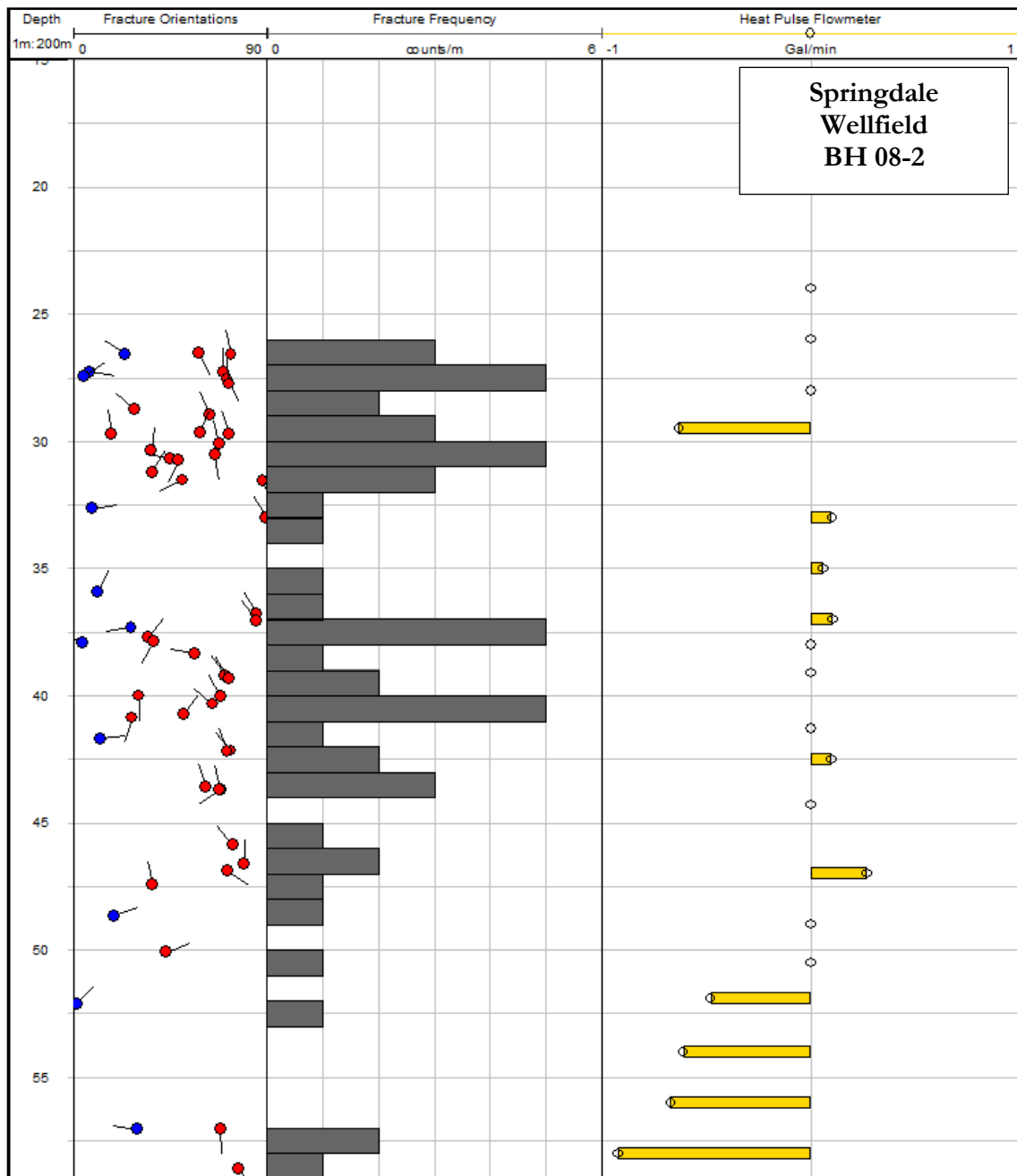












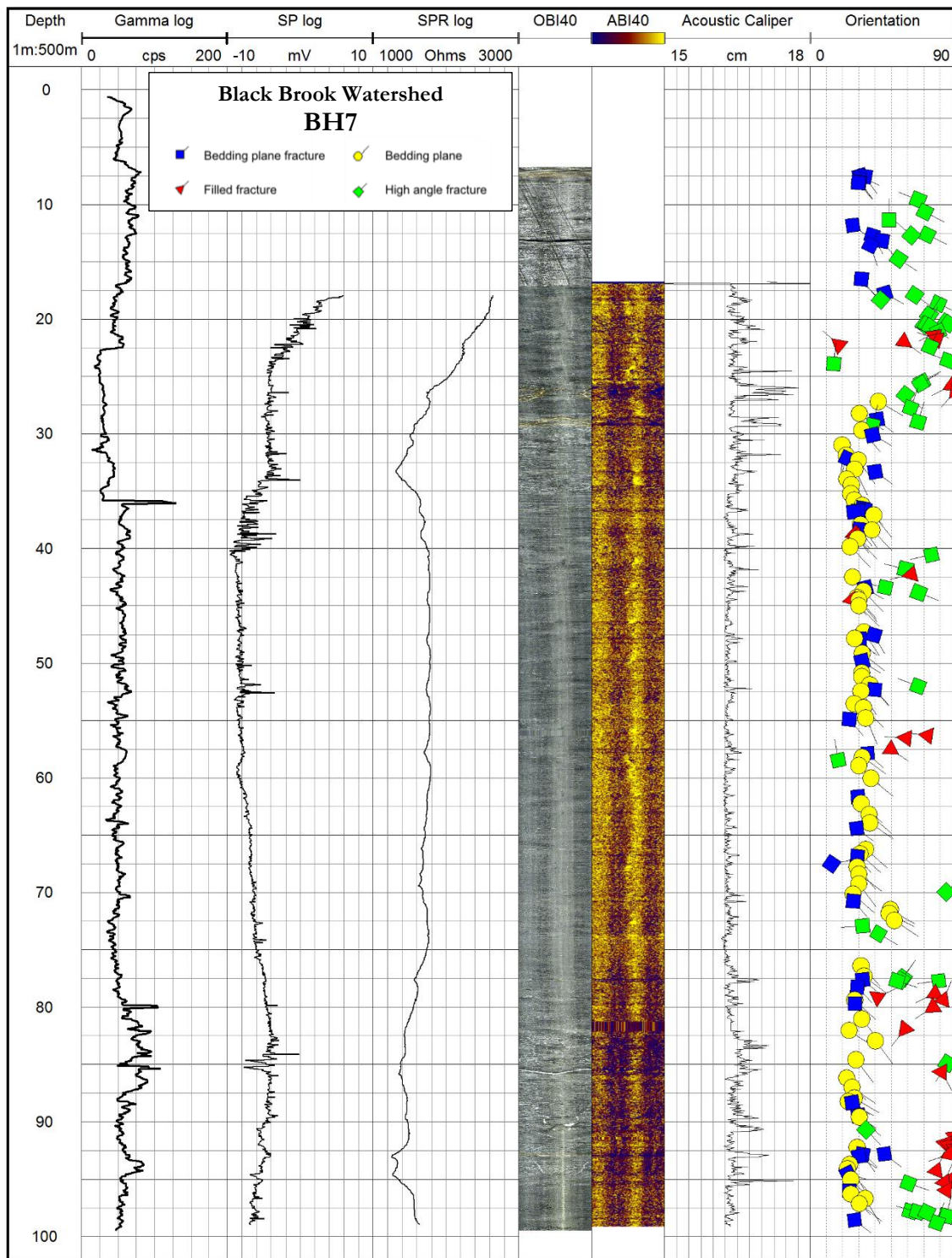
Appendix B:

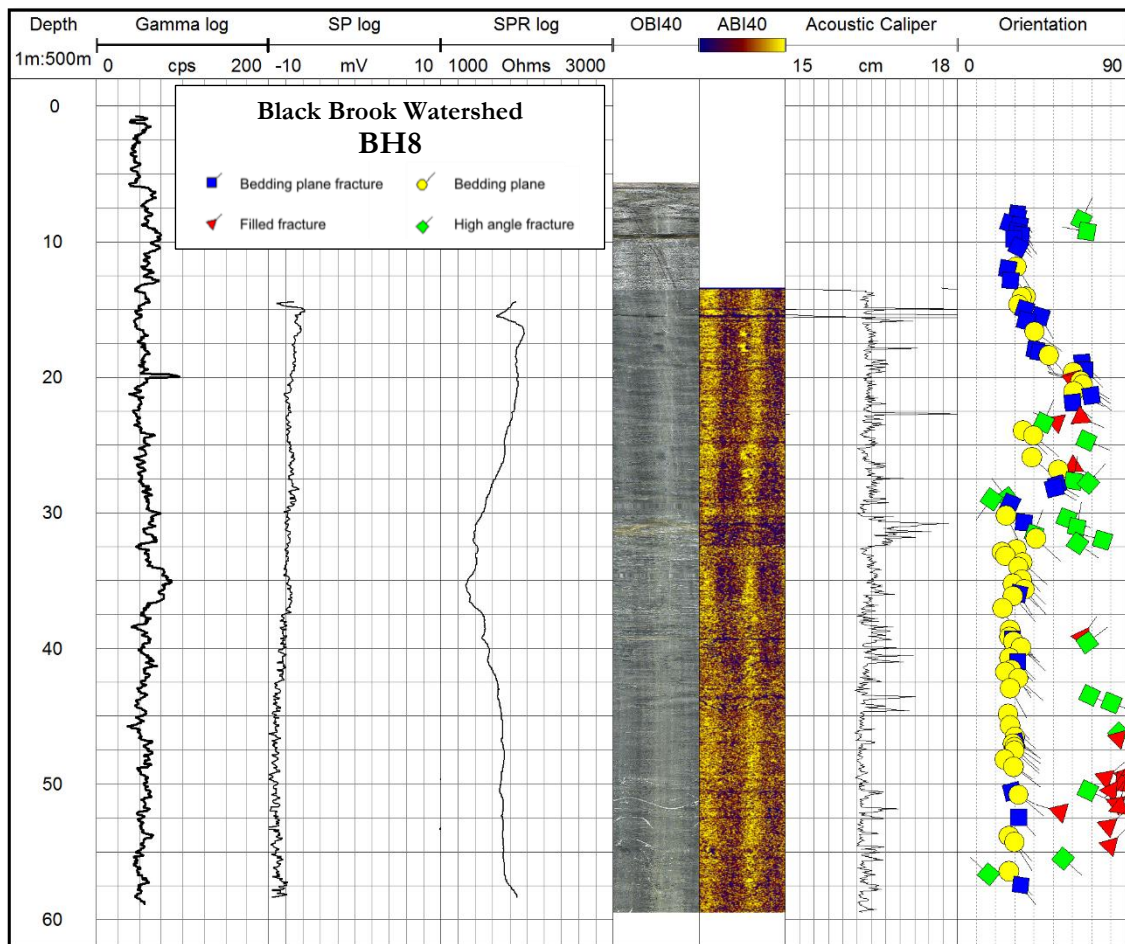
Compilation of borehole geophysical logs recorded within the Black Brook Watershed

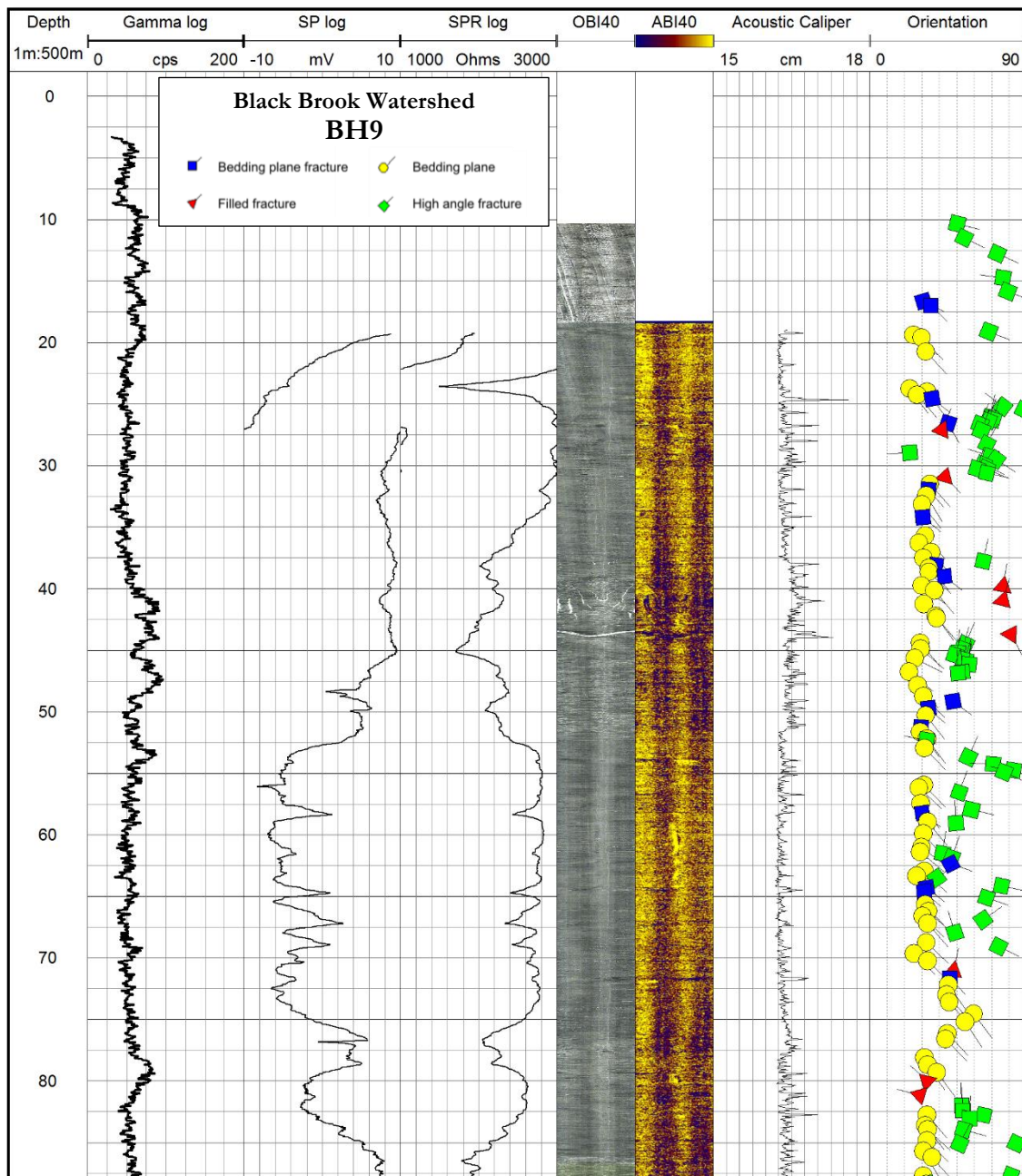
Borehole	X coordinate* (m)	Y coordinate* (m)	Depth to Bedrock (m)	Total Depth (m)	Top of Well Casing Elevation (m)
BH7	2404484	7568630	2.1	100.0	254.5
BH8	2404490	7568637	3.4	60.1	252.4
BH9	2404526	7568750	2.7	141.1	238.1
BH10	2404858	7569029	5.2	70.1	246.6
BH11	2404701	7568894	1.8	40.2	227.7

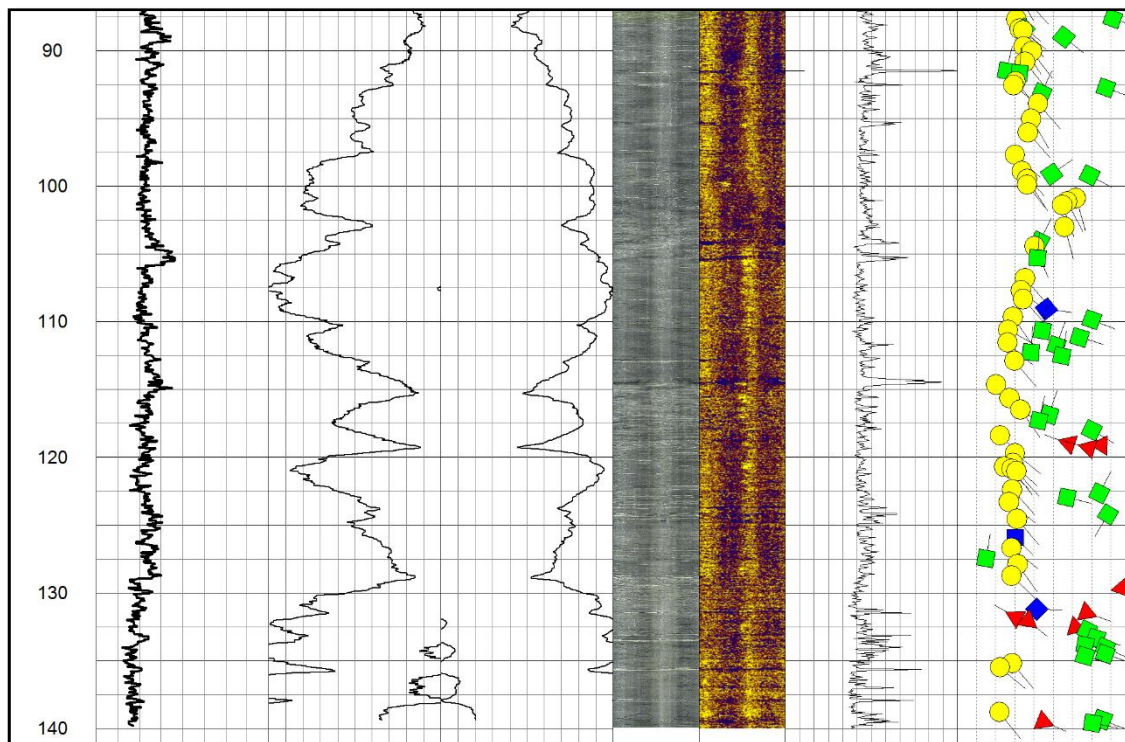
* NAD83 (CSRS) NB Double Stereographic projection

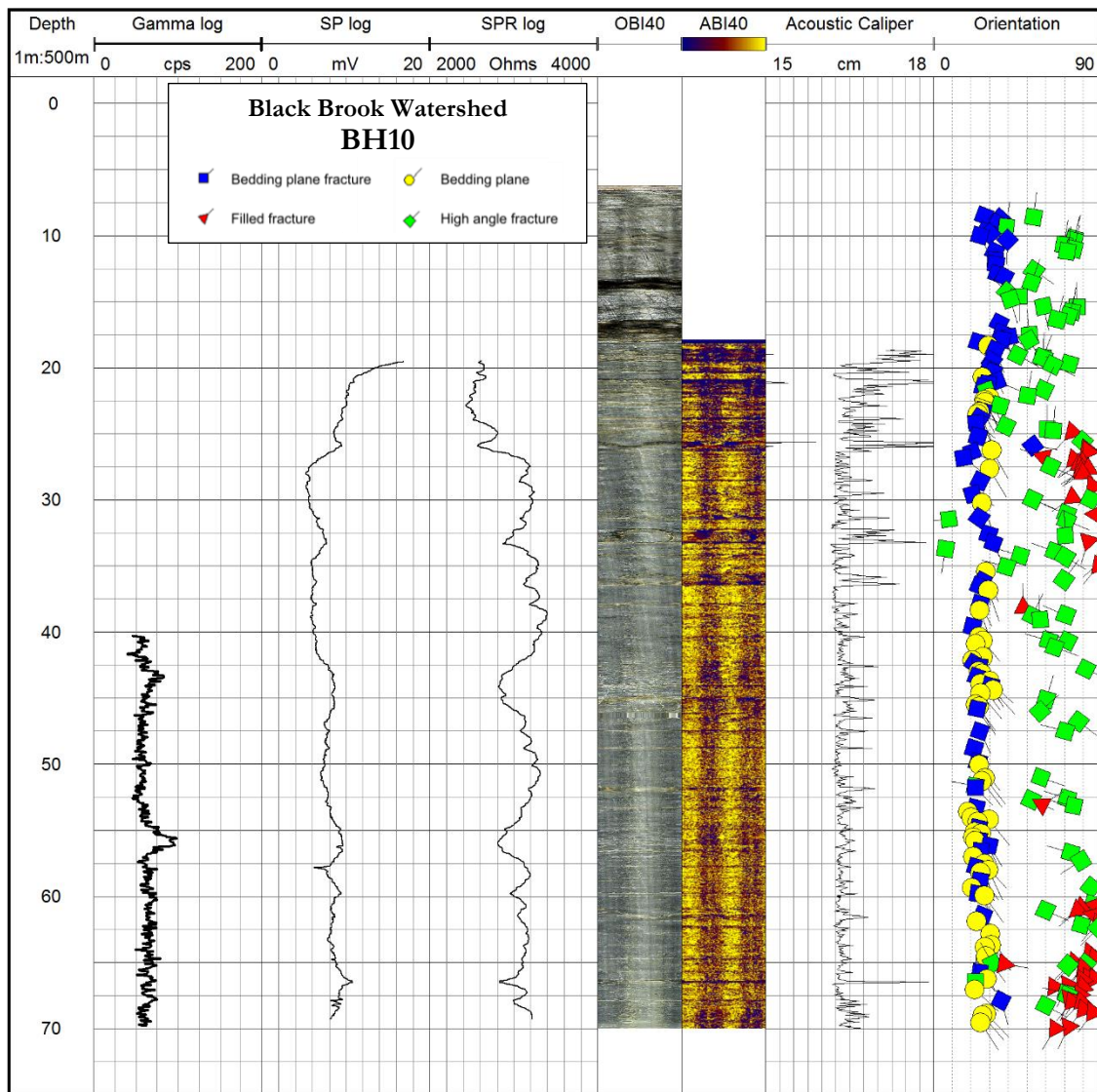
Borehole Geophysical Probe	Description
2PGA-1000	Measurement of gamma log, single point resistance log, and self-potential
ALT OBI-40	Optical televiewer imaging
ALT ABI-40	Acoustic televiewer imaging

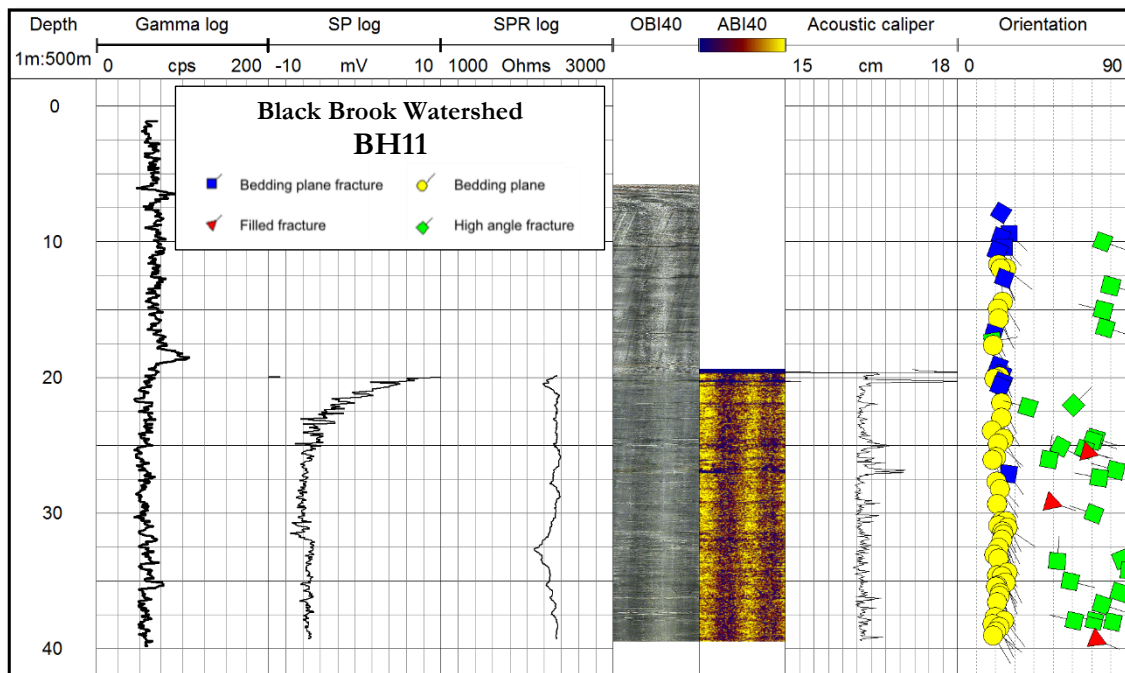












Appendix C:

**Springdale Wellfield fractures interpreted from televiewer
images**

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH07-1	25.53	240.3	66.66	4.36	High-angle fracture
BH07-1	25.55	141.4	84.93	4.07	High-angle fracture
BH07-1	25.56	160.06	70.23	4.74	High-angle fracture
BH07-1	25.57	259.26	64.62	4.29	High-angle fracture
BH07-1	25.72	334.3	67.61	na	High-angle fracture
BH07-1	25.95	264.5	69.08	6.43	High-angle fracture
BH07-1	26.34	86.66	56.36	3.88	High-angle fracture
BH07-1	26.76	82.75	62.66	4.59	High-angle fracture
BH07-1	27.32	245.88	75.21	8.93	High-angle fracture
BH07-1	27.64	236.44	66.54	na	High-angle fracture
BH07-1	27.64	246.36	58.75	9.34	High-angle fracture
BH07-1	27.79	263.02	67.39	na	High-angle fracture
BH07-1	27.9	107.9	75.93	1.95	High-angle fracture
BH07-1	27.99	266.49	64.22	na	High-angle fracture
BH07-1	28.37	212.32	69.02	2.51	High-angle fracture
BH07-1	28.66	240.25	73.88	1.94	High-angle fracture
BH07-1	28.75	109.68	52.02	6.77	High-angle fracture
BH07-1	29.14	34.42	56.34	13.86	High-angle fracture
BH07-1	29.18	236.25	61.21	8.67	High-angle fracture
BH07-1	30.25	120.64	68.71	7.62	High-angle fracture
BH07-1	30.59	18.97	14.1	38.98	Bedding plane fracture
BH07-1	31.17	97.34	17.64	na	Bedding plane fracture
BH07-1	31.5	275.87	68.74	6.16	High-angle fracture
BH07-1	31.72	275.43	58.22	7.37	High-angle fracture
BH07-1	32.26	261.21	59.1	na	High-angle fracture
BH07-1	32.31	259.07	61.94	9.88	High-angle fracture
BH07-1	32.43	151.46	67.41	na	High-angle fracture
BH07-1	32.58	247.51	50.02	na	High-angle fracture
BH07-1	32.65	156.3	0.13	na	Bedding plane fracture
BH07-1	33.11	316.49	30.08	na	Bedding plane fracture
BH07-1	33.19	349.7	85.23	2.58	High-angle fracture
BH07-1	33.3	319.05	36.11	na	Bedding plane fracture
BH07-1	33.6	355.46	76.29	4.27	High-angle fracture
BH07-1	33.65	29.77	55.4	7.95	High-angle fracture
BH07-1	35.52	290.5	60.08	na	High-angle fracture
BH07-1	35.52	238.01	69.5	3.85	High-angle fracture
BH07-1	35.84	55.17	45.81	7.67	High-angle fracture
BH07-1	36.09	328.77	63.72	6.2	High-angle fracture
BH07-1	36.32	311.99	70.43	5.02	High-angle fracture
BH07-1	36.71	309.5	72.8	4.14	High-angle fracture
BH07-1	36.77	319.76	64.74	na	High-angle fracture
BH07-1	36.79	313.61	74.34	4.59	High-angle fracture
BH07-1	36.87	316.31	67.89	na	High-angle fracture
BH07-1	37.08	313.53	73.82	na	High-angle fracture
BH07-1	37.08	306.36	75.66	na	High-angle fracture

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH07-1	37.46	147.71	81.66	8.27	High-angle fracture
BH07-1	38.16	151.02	17.84	131.46	Bedding plane fracture
BH07-1	38.98	279.09	80.29	12.48	High-angle fracture
BH07-1	39.89	252.01	60.87	6.82	High-angle fracture
BH07-1	39.94	226.12	66.59	4.37	High-angle fracture
BH07-1	39.95	31.38	80.89	3.96	High-angle fracture
BH07-1	41.51	266.65	46.52	9.63	High-angle fracture
BH07-1	41.57	54.7	61.39	8.62	High-angle fracture
BH07-2	24.81	343.93	23.32	19.28	High-angle fracture
BH07-2	25.23	333.7	23.17	42.29	High-angle fracture
BH07-2	26.08	289.71	9.69	6.9	Bedding plane fracture
BH07-2	26.28	143.47	34.67	11.52	High-angle fracture
BH07-2	26.44	273.96	12.65	na	Bedding plane fracture
BH07-2	26.67	309.38	6.89	na	Bedding plane fracture
BH07-2	26.88	346.31	7.98	10.89	Bedding plane fracture
BH07-2	27.08	80.74	4.15	na	Bedding plane fracture
BH07-2	27.22	96.54	6.39	na	Bedding plane fracture
BH07-2	27.34	64.66	6.14	na	Bedding plane fracture
BH07-2	27.56	89.43	47.33	na	High-angle fracture
BH07-2	27.58	263.41	64.23	6.09	High-angle fracture
BH07-2	27.61	274.55	67.35	na	High-angle fracture
BH07-2	27.66	291.43	69.35	7.41	High-angle fracture
BH07-2	27.82	118.92	9.13	na	Bedding plane fracture
BH07-2	27.94	305.67	40.52	34.97	High-angle fracture
BH07-2	28.5	120.99	17.58	na	Bedding plane fracture
BH07-2	29.1	338.03	9.43	10.85	Bedding plane fracture
BH07-2	29.29	51.78	20.09	9.39	Bedding plane fracture
BH07-2	29.32	251.36	42.07	5.2	High-angle fracture
BH07-2	30.11	353.63	34.98	22.97	High-angle fracture
BH07-2	30.2	160.95	4.34	13.96	Bedding plane fracture
BH07-2	31.02	277.2	8.97	16.79	Bedding plane fracture
BH07-2	32.69	91.44	37.54	5.55	High-angle fracture
BH07-2	32.83	97.04	22.53	9.24	High-angle fracture
BH07-2	33.15	77.8	66.78	7.1	High-angle fracture
BH07-2	33.96	352.19	63.83	9.26	High-angle fracture
BH07-2	34.86	277.37	53.66	8.3	High-angle fracture
BH07-2	35.12	205.26	73.09	4.95	High-angle fracture
BH08-1	19.43	246	82.83	1.75	High-angle fracture
BH08-1	19.61	237.4	81.39	2.1	High-angle fracture
BH08-1	19.68	234.35	83.11	2.16	High-angle fracture
BH08-1	20.17	15.03	20.24	9.38	Bedding plane fracture
BH08-1	20.33	299.99	80.31	3.03	High-angle fracture
BH08-1	22.31	115.4	33.16	9.21	Bedding plane fracture
BH08-1	22.52	271.16	78.9	3.46	High-angle fracture
BH08-1	22.7	319.8	88.13	na	High-angle fracture
BH08-1	22.99	355.01	75.36	na	High-angle fracture
BH08-1	23.1	331.86	62.09	na	High-angle fracture

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH08-1	23.3	313.36	85.99	1.05	High-angle fracture
BH08-1	23.43	346.04	69.94	6.18	High-angle fracture
BH08-1	23.51	223.47	86.22	na	High-angle fracture
BH08-1	23.68	358.31	40.68	6.07	Bedding plane fracture
BH08-1	23.87	347.85	31.46	5.97	Bedding plane fracture
BH08-1	23.95	304.37	59.44	3.56	High-angle fracture
BH08-1	24.22	33.69	26.21	6.28	Bedding plane fracture
BH08-1	24.33	177.55	54.65	na	High-angle fracture
BH08-1	24.51	214.47	65.93	2.86	High-angle fracture
BH08-1	24.53	313.58	12.39	9.77	Bedding plane fracture
BH08-1	24.54	161.62	72.61	2.99	High-angle fracture
BH08-1	24.94	229.1	76.49	3.97	High-angle fracture
BH08-1	25.73	72.29	34.27	5.78	Bedding plane fracture
BH08-1	26.07	37.91	20.78	10.28	Bedding plane fracture
BH08-1	27.23	89.67	22.74	143.74	Bedding plane fracture
BH08-1	27.59	59.8	17.21	86.83	Bedding plane fracture
BH08-1	28.04	330.07	86.09	5.11	High-angle fracture
BH08-1	28.13	106.57	74.83	12.04	High-angle fracture
BH08-1	28.49	110.93	76.22	3.34	High-angle fracture
BH08-1	29.48	73.05	15.97	2.88	Bedding plane fracture
BH08-1	29.8	320.63	21.7	15.8	Bedding plane fracture
BH08-1	30.15	232.18	79.23	5.79	High-angle fracture
BH08-1	31.34	76.32	60.37	na	High-angle fracture
BH08-1	31.55	163.12	3.97	na	Bedding plane fracture
BH08-1	31.92	129.52	55.73	3.94	High-angle fracture
BH08-1	31.97	155.47	63.01	3.18	High-angle fracture
BH08-1	32.73	0.04	18.54	13.27	Bedding plane fracture
BH08-1	33.2	285.04	86.38	4.42	High-angle fracture
BH08-1	33.27	125.51	25.51	na	Bedding plane fracture
BH08-1	33.58	325.44	85.23	3.49	High-angle fracture
BH08-1	33.77	248.17	85.71	5.01	High-angle fracture
BH08-1	34.29	95.83	11.63	13.71	Bedding plane fracture
BH08-1	34.63	65.76	81.31	3.17	High-angle fracture
BH08-1	34.63	227.59	79.13	1.32	High-angle fracture
BH08-1	35.21	325.62	86	1.47	High-angle fracture
BH08-1	36.08	44.14	82.43	na	High-angle fracture
BH08-1	36.35	135.66	7.13	13.89	Bedding plane fracture
BH08-1	37	310.5	38.69	5.46	Bedding plane fracture
BH08-1	37.56	296.13	59.59	7.09	High-angle fracture
BH08-1	37.83	96.63	72.95	7.04	High-angle fracture
BH08-1	38.12	89.35	73.69	5.9	High-angle fracture
BH08-1	38.22	258.99	70.28	7.09	High-angle fracture
BH08-1	39.22	337.93	22.22	na	Bedding plane fracture
BH08-1	39.34	247.01	89.18	1.66	High-angle fracture
BH08-1	39.78	97.18	30.57	65.92	Bedding plane fracture
BH08-1	40.05	336.74	76.61	na	High-angle fracture
BH08-1	40.36	40.73	12.13	6.84	Bedding plane fracture

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH08-1	40.4	340.23	77.46	na	High-angle fracture
BH08-1	40.5	283.1	75.77	na	High-angle fracture
BH08-1	42.15	131.86	84.89	2.5	High-angle fracture
BH08-1	42.21	210.15	54.39	na	High-angle fracture
BH08-1	42.31	334.56	83.15	2.51	High-angle fracture
BH08-2	26.52	151.76	58.06	7.41	High-angle fracture
BH08-2	26.56	302.65	23.86	na	Bedding plane fracture
BH08-2	26.58	347.48	73.11	5.23	High-angle fracture
BH08-2	27.27	0.37	69.46	3.86	High-angle fracture
BH08-2	27.27	97.47	7.33	10.64	Bedding plane fracture
BH08-2	27.45	56.3	5.06	31.5	Bedding plane fracture
BH08-2	27.53	149.63	71.25	4.12	High-angle fracture
BH08-2	27.74	355.23	72.21	4.84	High-angle fracture
BH08-2	28.71	311.29	28.2	9.69	High-angle fracture
BH08-2	28.94	336.68	63.04	4.13	High-angle fracture
BH08-2	29.68	22.46	58.72	5.71	High-angle fracture
BH08-2	29.71	340.85	72.46	4.22	High-angle fracture
BH08-2	29.73	350.01	17.7	na	Bedding plane fracture
BH08-2	30.08	346.39	67.49	11.87	High-angle fracture
BH08-2	30.57	171.58	65.84	29.06	High-angle fracture
BH08-2	30.65	279.21	44.86	52.46	High-angle fracture
BH08-2	30.74	203.88	48.82	16.46	High-angle fracture
BH08-2	31.19	31.03	36.46	12.06	High-angle fracture
BH08-2	31.5	245.22	50.23	13.43	High-angle fracture
BH08-2	31.53	153.84	87.86	2.24	High-angle fracture
BH08-2	32.65	82.32	8.72	74.82	Bedding plane fracture
BH08-2	33.06	330.95	89.58	0.68	High-angle fracture
BH08-2	35.95	26.54	11.04	14.72	Bedding plane fracture
BH08-2	36.72	328.91	84.96	9.57	High-angle fracture
BH08-2	37.31	263.02	26.76	3.57	Bedding plane fracture
BH08-2	37.68	39.88	34.73	13.97	High-angle fracture
BH08-2	37.85	210.51	37.27	7.96	High-angle fracture
BH08-2	37.93	291.14	4.15	6.98	Bedding plane fracture
BH08-2	38.35	279.47	56.26	na	High-angle fracture
BH08-2	39.19	325.21	70.51	6	High-angle fracture
BH08-2	39.31	328.64	72.04	3.08	High-angle fracture
BH08-2	40.01	331.46	68.24	8.9	High-angle fracture
BH08-2	40.01	178.63	30.54	12.06	High-angle fracture
BH08-2	40.32	310.62	64.39	6.05	High-angle fracture
BH08-2	40.74	37.46	51.06	4.4	High-angle fracture
BH08-2	40.87	197.12	27.17	16.01	High-angle fracture
BH08-2	41.69	84.15	12.56	6.83	Bedding plane fracture
BH08-2	42.15	321.79	73.19	3.47	High-angle fracture
BH08-2	42.17	343.66	71.43	4.46	High-angle fracture
BH08-2	43.58	343.31	61.6	9.99	High-angle fracture
BH08-2	43.7	346.99	68.24	3.71	High-angle fracture
BH08-2	43.73	235.96	67.66	8.36	High-angle fracture

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH08-2	45.86	320.96	73.95	2.77	High-angle fracture
BH08-2	46.57	0.71	79.47	5.85	High-angle fracture
BH08-2	46.92	123.15	71.83	5.61	High-angle fracture
BH08-2	47.42	348.75	36.74	16.83	High-angle fracture
BH08-2	48.66	68.78	18.56	na	Bedding plane fracture
BH08-2	50.09	69.68	43.14	100.64	High-angle fracture
BH08-2	52.13	43.8	1.33	43	Bedding plane fracture
BH08-2	57.02	178.43	68.17	5.21	High-angle fracture
BH08-2	57.03	278.49	29.41	14.81	Bedding plane fracture
BH08-2	58.57	142.73	76.64	4.85	High-angle fracture
BH09-2	21.69	349.56	26.59	281.57	Bedding plane fracture
BH09-2	23.82	55.36	81.92	33.6	High-angle fracture
BH09-2	25.4	172.68	79.17	4.13	High-angle fracture
BH09-2	25.54	23.83	43.07	na	High-angle fracture
BH09-2	27.21	177.4	78.04	2.9	High-angle fracture
BH09-2	27.38	292.12	80.45	3.49	High-angle fracture
BH09-2	28.26	169.3	72.48	4.22	High-angle fracture
BH09-2	28.47	158.61	84.73	3.58	High-angle fracture
BH09-2	28.7	178.67	69.72	9.71	High-angle fracture
BH09-2	30.39	336.87	82.74	4.04	High-angle fracture
BH09-2	30.47	82.13	76.2	10.97	High-angle fracture
BH09-2	32.52	349.71	79.79	4.43	High-angle fracture
BH09-2	33.11	205.88	30.68	390.26	Bedding plane fracture
BH09-2	33.7	10.35	75.21	6.12	High-angle fracture
BH09-2	33.89	114.26	58.73	na	Bedding plane fracture
BH09-2	33.96	254.54	81.47	5.78	High-angle fracture
BH09-2	34.14	119.74	77.28	6.17	High-angle fracture
BH09-2	34.94	277.71	80.26	3.55	High-angle fracture
BH09-2	35.5	287.56	72.63	7.46	High-angle fracture
BH09-2	36.98	345.68	60.87	6.82	High-angle fracture
BH09-2	38	330.02	34.47	11.54	Bedding plane fracture
BH09-2	40.38	346.32	42	10.4	Bedding plane fracture
BH09-2	43.38	79.44	77.43	15.25	High-angle fracture
BH09-2	45.05	82.47	20.3	21.38	Bedding plane fracture
BH09-2	45.86	286.48	40.59	6.43	High-angle fracture
BH09-2	46.1	272.39	57.14	5.22	High-angle fracture
BH09-2	47.42	24.16	17.64	9.53	Bedding plane fracture
BH09-2	47.5	351.03	30.94	8.58	Bedding plane fracture
BH09-2	48.22	130.36	18.85	18.43	Bedding plane fracture
BH09-2	48.69	359.2	65.69	6.28	High-angle fracture
BH09-2	50.24	220.77	21.55	48.15	Bedding plane fracture
BH09-2	52.33	88.61	11.71	136.28	Bedding plane fracture
BH09-2	53.4	351.83	12.58	40.27	Bedding plane fracture
BH09-2	54.76	264.91	35.61	70.55	Bedding plane fracture
BH09-2	55.4	148.54	10.83	7.86	Bedding plane fracture
BH09-2	55.76	232.23	87.82	5.5	High-angle fracture
BH09-2	56.42	93.04	31.24	3.42	Bedding plane fracture

Borehole ID	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH09-2	56.92	48.72	77.29	17.28	High-angle fracture
BH09-2	57.98	226.26	58.83	46.9	High-angle fracture
BH09-2	58.6	120.57	12.26	10.75	Bedding plane fracture
BH09-2	60.11	313.4	13.97	na	Bedding plane fracture
BH09-2	60.51	125.55	43.88	10.09	Bedding plane fracture
BH09-2	62.39	336.64	20.33	6.56	Bedding plane fracture
BH09-2	64.83	14.81	7.9	510.29	Bedding plane fracture
BH09-2	65.93	264.7	62.58	4.61	High-angle fracture
BH09-2	66.82	305.25	65.91	7.35	High-angle fracture
BH09-2	67.62	356.05	28.26	7.05	Bedding plane fracture
BH09-2	67.99	282.54	55.47	5.67	Bedding plane fracture
BH09-2	68.22	310.22	61.69	8.06	High-angle fracture
BH09-2	68.39	278.97	37.71	7.91	Bedding plane fracture
BH09-2	69.21	311.35	25.42	9.93	Bedding plane fracture
BH09-2	69.34	132.8	0.81	25	Bedding plane fracture
BH09-2	69.54	240.97	33.96	17.42	Bedding plane fracture
BH09-2	71.11	87.94	17.13	na	Bedding plane fracture
BH09-2	71.64	352.87	13.79	10.68	Bedding plane fracture
BH09-2	71.69	88.69	85.39	1.13	High-angle fracture
BH09-2	71.79	229.67	84	1.46	High-angle fracture
BH09-2	72.21	359.12	38.48	10.96	Bedding plane fracture
BH09-2	72.25	279.89	70	3.76	High-angle fracture
BH09-2	72.32	26.3	73.89	3.88	High-angle fracture
BH09-2	73.7	210.87	35.86	6.49	Bedding plane fracture
BH09-2	73.76	156.19	37.77	8.69	Bedding plane fracture
BH09-2	74.26	313.88	77.13	1.56	High-angle fracture
BH09-2	74.38	325.71	48.16	4.67	Bedding plane fracture
BH09-2	75.83	228.33	76.77	3.2	High-angle fracture
BH09-2	77.78	26.03	82.49	6.93	High-angle fracture
BH09-2	77.78	7.99	66.57	7.95	High-angle fracture

Appendix D:

Black Brook Watershed fractures interpreted from televiewer images

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH7	7.49	117.24	30.94	15.88	bedding plane fracture
BH7	7.55	143.92	33.62	0.1	bedding plane fracture
BH7	7.77	147.81	30.12	0.1	bedding plane fracture
BH7	8.04	141.86	30.34	3.02	bedding plane fracture
BH7	8.12	140.28	29.93	0.1	bedding plane fracture
BH7	9.59	291.45	66.62	0.1	high-angle fracture
BH7	10.64	117.64	70.82	3.82	high-angle fracture
BH7	11.33	359.44	48.68	3.37	high-angle fracture
BH7	11.81	127.33	26.27	0.1	bedding plane fracture
BH7	12.67	116.44	72.24	0.1	high-angle fracture
BH7	12.7	150.26	38.33	6.82	bedding plane fracture
BH7	12.72	311.59	62.51	3.4	high-angle fracture
BH7	13.14	143.16	44.2	49.74	bedding plane fracture
BH7	13.54	161.49	37.16	4.45	bedding plane fracture
BH7	14.73	123.73	54.53	4.48	high-angle fracture
BH7	16.49	140.26	31.55	7.85	bedding plane fracture
BH7	17.76	119.54	46.1	0.1	bedding plane fracture
BH7	17.91	124.92	64.55	0.1	high-angle fracture
BH7	18.29	312.61	43.73	3.54	high-angle fracture
BH7	18.66	114.87	79.18	0.1	high-angle fracture
BH7	19.64	33.18	73.43	0.1	high-angle fracture
BH7	20.25	123.19	84.11	0.1	high-angle fracture
BH7	20.38	128.71	85.35	1.57	high-angle fracture
BH7	20.5	128.52	70.51	0.1	high-angle fracture
BH7	20.52	32.17	71.86	0.1	high-angle fracture
BH7	20.68	122.61	74.38	0.1	high-angle fracture
BH7	21.11	245.1	79.09	0.1	high-angle fracture
BH7	21.41	291.81	76.57	0.1	filled fracture
BH7	21.75	289.14	78.38	0.1	filled fracture
BH7	21.97	124.92	57.69	0.09	filled fracture
BH7	22.23	310.93	17.83	0.1	filled fracture
BH7	22.42	25.94	74.16	0.1	high-angle fracture
BH7	23.59	289.6	85.15	0.1	high-angle fracture
BH7	23.89	6.74	14.46	6.14	high-angle fracture
BH7	25.43	61.77	69.06	1.08	high-angle fracture
BH7	25.63	59.96	68.13	1.84	high-angle fracture
BH7	25.81	243.92	87.3	0.1	filled fracture
BH7	26.24	220.7	89.68	0.1	filled fracture
BH7	26.58	129.47	58.79	0.1	high-angle fracture
BH7	27.73	19.82	62.16	0.1	high-angle fracture
BH7	28.75	127.1	41.19	4.21	bedding plane fracture
BH7	28.93	17.18	66.76	6.15	high-angle fracture
BH7	29.24	12.98	38.65	5.78	high-angle fracture
BH7	30.11	120.84	38.45	4.36	bedding plane fracture
BH7	32.15	157.68	22.54	0.1	bedding plane fracture
BH7	33.31	125.95	40.14	4.31	bedding plane fracture
BH7	36.48	144.97	32.35	0.1	bedding plane fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH7	36.67	140.04	33.79	5.15	bedding plane fracture
BH7	36.74	136.39	29.5	0.1	bedding plane fracture
BH7	36.79	131.99	26.95	0.1	bedding plane fracture
BH7	38.28	142.44	30.29	0.1	bedding plane fracture
BH7	38.76	132.07	27.08	0.1	filled fracture
BH7	38.89	142.81	27.99	0.1	filled fracture
BH7	40.55	257.38	74.63	2.71	high-angle fracture
BH7	41.82	283.83	58.65	2.5	high-angle fracture
BH7	42.25	256.95	62.36	0.1	filled fracture
BH7	43.39	281.44	46.27	0.1	high-angle fracture
BH7	43.44	120.91	34.05	0.1	bedding plane fracture
BH7	43.87	111.13	67.14	2.71	high-angle fracture
BH7	44.45	136.59	25.95	5.81	filled fracture
BH7	47.53	150.83	39.33	0.1	bedding plane fracture
BH7	47.86	134.43	30.51	0.1	bedding plane fracture
BH7	49.87	121.06	31.78	0.1	bedding plane fracture
BH7	52.01	290.69	66.78	0.1	high-angle fracture
BH7	52.28	139.69	39.71	0.1	bedding plane fracture
BH7	54.92	139.03	24.13	0.1	bedding plane fracture
BH7	56.29	278.13	72.49	0.1	filled fracture
BH7	56.57	275.11	59.04	0.1	filled fracture
BH7	57.52	122.98	49.67	0.97	filled fracture
BH7	57.83	132.63	35.54	5.01	bedding plane fracture
BH7	58.44	348.66	17.4	5.25	high-angle fracture
BH7	61.63	130.28	29.58	0.1	bedding plane fracture
BH7	64.4	127.4	28.77	0.1	bedding plane fracture
BH7	66.85	138.87	29.09	0.1	bedding plane fracture
BH7	67.48	79.6	13.18	0.1	bedding plane fracture
BH7	69.96	44.85	83.92	2.78	high-angle fracture
BH7	70.77	138.17	26.7	5.79	bedding plane fracture
BH7	72.88	265.98	32.2	0.1	high-angle fracture
BH7	73.59	120.7	42.34	4.07	high-angle fracture
BH7	77.45	35.54	57.39	0.1	high-angle fracture
BH7	77.58	36.4	56.05	0.1	high-angle fracture
BH7	77.59	129.75	32.09	5.17	bedding plane fracture
BH7	77.68	262.86	79.52	0.1	high-angle fracture
BH7	77.72	108.29	53.43	0.1	high-angle fracture
BH7	78.25	137.51	29.23	0.1	bedding plane fracture
BH7	78.82	243.62	77.07	0.1	filled fracture
BH7	79.17	64.33	41.53	0.1	filled fracture
BH7	79.41	21.97	82.37	0.1	filled fracture
BH7	79.72	135.72	27.71	0.1	bedding plane fracture
BH7	79.98	237.15	75.7	0.1	filled fracture
BH7	81.91	218.42	58.39	0.1	filled fracture
BH7	84.89	212.75	84.64	0.1	high-angle fracture
BH7	85.63	151.76	81.29	26.78	filled fracture
BH7	88.3	126.22	25.87	0.1	bedding plane fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH7	89.46	125.37	29.7	0.1	bedding plane fracture
BH7	90.71	134.4	34.41	0.1	high-angle fracture
BH7	91.32	54.15	88.57	3.98	filled fracture
BH7	91.86	45.13	89.48	0.1	filled fracture
BH7	91.89	41.97	83.38	0.1	filled fracture
BH7	92.07	52.54	85.83	0.1	filled fracture
BH7	92.29	55.81	88.82	0.1	filled fracture
BH7	92.62	45.96	89.05	0.1	filled fracture
BH7	92.79	127.3	45.73	6.23	bedding plane fracture
BH7	92.85	252.28	85.69	0.1	filled fracture
BH7	92.87	143.8	32.28	0.1	bedding plane fracture
BH7	92.99	124.07	30.06	0.1	bedding plane fracture
BH7	94.32	22.41	77.96	0.1	filled fracture
BH7	94.51	106.4	23.13	0.1	bedding plane fracture
BH7	95.16	29.29	85.23	0.1	filled fracture
BH7	95.27	24.03	82.77	0.1	filled fracture
BH7	95.34	110.37	60.66	5.72	high-angle fracture
BH7	95.95	31.36	83.92	0.1	filled fracture
BH7	95.98	133.15	24.22	0.1	bedding plane fracture
BH7	97.82	108.11	61.69	0.1	high-angle fracture
BH7	97.86	285.01	66.09	0.1	high-angle fracture
BH7	97.9	114.26	71.87	0.1	high-angle fracture
BH7	98.19	14	84.06	0.1	high-angle fracture
BH7	98.53	127.7	27.17	7.94	bedding plane fracture
BH7	98.81	20.26	78.28	4.24	high-angle fracture
BH8	7.91	142.26	31.18	8.02	bedding plane fracture
BH8	8.15	146.4	31.16	5.11	bedding plane fracture
BH8	8.4	25.15	64.82	2.03	high-angle fracture
BH8	8.65	150.73	27.34	5.46	bedding plane fracture
BH8	8.82	145.79	32.27	5	bedding plane fracture
BH8	9.24	280.61	67.68	2.82	high-angle fracture
BH8	9.27	146.61	30.87	5.13	bedding plane fracture
BH8	9.52	140	33.15	4.92	bedding plane fracture
BH8	9.79	135.49	29.59	10.51	bedding plane fracture
BH8	10.4	162.98	31.8	2.97	bedding plane fracture
BH8	11.99	143.43	26.24	7.95	bedding plane fracture
BH8	12.85	139.89	27.54	5.44	bedding plane fracture
BH8	15.02	153.69	35.21	7.95	bedding plane fracture
BH8	15.51	150.3	43.19	20.61	bedding plane fracture
BH8	15.87	149.26	35.95	7.04	bedding plane fracture
BH8	17.93	144.96	40.75	4.55	bedding plane fracture
BH8	18.13	142.31	42.41	6.86	bedding plane fracture
BH8	18.89	132.46	65.07	3.53	bedding plane fracture
BH8	19.43	133.92	66.58	2.52	bedding plane fracture
BH8	20.03	283.92	60.12	0.1	filled fracture
BH8	20.22	287.27	60.16	0.1	filled fracture
BH8	21.37	125.75	69.93	3.45	bedding plane fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH8	21.84	129.48	60.33	3.99	bedding plane fracture
BH8	22.94	114.08	64.48	2.87	filled fracture
BH8	23.28	290.96	51.86	0.1	filled fracture
BH8	23.33	23.84	44.9	0.1	high-angle fracture
BH8	24.66	112.52	67.44	1.73	high-angle fracture
BH8	26.57	116.3	60.66	0.1	filled fracture
BH8	27.64	8.59	60.63	0.1	high-angle fracture
BH8	27.78	40.22	68.26	0.1	high-angle fracture
BH8	27.91	119.47	52.15	3.09	bedding plane fracture
BH8	28.15	120.81	50.64	0.1	bedding plane fracture
BH8	28.9	322.1	25.89	0.1	high-angle fracture
BH8	28.99	216.39	17.87	6.67	high-angle fracture
BH8	29.28	159.19	27.49	5.84	bedding plane fracture
BH8	30.41	111.87	57.2	2.55	high-angle fracture
BH8	30.71	138.03	34.64	0.1	bedding plane fracture
BH8	31.07	11.15	62.43	4.15	high-angle fracture
BH8	31.53	19.97	39.93	0.1	high-angle fracture
BH8	31.99	285.53	75.68	2.06	high-angle fracture
BH8	32.26	125.63	62.79	6.14	high-angle fracture
BH8	35.93	148.14	31.89	5.54	bedding plane fracture
BH8	39.14	34.61	65.06	0.1	filled fracture
BH8	39.32	134.31	28.63	5.5	bedding plane fracture
BH8	39.55	53.51	67.7	2.64	high-angle fracture
BH8	40.96	136.11	31.47	5.37	bedding plane fracture
BH8	43.47	113.24	69.15	4.06	high-angle fracture
BH8	44.01	111.8	80.4	5.17	high-angle fracture
BH8	46.24	44.76	84.29	0.87	high-angle fracture
BH8	46.65	46.16	84.05	0.1	filled fracture
BH8	46.91	123.98	30.41	5.9	bedding plane fracture
BH8	49.36	41.86	87.42	0.1	filled fracture
BH8	49.56	50.48	77.07	0.1	filled fracture
BH8	49.62	37.85	89.51	0.1	filled fracture
BH8	49.87	215.2	85.8	0.1	filled fracture
BH8	50.14	244.09	88.86	0.1	filled fracture
BH8	50.5	113.16	68.04	2.51	high-angle fracture
BH8	50.52	33.25	80.05	0.1	filled fracture
BH8	50.55	120.14	28.78	5.23	bedding plane fracture
BH8	51.44	48.98	83.14	5.6	filled fracture
BH8	51.71	9.89	84.55	0.1	filled fracture
BH8	52.05	287.32	53.56	0.1	filled fracture
BH8	52.47	136.64	31.82	0.1	bedding plane fracture
BH8	53.11	50.55	78.75	2.36	filled fracture
BH8	54.57	44.42	79.51	2.1	filled fracture
BH8	55.48	131.91	55.24	5.64	high-angle fracture
BH8	56.65	312.34	16.37	9.53	high-angle fracture
BH8	57.43	142.32	32.9	5.26	bedding plane fracture
BH9	10.34	104.46	50.4	13.84	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH9	11.52	116.17	54.3	7.25	high-angle fracture
BH9	12.76	114.26	73.43	4.02	high-angle fracture
BH9	14.74	277.8	76.42	5.64	high-angle fracture
BH9	15.84	111.07	79.29	2.59	high-angle fracture
BH9	16.66	117.37	30.9	0.1	bedding plane fracture
BH9	16.98	133.83	35.07	14.34	bedding plane fracture
BH9	19.08	112.38	68.32	5.53	high-angle fracture
BH9	24.53	126.94	35.76	9.9	bedding plane fracture
BH9	25.19	306.77	76.58	0.1	high-angle fracture
BH9	25.44	117.58	88.44	0.1	high-angle fracture
BH9	26.03	11.91	69	0.1	high-angle fracture
BH9	26.25	18.13	70.81	0.1	high-angle fracture
BH9	26.33	19.24	67.94	3.95	high-angle fracture
BH9	26.56	150.25	45.17	6.87	bedding plane fracture
BH9	26.63	22.22	63.42	0.1	high-angle fracture
BH9	27.09	23.31	41.35	0.1	filled fracture
BH9	27.17	21.91	64.18	0.1	high-angle fracture
BH9	28.26	27.84	67.21	3.18	high-angle fracture
BH9	28.93	266.32	23.04	0.1	high-angle fracture
BH9	29.27	287.66	68.93	1.67	high-angle fracture
BH9	29.5	123.28	72.7	1.68	high-angle fracture
BH9	29.88	115.26	65.06	3.42	high-angle fracture
BH9	30.23	113.55	66.91	2.02	high-angle fracture
BH9	30.25	16.32	61.54	0.1	high-angle fracture
BH9	30.58	102.3	67.27	2.62	high-angle fracture
BH9	30.88	136.46	43.24	0.1	filled fracture
BH9	31.93	139.15	33.6	8.37	bedding plane fracture
BH9	34.18	128.68	30.58	5.46	bedding plane fracture
BH9	37.76	12.08	65.05	0.1	high-angle fracture
BH9	38.06	142.48	37.95	0.1	bedding plane fracture
BH9	38.98	127.39	42.82	0.1	bedding plane fracture
BH9	39.82	10.28	76.58	0.1	filled fracture
BH9	40.93	16.56	76.37	0.1	filled fracture
BH9	43.72	151.84	81.28	25.88	filled fracture
BH9	44.52	23.84	54.96	2.52	high-angle fracture
BH9	44.92	16.59	53	3.33	high-angle fracture
BH9	45.21	5.18	50.55	0.1	high-angle fracture
BH9	45.41	17.71	48.94	0.1	high-angle fracture
BH9	45.75	17.27	54.55	1.59	high-angle fracture
BH9	46.17	5.16	56.9	2.19	high-angle fracture
BH9	46.77	356.32	53.32	3.47	high-angle fracture
BH9	46.86	357.11	50.82	4.83	high-angle fracture
BH9	49.13	124.65	47.72	0.1	bedding plane fracture
BH9	49.71	132.08	33.71	5.33	bedding plane fracture
BH9	51.27	139.72	29.55	0.1	bedding plane fracture
BH9	52.3	280.56	32.47	0.1	high-angle fracture
BH9	53.71	23.95	56.75	1.21	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH9	54.23	272.57	70.84	0.1	high-angle fracture
BH9	54.76	278.88	82.63	0.1	high-angle fracture
BH9	54.92	115.46	77.32	0.1	high-angle fracture
BH9	56.55	19.76	51.47	2.63	high-angle fracture
BH9	57.93	103.91	58.52	2.63	high-angle fracture
BH9	58.19	127.75	30.09	0.1	bedding plane fracture
BH9	59.06	357.28	49.78	5.49	high-angle fracture
BH9	61.51	9.46	41.88	2.59	high-angle fracture
BH9	61.83	20.23	47.19	3.2	high-angle fracture
BH9	62.33	108.21	46.54	5.88	bedding plane fracture
BH9	63.5	54.62	38.27	0.1	high-angle fracture
BH9	64.17	102.63	75.96	1.77	high-angle fracture
BH9	64.29	127.15	32.51	0.1	bedding plane fracture
BH9	64.58	131.83	31.07	0.1	bedding plane fracture
BH9	65.12	108.82	66.91	0.1	high-angle fracture
BH9	66.9	53.71	65.03	1.7	high-angle fracture
BH9	67.97	342.63	49.05	0.1	high-angle fracture
BH9	69.07	116.64	74.15	0.1	high-angle fracture
BH9	71.05	11.42	48.12	0.1	filled fracture
BH9	71.66	135.86	45.97	4.52	bedding plane fracture
BH9	79.98	312.54	32.91	0.1	filled fracture
BH9	81.15	285.5	28.73	0.1	filled fracture
BH9	82.02	358.85	52.7	3.07	high-angle fracture
BH9	82.39	358.78	53.49	1.8	high-angle fracture
BH9	82.74	283.11	65.4	2.93	high-angle fracture
BH9	83.07	1.69	57.44	0.1	high-angle fracture
BH9	83.96	106.04	53.73	0.1	high-angle fracture
BH9	85.04	113.13	84.28	0.1	high-angle fracture
BH9	85.09	25.83	51.65	0.1	high-angle fracture
BH9	87.64	108.95	81.32	0.1	high-angle fracture
BH9	88.39	135.9	34.78	0.1	high-angle fracture
BH9	89	129.89	55.48	11.62	high-angle fracture
BH9	91.49	13.2	25.39	0.1	high-angle fracture
BH9	91.67	4.66	32.26	0.1	high-angle fracture
BH9	92.76	110.63	77.45	0.1	high-angle fracture
BH9	93.16	25.38	43.91	0.1	high-angle fracture
BH9	99.12	54.66	49.37	3.08	high-angle fracture
BH9	99.19	118.76	69.03	7.04	high-angle fracture
BH9	104.06	28	42.77	4.58	high-angle fracture
BH9	105.3	4.61	41.96	10.06	high-angle fracture
BH9	109.06	98.06	46.45	0.1	bedding plane fracture
BH9	109.87	110.9	70.27	2.92	high-angle fracture
BH9	110.66	8.68	44.44	0.1	high-angle fracture
BH9	111.15	106.21	63.61	3.52	high-angle fracture
BH9	111.76	20.01	51.81	4.81	high-angle fracture
BH9	112.26	5.13	38.29	0.1	high-angle fracture
BH9	112.51	11.14	54.48	4.01	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH9	116.86	18.32	47.68	7.93	high-angle fracture
BH9	117.3	13.39	42.49	6.36	high-angle fracture
BH9	118.02	119.4	70.32	3.1	high-angle fracture
BH9	118.95	287.93	57.96	0.1	filled fracture
BH9	119.09	272.13	75.16	0.1	filled fracture
BH9	119.3	284.95	68.66	0.1	filled fracture
BH9	122.64	28.57	74.18	0.1	high-angle fracture
BH9	122.96	103.27	57.11	6.94	high-angle fracture
BH9	124.19	210.98	78.71	0.1	high-angle fracture
BH9	125.9	134.15	30.2	0.1	bedding plane fracture
BH9	127.44	9.34	15.01	0.1	high-angle fracture
BH9	129.63	140.95	86.04	0.1	filled fracture
BH9	131.18	91.48	41.32	0.1	bedding plane fracture
BH9	131.45	109.41	67.35	0.1	filled fracture
BH9	131.8	299.18	30.43	0.1	filled fracture
BH9	132.03	129.28	36.68	0.1	filled fracture
BH9	132.52	105.91	61.7	0.1	filled fracture
BH9	132.76	116.78	67.73	0.1	high-angle fracture
BH9	133.34	105.85	72.42	0.1	high-angle fracture
BH9	133.85	101.46	66.77	7.6	high-angle fracture
BH9	134.05	114.64	77.27	2.98	high-angle fracture
BH9	134.49	109.4	77.21	2.13	high-angle fracture
BH9	134.63	106.67	66.74	0.1	high-angle fracture
BH9	139.33	111.28	75.53	0.1	high-angle fracture
BH9	139.48	111.59	44.63	0.1	filled fracture
BH9	139.62	100.54	70.47	0.1	high-angle fracture
BH10	8.51	155.95	26.96	0.1	bedding plane fracture
BH10	8.59	8.2	53.49	0.1	high-angle fracture
BH10	8.69	175.63	35.57	0.1	bedding plane fracture
BH10	9.04	166.27	33.94	4.89	bedding plane fracture
BH10	9.29	6.97	39.07	6.92	high-angle fracture
BH10	9.68	159.87	28.05	8.54	bedding plane fracture
BH10	9.86	179.09	29.99	0.1	bedding plane fracture
BH10	9.99	153.13	24.67	0.1	bedding plane fracture
BH10	10.19	16.24	75.28	8.5	high-angle fracture
BH10	10.3	180.86	39.81	0.1	bedding plane fracture
BH10	10.39	14.16	75.17	4.79	high-angle fracture
BH10	10.69	8.38	69.84	0.1	high-angle fracture
BH10	10.98	14.31	75.29	1.87	high-angle fracture
BH10	11.2	7.94	71.64	3.83	high-angle fracture
BH10	11.2	162.91	32.47	10.05	bedding plane fracture
BH10	11.86	140.12	32.81	7.85	bedding plane fracture
BH10	12.14	143.45	33.33	4.94	bedding plane fracture
BH10	12.66	124.17	53.74	4.34	high-angle fracture
BH10	12.88	154.71	34.05	7.66	bedding plane fracture
BH10	13.1	163.33	37.96	4.51	bedding plane fracture
BH10	13.57	19.39	52.42	59.95	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH10	14.27	312.56	39.2	4.39	high-angle fracture
BH10	14.58	178.44	45.88	5.37	high-angle fracture
BH10	14.78	163.9	41.1	0.1	high-angle fracture
BH10	15.36	167.36	58.84	2.58	high-angle fracture
BH10	15.39	358.45	76.8	0.1	high-angle fracture
BH10	15.74	9.32	74.18	3.11	high-angle fracture
BH10	16.06	16.1	72.27	0.1	high-angle fracture
BH10	16.37	14	66.29	0.1	high-angle fracture
BH10	16.58	162.14	35.08	12.29	bedding plane fracture
BH10	17.36	155.05	36.39	9.31	bedding plane fracture
BH10	17.51	354.18	51.32	11.63	high-angle fracture
BH10	17.66	121.57	40.7	8.49	bedding plane fracture
BH10	17.84	147.15	36.68	6.61	bedding plane fracture
BH10	17.88	148.69	50.89	3.29	high-angle fracture
BH10	18.03	150.3	23.83	11.68	bedding plane fracture
BH10	18.63	158.83	33.04	7.26	bedding plane fracture
BH10	19.07	118.98	44.87	3.91	high-angle fracture
BH10	19.21	3.43	58.39	2.8	high-angle fracture
BH10	19.27	21.15	58.57	4.85	high-angle fracture
BH10	19.51	168.97	31.15	10.51	bedding plane fracture
BH10	19.69	287.34	72.81	0.1	high-angle fracture
BH10	19.8	116.7	63.63	2.22	high-angle fracture
BH10	20.08	164.59	29.34	5.47	bedding plane fracture
BH10	20.28	154.76	31.8	0.1	bedding plane fracture
BH10	20.98	113.96	33.83	4.99	bedding plane fracture
BH10	21.2	145.8	29.22	5.48	bedding plane fracture
BH10	21.46	152.37	26.41	5.77	bedding plane fracture
BH10	21.65	23.82	59.05	6.85	high-angle fracture
BH10	21.72	159.42	28.45	0.1	high-angle fracture
BH10	22.08	8.57	50.41	3	high-angle fracture
BH10	22.83	288.26	35.35	0.1	high-angle fracture
BH10	23.24	150.89	26.13	5.84	bedding plane fracture
BH10	23.82	167.78	24.33	9.43	bedding plane fracture
BH10	24.17	129.34	22.75	0.1	bedding plane fracture
BH10	24.38	295.42	38.72	0.1	high-angle fracture
BH10	24.67	359.33	61.12	5.65	high-angle fracture
BH10	24.75	3.9	64.05	3.26	high-angle fracture
BH10	24.83	223.33	74.37	0.1	filled fracture
BH10	25.21	149.48	23.68	6.08	bedding plane fracture
BH10	25.54	131.3	79.84	0.1	high-angle fracture
BH10	25.88	98.39	53.11	17.81	bedding plane fracture
BH10	26.13	218.07	81.93	0.1	filled fracture
BH10	26.28	216.33	82.83	1.52	filled fracture
BH10	26.34	115.16	20.48	0.1	bedding plane fracture
BH10	26.76	288.11	58.72	0.1	filled fracture
BH10	26.81	126.85	15.84	6.57	bedding plane fracture
BH10	26.88	221.62	75.12	0.1	filled fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH10	27.13	238.46	79.72	0.1	filled fracture
BH10	27.28	236.03	78	0.1	filled fracture
BH10	27.45	222.98	79.36	0.1	filled fracture
BH10	27.45	113.28	62.66	2.46	high-angle fracture
BH10	27.67	239.1	83.06	0.1	filled fracture
BH10	27.81	238.12	79.51	0.1	filled fracture
BH10	28.01	237.99	80.14	0.1	filled fracture
BH10	28.55	163.85	24.77	6.1	bedding plane fracture
BH10	28.95	216.63	85.23	0.1	filled fracture
BH10	29.56	118.33	20.88	6.19	bedding plane fracture
BH10	29.81	234.33	74.56	0.1	filled fracture
BH10	29.93	118.59	83.84	1.98	high-angle fracture
BH10	29.95	112.8	53.17	3.26	high-angle fracture
BH10	31.05	291.76	71.19	4.59	high-angle fracture
BH10	31.11	31.84	86.83	0.1	filled fracture
BH10	31.36	169.74	24.61	6.18	bedding plane fracture
BH10	31.47	262.97	8.22	10.72	high-angle fracture
BH10	31.57	112.69	71.15	0.1	high-angle fracture
BH10	32.6	153.34	29.45	2.46	bedding plane fracture
BH10	32.7	357.11	70.5	3.2	high-angle fracture
BH10	33.09	196.61	82.81	0.1	filled fracture
BH10	33.24	154.62	31.85	11.2	bedding plane fracture
BH10	33.72	188.41	6.25	0.1	high-angle fracture
BH10	33.9	289.4	64.82	2.79	high-angle fracture
BH10	34.18	285.9	46.5	5.47	high-angle fracture
BH10	34.31	121.9	70.59	3.13	high-angle fracture
BH10	34.55	226.94	89.88	0.1	filled fracture
BH10	34.96	219.23	88.59	0.1	filled fracture
BH10	35.04	285.54	39.05	4.23	high-angle fracture
BH10	36.06	36.45	69.82	3.29	high-angle fracture
BH10	36.16	162.25	26.25	9.57	bedding plane fracture
BH10	36.48	160.8	24.45	8.87	bedding plane fracture
BH10	37.85	152.29	24.65	6.18	bedding plane fracture
BH10	38.15	358.03	48.08	0.1	high-angle fracture
BH10	38.72	291.89	70.84	0.1	high-angle fracture
BH10	38.74	26.18	52.94	2.36	high-angle fracture
BH10	39.06	357.24	57.09	0.1	high-angle fracture
BH10	39.62	148.77	20.8	7.35	bedding plane fracture
BH10	40.66	296.85	71.73	1.48	high-angle fracture
BH10	40.67	286.36	61.7	2.08	high-angle fracture
BH10	41.14	109.27	64.9	2.32	high-angle fracture
BH10	42.34	165.97	21.67	6.5	bedding plane fracture
BH10	42.61	141.74	24.5	7.07	bedding plane fracture
BH10	42.81	117.36	81.38	1.85	high-angle fracture
BH10	43.33	154.23	23.16	6.34	bedding plane fracture
BH10	44.11	148.79	30.43	5.76	bedding plane fracture
BH10	45.12	16.15	60.41	3.86	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH10	45.8	147.2	23.28	8.15	bedding plane fracture
BH10	46.01	49.56	56.96	1.82	high-angle fracture
BH10	46.74	127.85	77.67	2.61	high-angle fracture
BH10	47.51	154.54	24.68	9.84	bedding plane fracture
BH10	47.52	285.67	70.38	0.1	high-angle fracture
BH10	48.79	151.63	21.48	6.47	bedding plane fracture
BH10	49.83	152.01	23.78	0.1	bedding plane fracture
BH10	51.01	110.8	57.33	4.75	high-angle fracture
BH10	51.66	280.86	22.26	0.1	high-angle fracture
BH10	51.76	134.59	22.43	9.84	bedding plane fracture
BH10	52.61	288.72	71.61	2.36	high-angle fracture
BH10	52.69	113.04	52.13	0.1	high-angle fracture
BH10	53.05	57.87	58.17	0.1	filled fracture
BH10	53.15	282.76	75.21	1.96	high-angle fracture
BH10	53.32	149.58	22.84	12.86	bedding plane fracture
BH10	54.87	149.02	23.57	6.41	bedding plane fracture
BH10	55.77	143.77	23.98	0.1	bedding plane fracture
BH10	56.15	148.13	29.79	0.1	bedding plane fracture
BH10	56.55	138.87	25.31	6.18	bedding plane fracture
BH10	56.69	286.3	73.3	0.1	high-angle fracture
BH10	57.3	237.97	79.08	0.1	high-angle fracture
BH10	57.73	151.69	22.83	0.1	bedding plane fracture
BH10	58.82	144.96	24.69	6.34	bedding plane fracture
BH10	59.29	39.46	84.4	1.55	high-angle fracture
BH10	59.79	143.31	23.85	0.1	bedding plane fracture
BH10	60.6	106.66	85.35	1.11	high-angle fracture
BH10	60.64	39.65	86.74	0.1	filled fracture
BH10	60.78	232.31	79.02	0.1	filled fracture
BH10	61.05	229.64	77.15	0.1	filled fracture
BH10	61.06	114.55	60	0.1	high-angle fracture
BH10	61.4	42.63	85.68	2.84	filled fracture
BH10	61.47	156.21	26.2	0.1	bedding plane fracture
BH10	62.07	222.2	88.19	0.1	filled fracture
BH10	62.13	106.76	79.04	0.1	high-angle fracture
BH10	62.39	224.4	88.17	0.1	high-angle fracture
BH10	64.27	220.32	83.73	0.1	filled fracture
BH10	64.61	225.88	86.18	0.1	filled fracture
BH10	64.95	222.24	75.09	0.1	filled fracture
BH10	65.08	330.79	31.54	0.1	high-angle fracture
BH10	65.08	217.5	81.43	0.1	high-angle fracture
BH10	65.09	100.53	38.06	0.1	filled fracture
BH10	65.21	222.85	71.84	0.1	high-angle fracture
BH10	65.42	222.15	80.23	0.1	filled fracture
BH10	65.71	142.59	24.97	0.1	bedding plane fracture
BH10	66.17	211.67	83.39	0.1	filled fracture
BH10	66.22	214.51	86.34	0.1	filled fracture
BH10	66.32	216.3	79.71	0.1	filled fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH10	66.43	1.18	22.43	19.34	high-angle fracture
BH10	66.68	205.82	72	0.1	filled fracture
BH10	66.78	213	64.64	0.1	filled fracture
BH10	66.85	226.53	80.03	0.1	filled fracture
BH10	67.35	283.85	71.12	0.1	high-angle fracture
BH10	67.41	214.05	78.45	0.1	filled fracture
BH10	67.5	115	71.74	0.1	high-angle fracture
BH10	67.72	201.46	72.58	0.1	filled fracture
BH10	67.92	163.96	35.99	9.24	bedding plane fracture
BH10	67.96	218.92	74.12	0.1	filled fracture
BH10	67.97	224.02	79.54	0.1	filled fracture
BH10	68.25	118.76	59.89	6.42	high-angle fracture
BH10	68.54	220.61	80.41	0.1	filled fracture
BH10	68.77	216.5	85.5	0.1	filled fracture
BH10	69.83	205.44	72.25	0.1	filled fracture
BH10	70.01	212.02	65.11	0.1	filled fracture
BH11	7.86	164.84	22.51	0.1	bedding plane fracture
BH11	9.38	134.13	27.02	5.53	bedding plane fracture
BH11	9.66	155.18	22.56	5.91	bedding plane fracture
BH11	10.02	108.89	76	2.26	high-angle fracture
BH11	10.4	137.26	24.34	19.08	bedding plane fracture
BH11	10.69	157.03	20.97	9.5	bedding plane fracture
BH11	12.72	154.82	24.32	9.06	bedding plane fracture
BH11	13.23	104.88	80.12	1.13	high-angle fracture
BH11	15.06	284.11	76.23	1.75	high-angle fracture
BH11	16.37	106.4	77.08	3.13	high-angle fracture
BH11	16.78	152.1	18.72	9.77	bedding plane fracture
BH11	17.37	79.08	18.32	6.25	high-angle fracture
BH11	19.13	155.63	21.32	6.87	bedding plane fracture
BH11	19.76	155.95	23.13	6	bedding plane fracture
BH11	20.35	156.71	23.47	6.01	bedding plane fracture
BH11	20.64	156.44	22.32	6.1	bedding plane fracture
BH11	22.05	44.93	60.46	0.1	high-angle fracture
BH11	22.21	284.99	36.97	0.1	high-angle fracture
BH11	24.49	106.68	72.1	2.91	high-angle fracture
BH11	24.63	108.51	70.41	4.89	high-angle fracture
BH11	25.08	118.54	53.71	0.1	high-angle fracture
BH11	25.28	110.03	66.16	0.1	high-angle fracture
BH11	25.52	105.48	68.03	0.1	filled fracture
BH11	26.04	8.29	47.72	3.02	high-angle fracture
BH11	26.83	284.69	82.5	0.1	high-angle fracture
BH11	27.08	144.92	26.68	5.78	bedding plane fracture
BH11	27.41	280.91	74.07	0.1	high-angle fracture
BH11	29.23	107.72	49.36	4.94	filled fracture
BH11	30.03	288.94	71.22	0.1	high-angle fracture
BH11	33.4	157.95	86.15	3.84	high-angle fracture
BH11	33.54	3.96	52.01	3.25	high-angle fracture

Borehole	Depth (m)	Azimuth (deg)	Dip (deg)	Aperture (mm)	Fracture type
BH11	34.25	93.6	89.69	0.1	high-angle fracture
BH11	35.06	103.65	58.67	0.1	high-angle fracture
BH11	35.87	107.83	84.61	0.1	high-angle fracture
BH11	36.76	108.64	75.39	3.21	high-angle fracture
BH11	37.93	282.82	71.53	0.1	high-angle fracture
BH11	37.96	283.67	60.82	0.1	high-angle fracture
BH11	38.03	282.56	80.97	0.1	high-angle fracture
BH11	39.34	112.09	72.31	2.57	filled fracture

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Peer-reviewed Articles, Published in Refereed Journals:

DesRoches AJ, and Butler KE (2016) Monitoring and modelling of pumping-induced self-potentials for transmissivity estimation within a heterogeneous confined aquifer, *Geophysical Journal International*. 207(3):1722-1738.

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