Hydrogeophysics of Coastal Aquifers -
geophysical applications to the assessment of
groundwater properties in coastal gravel and karst
aquifers

A thesis submitted to the National University of Ireland, Galway
for the degree of Doctor of Philosophy

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Declaration

This work has not previously been submitted for a degree in any university. To the best of my knowledge and belief, the thesis contains no material previously published or written by another person except where due reference is made in the thesis itself.

Yvonne O'Connell
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Abstract

Airborne, terrestrial and marine geophysics have been employed to examine coastal zone hydrogeology in two very different aquifer types in Ireland. Modelling of regional airborne electromagnetic data informed by ground-based ERT allowed the determination of the aquifer electrical resistivity/conductivity distribution within a locally important coastal sand and gravel aquifer on the east coast. The effective electrical conductivity as a function of porosity for the groundwater body was estimated with salinity and clay mass fraction constraints from supplementary drilling. The hydraulic conductivity depth function was estimated after calibration with a hydraulic conductivity measurement in the field allowing a determination of lateral variations in groundwater hydraulic conductivity across the sand and gravel aquifer, showing potential for upscaling to catchment scale water resource management. On the west coast of Ireland, groundwater movement in two catchments of a regionally important coastal karst aquifer was examined with terrestrial and marine geophysics. This hydrogeophysical examination confirmed the extension of faulting offshore and groundwater egress/saltwater ingress along fault. The examination of subterranean conduit and epikarstic flow paths and intertidal and submarine groundwater discharge provided new evidence for previously predicted terrestrial and submarine pathways confirming the existence of active SGD. Discrete conduit network modelling, informed by the terrestrial geophysics coupled with ancillary hydrogeological observations characterised the hydrogeology of one of the catchments by determining flow pathways and their likely hydraulic mechanisms indicating a complex conduit network of constrictions and bypass channels with multiple inlets and outlets. This work augments our understanding of groundwater movement in the coastal zone for the management of coastal aquifer systems and provides transferrable practices to assist in catchment-scale assessments of coastal aquifers.
Chapter 1: Introduction

1.1 General Introduction & Key Concepts

Groundwater accounts for more than 97% of all fresh water available on Earth (excluding the water locked up in glaciers and ice caps) and is the largest source of drinking water (~75% of the supply) across Europe (European Commission, 2008). This global resource is under threat as aquifers are vulnerable to numerous pressures including extreme flood events that can contaminate groundwater from rapidly percolating surface pollution and anthropogenic exploitation and pollution, predominantly from industrial and agricultural activities (Daly, 2009; Foster et al., 2013; Wada et al., 2010). Coastal aquifers are subject to additional pressures. In the coastal zone, groundwater is exchanged across a transition zone between land and sea, with increased abstraction of groundwater potentially drawing seawater into coastal aquifers (Changming et al., 2001; Ghiglieri et al., 2012; Petelet-Giraud et al., 2016; Trabelsi et al., 2012). Climatic variations, giving rise to changes in the hydrologic cycle such as decreased recharge and increased evapotranspiration (e.g., Meixner et al., 2016; Taylor et al., 2012), coupled with predicted sea-level fluctuations (IPCC, 2014), are potential drivers of saltwater intrusion (Comte et al., 2016; Werner et al., 2013) and this is a critical consideration for the EU given that 200 million citizens live in coastal regions.

Submarine Groundwater Discharge (SGD) forms a major component of the coastal hydrological cycle, consisting of terrestrial groundwater mixed with seawater which infiltrates coastal aquifers (Moore, 2010). SGD has been examined globally by various authors e.g., Breier et al. (2005), Viso et al. (2009) and Beck et al. (2016) in North America, Moore (2006), Roxburgh (1985) and (Zektser and Dzyuba, 2014) in Europe, Hwang et al. (2016), Swarzenski et al. (2006)) and Taniguchi et al. (1998) in Asia and Stewart et al. (2015) in Australia. SGD broadly refers to any and all flows of water upwards across the sea floor and may be made up of fresh water, recirculated seawater or a combination of both (SCOR/LOICZ, 2004). It is becoming increasingly important to consider SGD in light of climate-change impacts which can stress coastal aquifers from two opposing directions. Sea level rise (Michael et al., 2013; Moore, 2010) has the potential to
increase the volume of saltwater entering the aquifer and altering flows leaving the system e.g. there can be a shift inland of SGD locations or changes in discharge such as more of less diffuse SGD. A greater frequency and magnitude of storm events, which can lead to increased sea surge events resulting in salinisation in low lying areas, can increase contamination of the groundwater from rapidly percolating surface pollution from anthropogenic activities, for which SGD can potentially act as a pathway to the sea (Taniguchi et al., 2002). Characterising groundwater movement and determining the hydrogeological parameters across the coastal zone is therefore critical for management strategies and monitoring the impacts of climactic events on aquifer resources.

In coastal aquifers, SGD investigation traditionally includes monitoring of freshwater outputs using seepage meters (Lee, 1977; Shinn et al., 2002) monitoring wells (Perriquet et al., 2014) and natural tracers (colour, temperature, salinity e.g., Taniguchi et al. (1998), Martin et al. (2006) and Wilson and Rocha (2012)) and geochemical tracers such as radium isotopes, radon, methane, hydrogen sulphide or carbon dioxide e.g., Porcelli and Swarzenski (2003), Moore (2006) and Schubert et al. (2015). SGD can also be determined through indirect methods such as water balance approaches e.g., Sekulic and Vertacnik (1996) in the Adriatic, and Sellinger (1995) in the USA, hydrograph separation techniques e.g., Zektser and Dzhamalov (1981) for the Pacific Ocean Rim and Boldovski (1996) in Russia, and through theoretical analysis and numerical simulations e.g., Bokuniewicz (1992) and Fukuo and Kaihotsu (1988). Concentrated SGD monitoring in Ireland has only recently begun with work by Cave and Henry (2011), Einsiedl (2012) and Wilson and Rocha (2012).

In Ireland, groundwater provides not only ~25% of the drinking water supply (Daly, 2009) but it is also an integral resource for industry and agriculture. EU policies for the protection, improvement and sustainable use of waters, specifically the Water Framework Directive 2000/60/EC (WFD) and the associated Groundwater Directive 2006/118/EC, have been implemented in Ireland to establish a framework for the protection of all waters including groundwater, transitional and coastal waters (RIA, 2009), driving research e.g., Flynn (2010) and Gill et al. (2013). With more than 50% of the Irish
population living in coastal areas (Devoy, 2008), coastal aquifers play an important role in providing fresh water supplies to coastal communities and as sources of nutrients to estuaries and bays, contributing to aquaculture e.g. Smith and Cave (2012). The Marine Strategy Framework Directive, MSFD 2008/56/EC (EU, 2008) was established by the EU in 2008 to protect marine waters and its implementation in Ireland is ongoing (DECLG, 2016). Many coastal zone management strategies have been initiated globally e.g. New Zealand (New Zealand Government, 2010) and Australia (EA, 1998), however as yet, no coastal zone management strategy has been implemented in Ireland. In 2013, the EU established a joint initiative to establish a framework on integrated coastal management and maritime spatial planning (EU, 2013) to assist with the implementation of other EU directives (including MSFD and WFD) in the coastal area. This initiative recognises that coastal zones are among the most vulnerable areas to climate change and natural hazards, at risk from sea level rise, flooding and extreme weather events.

In Ireland, all rocks, whether consolidated or unconsolidated, are classed as aquifers. In this work, two coastal aquifers comprising of a consolidated karst limestone aquifer and an unconsolidated sand and gravel aquifer are the focus of attention (Figure 1.1). While the majority of consolidated rock aquifers are classified as poor to moderately productive, locally important aquifers, karst limestone aquifers, which make up the majority of the regionally important aquifers, are very productive. Karst limestone aquifers occur intermittently along the west coast from Glenbeigh in Co. Kerry to Killybegs in Co. Donegal, in the southeast on the Co. Waterford and Co. Wexford coastlines and locally at Bettystown in Co. Louth. Sand and gravel (a catch-all description for unconsolidated materials) aquifers are generally classified as locally important with some deposits of regional importance in the southeast of Ireland. Locally important sand and gravel aquifers are mapped along much of the coastline of County Louth (GSI, 2007b) with only sporadic sand and gravel aquifers dotted along the rest of the Irish coastline in Counties Wexford, Mayo, Sligo and Donegal.
Figure 1.1: Bedrock aquifer types (colour blocks), sand and gravel aquifer locations (red polygons), coastal locations examined within this thesis and the Tellus Border airborne survey coverage (GSI, 2006, 2007b).

Under the terms of the WFD, all aquifers are categorised in whole or part as groundwater bodies (GWB) defined as distinct volumes of groundwater in an aquifer for the purposes of groundwater management (WFD Working Group on Groundwater, 2005). Characterisation of water bodies is a key element of the WFD in which the physical characteristics including water body boundaries, their geology and hydrogeology, linked groundwater and surface water systems are determined. Other EU policies such as the Floods Directive 2007/60/EC, Drinking Water Directive 98/83/EC and the Bathing Water Directive 2007/7/EC also necessitate a comprehensive understanding of aquifer resources and vulnerabilities.
1.2 Hydrological properties

Groundwater movement within an aquifer is governed by the hydraulic properties of effective porosity, permeability, hydraulic conductivity and transmissivity as detailed in the standard texts e.g., Domenico and Schwartz (1998), Fetter (2001), Hiscock (2005). Porosity is the volume of pore space as a percentage of the bulk volume, with effective porosity being the porosity available for fluid flow. Carbonate rocks e.g. limestone, can be subject to dissolution by chemically aggressive water, increasing secondary porosity along fractures, joints and fissures which can lead to the development of conduits that act as the dominant pathway for groundwater flow (Ford and Williams, 2007). The resultant karst limestones have a combination of primary, secondary and tertiary porosities, the percentages for which can vary greatly (Worthington, 1999). In unconsolidated sediments the porosity is typically primary (or intergranular) and depends predominantly on grain size distribution, grain shape, packing and sorting. Unconsolidated sediment porosities can range from 5% in fluvial deposits to as high as 70% in silts and clays (Hiscock and Bense, 2014). Additionally, the matrix is not rigid and is subject to consolidation and compaction, resulting in a changeable porosity.

Intrinsic permeability, $k$, is a measure of the ability of a medium (rock or sediment) to allow the flow of any fluid and is a function of the size of the openings through which the fluid can move (Fetter, 2001) having dimensions of area. Permeability in sediments is greatest in well sorted sands and gravels, diminishing with increased silt and clay content while consolidated rock permeability is a function of primary and secondary openings formed during and post rock formation respectively. The hydraulic conductivity, $K$, is synonymous with permeability, measuring the ease with which a fluid (usually water) passes through a porous medium and has units of length/time. It is a property of both the medium and the fluid:

$$ K = \frac{kp g}{\mu} $$

where $\rho$ is the density, $g$ is the acceleration due to gravity and $\mu$ is the dynamic viscosity of the pore fluid e.g. Hiscock and Bense, 2014. $K$ is typically determined either experimentally through field and laboratory testing (e.g. pump tests or falling head tests) or empirically using estimates
of intrinsic properties. Typically, hydraulic conductivity values in well sorted sands and gravels are high, diminishing with increased silt and clay content.

1.3 Hydrogeophysical Properties

The minerals that comprise soils and rocks are predominantly electrical insulators such that electrical conduction of current is mainly electrolytic (i.e. via dissolved ions in the pore fluid within the soil and rock matrix) e.g., Archie (1942), Parkhomenko (1967), Jackson et al. (1978). Electrical resistivity is a measure of the resistance of a medium to the flow of current. It is governed by Ohm’s Law which states that the resistance of the medium is proportional to the applied voltage and inversely proportional to the electrical current flow. Electrical resistivity has units of measurement of Ohm metres (Ωm) and is the inverse of electrical conductivity, measured in Siemens per metre (Sm⁻¹). Electrical conductivity is therefore a measure of the ability of a material to permit current flow.

Archie (1942) determined that the bulk resistivity (i.e. the resistivity for the volume of the medium) of a granular medium is a function of the porosity, the pore fluid resistivity and the degree of saturation. Some typical bulk values for common materials are contained in Table 1.

Table 1: Typical resistivity and conductivity for common geologic materials and water (Reynolds, 1997; Freeze and Cherry, 1979).

<table>
<thead>
<tr>
<th>Material</th>
<th>Nominal Resistivity (Ωm)</th>
<th>Nominal Conductivity (mS/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay</td>
<td>1-150</td>
<td>7 - 10³</td>
</tr>
<tr>
<td>Boulder Clay</td>
<td>15-35</td>
<td>30 - 70</td>
</tr>
<tr>
<td>Sand &amp; Gravel</td>
<td>30-225</td>
<td>4 - 33</td>
</tr>
<tr>
<td>Limestone</td>
<td>5 X 10⁻¹⁰⁻⁶</td>
<td>6x10⁻⁷⁻ 20</td>
</tr>
<tr>
<td>Sandstone</td>
<td>1.7 X 10⁸⁻ 1 X 10⁻¹¹</td>
<td>1 x 10⁻¹⁵⁻ 10⁻¹⁰</td>
</tr>
<tr>
<td>Slate</td>
<td>6 X 10⁻⁷⁻ 4 X 10⁻⁶</td>
<td>2.5 x 10⁻⁷⁻ 2</td>
</tr>
<tr>
<td>Marble</td>
<td>10⁻²⁻ 2.5 X 10⁻⁷</td>
<td>4x10⁻⁶⁻ 10</td>
</tr>
<tr>
<td>Granite</td>
<td>3 X 10⁻⁷⁻ X 10⁻⁸</td>
<td>3x10⁻⁴⁻ 3</td>
</tr>
<tr>
<td>Basalt</td>
<td>10⁻¹⁻ 1.3 X 10⁻⁷</td>
<td>8x10⁻³⁻ 100</td>
</tr>
<tr>
<td>Fresh water</td>
<td>&gt; 7</td>
<td>&lt;140</td>
</tr>
<tr>
<td>Brackish water</td>
<td>0.7 – 7</td>
<td>140 – 1.4x10⁻⁷</td>
</tr>
<tr>
<td>Seawater</td>
<td>0.07 - 0.7</td>
<td>1.4x10⁻⁷⁻ 10⁻⁰</td>
</tr>
</tbody>
</table>
In a saturated system, where porosity is relatively constant and not subject to significant change over short periods of time, a basic assumption can be made that changes in bulk resistivity can be attributed to changes in pore water salinity i.e. an increase in the total dissolved solids (Falgàs et al., 2011; Henderson et al., 2009), an assumption that can be exploited to examine seawater influence in to an aquifer. Archie’s Law assumes little or no clay content such that electrical conduction occurs by the migration of ions through the bulk pore water only and has proven generally valid only in resistive granular material (Jackson et al., 1978; Niwas et al., 2011; Wyllie and Gregory, 1953). However the determination of the porosity and related hydraulic properties is complicated by salinity and/or clay content. Increased pore water salinity reduces pore water resistivity e.g. Freeze and Cherry (1979), while the contribution of clay particles, which are capable of ion exchange through complex conduction mechanisms (Sen et al., 1981), must also be considered though is regularly ignored as the clay particulate is often not substantiated. The bulk resistivity becomes a combination of the pore fluid resistivity and the surface conduction at the grain/pore fluid interface, with surface conduction being a function of the surface area (Schwarz, 1962) and the surface chemical properties at the grain-fluid interface (Lesmes and Friedman, 2005). This has been addressed through various models (Bussian, 1983; Clavier et al., 1977; Gelius and Wang, 2008; Revil et al., 1998; Sen et al., 1988; Waxman and Smits, 1968).

By determining bulk electrical resistivities or conductivities, the hydraulic properties of porosity, permeability and hydraulic conductivity can be estimated, providing constraints for hydrogeological modelling, a key objective for water saturated environments given the difficulty of characterising the hydraulic properties of aquifers from limited numbers of boreholes e.g. Lesmes and Friedman (2005). Both electrical and hydraulic parameters are dependent on each other, being generally governed by the same physical parameters and lithological attributes e.g. lithology, porosity, pore size and interconnectivity, grain size, shape and mineralogy, compaction, consolidation and cementation. The interconnection between both sets of parameters has been examined by many authors (Abu-Hassanein et al., 1996; 1984; Jackson et al., 1978; Niwas et al., 2011; Revil, 2013; Rhoades et al., 1976; Salem, 1999; Shah and Singh, 2005; W;
Zhu et al., 2016), largely focused on the examination of reservoirs for oil exploration (Archie, 1942; Bussian, 1983; Clavier et al., 1984; Revil et al., 1998; Revil and Florsch, 2010; Sen et al., 1988; Waxman and Smits, 1968). The electrical and hydraulic conductivity relationship within groundwater aquifers specifically has also been examined (Domenico and Schwarz, 1990; Niwas and Celik, 2012; Niwas et al., 2011; Purvance and Andricevic, 2000).

1.4 Geophysical techniques
A variety of geophysical techniques can be employed to assist with the understanding of subsurface geology, with electrical and electromagnetic techniques generally proving most applicable in hydrogeological investigations e.g., Comte et al. (2012), Gondwe et al., (2012) and Martorana et al. (2014), in addition to seismic and potential field methods (see Table 2). The work presented in this thesis employs DC resistivity and electromagnetic induction techniques.

DC resistivity electrical imaging surveys are routinely used for hydrogeological applications including groundwater/surface water interactions e.g., Nyquist et al. (2008), seawater intrusion e.g., Comte & Banton (2007) and Nguyen et al. (2009), salinity studies of lakes and water reservoirs e.g., Amidu and Dunbar (2008) and in the detection of zones of karstification and conduits in the examination of groundwater movement through highly heterogeneous karst regions e.g., Zhu et al. (2011). The advent of electrical imaging systems capable of operating in a marine environment, previously prohibitive due to the conductive influence of the seawater, have allowed qualitative investigation of diffuse SGD through sediments (Belaval, 2003; Breier et al., 2005; Day-Lewis et al., 2006; Manheim et al., 2004).

Electromagnetic induction techniques allow the determination of subsurface conductivities and are classified as either frequency-domain (FDEM or FEM) or time-domain (TDEM or TEM) (see Appendix 1 for more detailed discussion). FEM is typically used in shallow applications e.g., engineering, archaeological, and groundwater applications while TEM, with a greater depth of investigation, has typically been used for deeper exploration e.g.,
groundwater or mineral exploration. A comparison of vertical ranges achievable from FEM, TEM and ERT (Figure 1.2, from Binley (2015)) indicates TEM generally providing greater depths than FEM and ERT. See Appendix 1 for a more detailed discussion on depths of investigation.

Table 2: List of commonly used geophysical methods in hydrology and the geophysical properties they sense (adapted from Binley (2015)).

<table>
<thead>
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<th>Geophysical Method</th>
<th>Geophysical Properties</th>
<th>Examples of derived properties and states</th>
<th>Key references</th>
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<tr>
<td>Electrical</td>
<td>DC resistivity e.g. VES, ERT</td>
<td>Electrical conductivity</td>
<td>Water content, clay content, pore water conductivity</td>
</tr>
<tr>
<td>Induced polarization</td>
<td>Electrical conductivity, chargeability</td>
<td>Water content, clay content, pore water conductivity, surface area, permeability</td>
<td>e.g., Börner et al., 1996; Slater and Lesmes, 2002</td>
</tr>
<tr>
<td>Spectral induced polarization</td>
<td>As above but with frequency dependence</td>
<td>Water content, clay content, pore water conductivity, surface area, permeability, geochemical transformations</td>
<td>e.g., Revil, 2010</td>
</tr>
<tr>
<td>Self-potential</td>
<td>Electrical sources, electrical conductivity</td>
<td>Water flux, permeability</td>
<td>e.g., Abaza and Clyde, 1969; Revil et al., 2003</td>
</tr>
<tr>
<td>Electromagnetic induction</td>
<td>Electromagnetic induction e.g. FDEM, TDEM</td>
<td>Electrical conductivity</td>
<td>Water content, clay content, salinity</td>
</tr>
<tr>
<td>Ground penetrating radar</td>
<td>Permittivity, electrical conductivity</td>
<td>Water content, porosity, stratigraphy</td>
<td>e.g., Huisman et al., 2003; Topp et al., 1980</td>
</tr>
<tr>
<td>Seismic</td>
<td>Seismic e.g. Refraction, Reflection</td>
<td>Elastic moduli and bulk density</td>
<td>Lithology, ice content, cementation state, pore fluid substitution</td>
</tr>
<tr>
<td>Seismoelectrics</td>
<td>Electrical current density</td>
<td>Water content, permeability</td>
<td>e.g., Revil et al., 2014; Sava et al., 2014</td>
</tr>
<tr>
<td>Nuclear magnetic resonance</td>
<td>Proton density</td>
<td>Water content, permeability</td>
<td>e.g., Herckenrath et al., 2012; Vilhelmsen et al., 2014</td>
</tr>
<tr>
<td>Gravity</td>
<td>Bulk density</td>
<td>Water content, porosity</td>
<td>e.g., Gehman et al., 2009; Blainey et al., 2007</td>
</tr>
</tbody>
</table>
Chapter 1

Figure 1.2. Comparison of vertical survey scales for ERT, FEM, and TEM assuming 1 and 5 m spaced electrode arrays for ERT(1) and ERT(5) respectively. 1 m coil and 3.5m coil instruments for FEM(1) and FEM(3.5) respectively and a 50 m loop sounding at multiple stations along a transect TEM (Binley, 2015).

The Tellus and Tellus Border airborne surveys acquired across Northern Ireland and the border counties of the Republic of Ireland (Beamish et al., 2006; Hodgson and Ture, 2013) (Figure 1.1), while focused on magnetic and radiometric acquisition for natural resources exploration, acquired high-resolution airborne FEM data. Airborne electromagnetic surveying (either FEM or TEM), traditionally used for natural resource and environmental applications, is increasingly being used for hydrogeological applications e.g. the examination of groundwater quality in a crystalline aquifer (de Souza Filho et al., 2010) and structural mapping in karst aquifers to determine groundwater flow paths (Gondwe et al., 2012). Other applications have been concerned with anthropogenic groundwater contamination (Wilson et al., 2013) in addition to groundwater contamination through coastal saltwater intrusion (Beamish, 2012; Fitterman and Deszcz-Pan, 1998; Kirkegaard et al., 2011). Its application
to hydrogeophysics continues to advance through research groups such as the Hydrogeophysics Group in Denmark (http://hgg.au.dk/).

The Tellus airborne survey, through its acquisition of FEM data, allows the development of new hydrogeophysical applications. This work aims to expand these applications to maximise the benefits of the Tellus FEM dataset by applying petrophysical modelling more commonly used in the lab or oil industry. To date, the Tellus airborne survey has expanded to cover the northern half of the island of Ireland with future plans for all Ireland coverage. This offers an unrivalled opportunity to develop hydrogeophysical applications of the FEM dataset at catchment and regional scales.

1.5 Research Challenge

With limited information or evidence for impacts of climate change on coastal aquifers (Comte et al., 2016) the determination of hydrogeological parameters and characterisation of groundwater movement across the coastal zone is critical for management strategies and monitoring the impacts of climactic events on aquifer resources. The pre-requisite for the modelling strategy required to forecast impacts on groundwater is an assessment of the contribution to groundwater fluxes from, *inter alia*, diffuse sources and discrete pathways. As such the characterisation of SGD and groundwater pathways in coastal aquifers is critical for understanding climate-change and anthropogenic impacts e.g., Kløve et al. (2014) and Sweeney et al. (2008). Determination of the hydrogeology and hydrological properties of aquifers generally suffers from a paucity of data. *In-situ* sensors are commonly absent or sparse and often prohibitively expensive for wide-scale application, so complementary approaches to assess groundwater potential at catchment scales are required.

Geophysics is increasingly being used to image the subsurface, laterally and as a function of depth. Ground based electrical and electromagnetic techniques and remotely sensed airborne techniques offer an opportunity to examine microscale electrical properties of sediments and rocks and potentially upscale these observations to determine aquifer hydrogeology.
and hydraulic properties on much broader vertical and lateral scales than the traditional point based in situ data e.g. (Binley et al., 2015; Sophocleous, 2002; von Hebel et al., 2014). The drawbacks of traditional in situ measurements of system properties (e.g. permeability) are the reasons why geophysical techniques have emerged as valuable tools for upscaling to broader spatial scales. There are, however, still considerable challenges in how to link hydrogeologically-relevant subsurface properties to measurable geophysical parameters, as outlined in the review by Binley et al. (2015).

As part of the Griffith Geoscience Research Awards Programme ¹, the work presented here aims to determine processes that contribute to water and land management at catchment scale by providing low-cost, low-disruption strategies that can upscale limited measurements to quantify catchment-scale groundwater resources for sustainable abstraction rates and the mitigation of flood or drought risk.

### 1.6 Scope of Works

Limited coastal hydrogeophysics have been carried out on a non-commercial level in Ireland to date (Gibson et al., 2012) and tend to be spearheaded by the Integrated Mapping for the Sustainable Development of Ireland’s Marine Resource (INFOMAR) under the direction of the Marine Institute of Ireland and the Geological Survey of Ireland. Marine geophysical research in Ireland is typically focused on the offshore environment e.g., Garcia et al. (2014) and Sacchetti et al. (2012). This is the broad focus of research groups such as the In situ Marine Laboratory for Geosystems Research (iMARL) and INFOMAR.

This thesis seeks to show how geophysical techniques including terrestrial and marine electrical methods and remotely sensed airborne techniques can be employed to examine the hydrogeological properties of aquifers. The determination of the microscale electrical properties of the sediments and rocks, to the macroscale examination of groundwater flowpaths is

¹ The Griffith Geoscience Research Awards Programme was established in 2007 to drive the implementation of the National Geoscience Programme and to stimulate research capacity.
explored to increase our understanding of groundwater movement in two very different aquifers along the Irish coastal zone. A sand and gravel aquifer found along the County Louth coast in the east of Ireland (Location 1, Figure 1.1) and a karst aquifer on the south coast of Galway Bay in the west of Ireland (Location 2, Figure 1.1).

**Co. Louth Sand & Gravel Aquifer**

The Dromiskin Gravel GWB sand & gravel aquifer is one of the largest aquifers of its type in Ireland. The sand and gravel aquifer, defined under the WFD as the Dromiskin Gravels GWB is categorised as a locally important aquifer based on having an areal extent <10 km² and saturated thickness of superficial deposits >5 m (DELG/EPA/GSI, 1999). The broader region is dominated by poorly productive regions. The GWB covers a surface area of 8.3 km² but its thickness and saturation are unknown and only estimates have been made of its hydraulic properties (GSI and RBD Consultants, 2005). In Ireland, sand and gravel aquifers generally consist of unconsolidated coarse-grained material usually with <8% fines (O’Suilleabháin, 2000) and are generally described as having primary porosity with inter-granular flow, high permeability and high effective porosity (River Basin Districts, 2005). Groundwater discharges to rivers and streams in the north of the Dromiskin Gravels GWB and to the sea in the east. Given the low-lying topography of the area and proximity to the coast there is a high potential for saline intrusion into the aquifer.

The Tellus regional airborne survey, spearheaded by the Geological Survey of Northern Ireland (GSNI) was completed in 2008. The Tellus Border survey was acquired across six border counties, by the Geological Surveys of Ireland (GSI) and GSNI from 2011-2013. The data was examined in a report by O'Connell and Daly (2013) to identify groundwater pathways in the coastal zone (Fig 1.3). The Dromiskin Gravel GWB aquifer was highlighted as a region of interest due to its coastal position, potential saline influences and the heterogeneity of its deposits as suggested by the airborne electromagnetic and radiometric data.
The hydraulic properties of effective porosity and hydraulic conductivity within this coastal aquifer deposit are examined in this thesis through the application of predictive petrophysical models (Bussian, 1983; Revil et al., 1998) informed by modelling of the remotely sensed airborne geophysical data to determine the electrical conductivity distribution, and substantiated by a ground-truthing ERT campaign, coupled with auxiliary drilling and lab testing of the clay cation exchange capacity (CEC) and grain size distribution. This integrated approach aims to demonstrate how remote sensing airborne geophysical data can contribute to water and land management at catchment scale and larger and explores the limits of quantifying the properties of near-surface clay-sand-gravel aquifers with sparse ground-based hydrogeological information.

**South Galway Bay coastal Karst Aquifer**

The karst aquifer system underlies the Burren and Gort-Kinvarra Lowlands in South Co. Galway and Co. Clare (Fig 1.4) and is characterised as a
regionally important aquifer (DELG/EPA/GSI, 1999). This coastal karst aquifer is the most extensive coastal karst system in Ireland. The dominant groundwater flow is through karstified conduits discharging via both the intertidal zone and submarine springs (McCormack et al., 2014; Perriquet et al., 2014). Groundwater velocities suggest flow rates to intertidal springs at Kinvarra >0.4 ms\(^{-1}\) (GSI, 2004). A report examining flooding problems in the region (OPW, 1998) concluded that the karst system has a specific capacity that when surpassed can result in extensive flooding. Modelling of the groundwater conduit system carried out for the 1998 OPW report, based on extensive dye tracing, water level monitoring, hydrochemical sampling and geological mapping and drilling, proposed a conceptual conduit flow model for the region which has been further examined by Gill et al. (2013) and McCormack et al. (2014). Geochemical and isotopic analysis of groundwater/seawater interactions have highlighted the nature of the seawater influence into this karst aquifer (Petrunic et al., 2012) and have been used to assess fresh water contributions (Einsiedl, 2012; Schubert et al., 2015).

To date, research has focused on groundwater chemistry, biogeochemical reactions, nutrient and contaminant loading, thermal imaging, saltwater intrusion, groundwater flow paths, groundwater discharge and residence times in a coastal karst aquifer (Cave and Henry, 2011; McCormack et al., 2014; Perriquet et al., 2014; Smith and Cave, 2012; Wilson and Rocha, 2012). The hydrodynamic properties of a coastal catchment within the karst aquifer have been determined by Perriquet et al. (2014) providing a baseline for future modelling in Bell Harbour. Cave and Henry (2011) and (McCormack et al., 2014) determined SGD rates by examining rainfall rates, temperature and salinity monitoring and nutrient sampling in Kinvarra Bay.
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Figure 1.3: Regional setting of southern Galway Bay showing the Bell Harbour and Gort-Kinvarra catchments and associated topography.

The work presented in this thesis uses electrical resistivity tomography and discrete conduit network modelling to characterise catchment hydrogeology by determining flow pathways and their likely hydraulic mechanisms. This work aims to substantiate existing hypotheses on groundwater movement and discharge in two catchments of the coastal karst aquifer (McCormack et al., 2014; Perriquet et al., 2014; Schubert et al., 2015) while examining the links between the electrical resistivities and hydraulic properties of pore water salinity and porosity. Towed marine, static terrestrial and time-lapse ERT techniques to examine groundwater flow paths and SGD are integrated with existing hydrogeological information to provide new insights in to how the structural geology and conduit networks influence SGD and saltwater ingress across the coastal zone. To date no authors have used the towed ERT technique in a karst aquifer setting. Informed by the ERT, conduit network modelling of the Bell Harbour catchment, aims to characterise flow pathways and their likely hydraulic mechanisms, with particular emphasis on conduit linkages between two ephemeral lakes and the sea.
1.7 Summary of Chapters

Chapter 2 details the assessment of hydraulic properties within the Dromiskin Gravels GWB. Chapter 3 details the hydrogeophysical examination of the karst aquifer on the southern coast of Galway Bay. Chapter 4 presents the examination of flow pathways in the Bell Harbour catchment. Chapter 5 includes a discussion of the work, conclusions made and recommendations for future work.
Chapter 2: Assessment of groundwater resources using airborne electromagnetic remote sensing

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2.1 Abstract
High-resolution airborne electromagnetic (EM) remote sensing data have been used to aid the assessment of groundwater resources and the quantification of hydrological properties of a clay-sand-gravel aquifer. The airborne EM data have been integrated with ground-based electrical resistivity data, and both data sets have been inverted to infer sub-surface electrical conductivity associated with variable pore fluid salinity and clay content. By using theoretical models to connect these microscale electrical properties with the macroscale conductivity inferred from the data, it is possible to constrain the porosity distribution in clay-sand-gravel mixtures after calibration with field measurements of cation exchange capacity, grain size distribution and pore fluid salinity. The hydraulic conductivity distribution with depth can be estimated using a variety of theoretical models calibrated with field hydraulic conductivity measurements. Airborne EM data can assist with the identification of groundwater resources and their quantification in clay-sand-gravel aquifers only if they are calibrated with surface electrical resistivity or EM data linked with hydrogeological data acquired from in-situ catchment-scale observation networks.

Keywords: hydrogeophysics, aquifer porosity, electrical and hydraulic conductivity
2.2 Introduction

As groundwater constitutes ~97% of all freshwater available on Earth, the characterisation of groundwater resources is essential for water management and fundamental to compliance with international directives. In-situ environmental sensors are commonly absent or sparse and often prohibitively expensive for wide scale application, so complementary approaches to assess groundwater potential at catchment scales are required. This paper demonstrates how remote sensing airborne geophysical data, especially electromagnetic data, can contribute to water and land management at catchment scale. In so doing, it explores the limits of quantifying the properties of near-surface clay-sand-gravel aquifers with sparse ground-based hydrogeological information.

Regional airborne geophysical surveys have been used for over 50 years for a variety of activities (Christensen, 2009) including basin analysis (Boyce and Morris, 2002; Gernigon et al., 2012), structural geology for mineral exploration (Airo and Mertanen, 2008; Holden et al., 2012), environmental assessments (Dent, 2007; Metternicht and Zinck, 2003) and various hydrogeological applications (de Souza Filho et al., 2010; Gondwe et al., 2012). Dickson et al. (2014) used airborne magnetic data with numerical upscaling techniques for regional groundwater modelling of heterogeneous aquifers in Northern Ireland. Airborne electromagnetic (AEM) data have been used and assessed for quantifying subsurface electrical conductivity on catchment scales with an opportunity for applications in groundwater resources and quality (e.g., Bedrosian et al. 2015; Gondwe et al., 2012; He et al., 2014; Siemon et al., 2011). AEM has also been extensively used for saltwater intrusion mapping e.g. Beamish, 2012; Fitterman and Deszcz-Pan, 1998; Rasmussen et al., 2013. Airborne radiometric data are sensitive to natural radioactive elements in the top metre of Earth’s surface, and so may be used to provide information on soil (Beamish and Farr, 2013b; Cook et al., 1996; Rawlins et al., 2009).

Airborne geophysical data can provide estimates of physical properties of the sub-surface and, while this is undoubtedly enhanced in conjunction with ground-based geophysics (Paine and Collins (2010); Oldenborger et al.
(2013) and geochemistry (Ranjbar et al. (2001), their spatial resolution is often sufficient to guide limited ground-based surveys in the reconnaissance of remote environments e.g. Studinger et al. (2003), in semi-urban areas with well-developed infrastructure e.g. Siemon et al. (2011), and in coastal zones where the environment varies sufficiently to preclude systematic ground mapping e.g. Teatini et al. (2011).

This paper utilises high-resolution airborne magnetic (AM), airborne electromagnetic (AEM) and airborne radiometric (AR) data that have been acquired over intensively farmed land in the coastal zone straddling the border between the Republic of Ireland and Northern Ireland. The data were collected as part of two mapping projects, Tellus (TEL: Beamish et al. (2006)) and Tellus Border (TELB: Hodgson and Ture (2013)), linking airborne geophysical and ground-based geochemical data to assess the potential for minerals, hydrocarbons, salt, construction materials, geothermal energy and groundwater resources. The frequency domain AEM data from the TEL survey have been used to map coastal saltwater intrusion (Beamish, 2012) and wetland areas (Beamish and Farr, 2013), while the AR data have assisted hydrogeological investigations associated with water saturation of near-surface soils (Beamish and Farr, 2013) and peat deposits (Beamish, 2013b; Keaney et al., 2013).

This paper presents results from the TELB survey. It focuses on the interpretation of AEM data, calibrated with ground-based electrical resistivity tomography (ERT), and uses the ERT data along with borehole measurements of clay cation exchange capacity (CEC) and grain size distribution to infer the electrical conductivity distribution within a clay-sand-gravel aquifer. These parameters provide input into theoretical models which can predict the hydraulic conductivity distribution within the aquifer after calibration with field-based hydraulic conductivity measurements. The aim is to provide a low-cost strategy that can upscale from single point in situ measurements to catchment scale, to quantify groundwater resources for sustainable abstraction rates in the face of increasing demand, and to help in the mitigation of flood or drought risk.
2.3 Data

2.3.1 Study Area

The region of interest (Figure 2.1a) lies within the Longford-Down terrane, dominated by Silurian greywacke, mudstone, sandstone and shale (Figure 2.1b) with an ENE-WSW Caledonian trend (Anderson, 2004; Chew and Stillman, 2001; Holland, 2001). Paleogene magmatism produced the Mourne, Carlingford and Slieve Gullion granitic intrusions and extensive WNW to NNW trending dyke swarms (Cooper and Johnston, 2004; Preston, 2001). Superficial deposits (Figure 2.1c) are dominated by tills (Fealy et al., 2009; GSNI, 2009) with extensive areas of sub-cropping/outcropping rock and the tills intermingle with deposits of marine sands and gravels, marine/estuarine silts/clays along the coastline. The area lies within the Neagh-Bann River Basin District (RBD), a water management region defined under the Water Framework Directive (EU, 2008; River Basin Districts, 2005), sub-divided into a number of ‘Bedrock Groundwater Bodies’ and ‘Gravel Groundwater Bodies’ (Figure 2.1d), i.e. distinct volumes of groundwater within aquifers (WFD Working Group on Groundwater, 2005). The bedrock groundwater bodies are generally poorly productive aquifers (GSI and RBD Consultants, 2004b; NIEA, 2012) while the gravel groundwater bodies, defined as having inter-granular flow, high permeability, high effective porosity (River Basin Districts, 2005), are locally important aquifers. The gravel groundwater bodies typically cover 1-10 km², their saturated thickness are usually >5 m and, where they consist of superficial deposits, they are usually referred to as ‘Sand and Gravel Aquifers’, regardless of their composition (DELG/EPA/GSI, 1999). The Dromiskin Gravels (DG) is one such groundwater body covering an area of 8.3 km² and unknown thickness and saturation (GSI and RBD Consultants, 2005), though anecdotal information from landowners and from a nearby borehole indicate depth to bedrock of approximately 20 m. The deposit consists of topographically low-lying (c. 5m OD), predominantly marine sands and gravels with a coastal strip of fine-grained, lagoonal and offshore silts and beach deposits, inter-bedded with sands and gravels. Groundwater discharges to rivers and streams to the north and to the sea in the east, the latter implying a high potential for saline intrusion given the nature and depth of the sediments and their topographic elevation.
Figure 2.1: Study area with (a) topography in metres above Ordnance Datum (m OD); CP – Cooley Peninsula, D – Dundalk, DB – Dundalk Bay. (b) bedrock geology; MM - Mourne Mountains, SG - Slieve Gullion, C – Carlingford; igneous complexes rise up to 850m while Ordovician and Silurian lithologies are generally <50 m. (c) soil types (Teagasc et al., 2006) and (d) superficial gravel and bedrock groundwater bodies (GSI, 2007b). The 5km by 5km area centred on the Dromiskin Gravels referred to in the text and subsequent diagrams is shown by a dashed line.
2.3.2 Acquisition and Processing

The TEL campaign was flown in 2006 using a JAC AEM05 wingtip-mounted, four-frequency AEM System (Leväniemi et al., 2009), a 256 Channel Exploranium GR820 spectrometer for the radiometric survey and caesium magnetometers for airborne magnetic (AM) acquisition (Beamish et al., 2006; Hautaniemi et al., 2005). The TELB survey was flown between 2011 and 2012 using the same instrumentation and similar acquisition specifications (SGL, 2012). The flight line angle was 345°, orthogonal to the main geological strike, the line spacing was 200m and survey speeds were from 60-70 ms⁻¹. Flight elevations were 56 m or 59 m for TEL and TELB respectively but with increased terrain clearance in urban areas and areas of steep topography.

The AEM system acquired data at four frequencies (0.912, 3.005, 11.962 and 24.510 kHz), at spatial intervals of 17.5 m (TEL) and 6 m (TELB). The recorded signal, M, for each frequency is a ratio between the induced secondary (Hₘ) and transmitted primary (Hₚ) magnetic fields, (Suppala et al., 2005); it is output in millivolts (mV) as components in-phase (real) and 90° out of phase (quadrature or imaginary) with the primary field (Hautaniemi et al., 2005). Both components were calibrated to parts per million (ppm) of the transmitted signal using a homogenous over-water half-space electrical conductivity model (Hautaniemi et al., 2005; SGL, 2012). These were transformed to apparent conductivities using look-up tables and a pseudo-layer half-space model (Beamish et al., 2006; Fraser, 1978; Hodgson and Ture, 2013). The AEM dataset in this paper consisted of in-phase and quadrature components for all four frequencies of the TELB survey, merged (TEL and TELB) apparent conductivities for the 3.005 and 11.962 kHz frequencies (Beamish, 2013a) and unmerged apparent conductivities for the 0.912 and 24.510 kHz frequencies. Following Beamish and White (2012), a maximum flight altitude of 100m above the terrain was chosen to exclude data with a low signal to noise ratio resulting from increased flight altitudes and to remove cultural noise identified from correlations between potential sources (road and rail networks, power-lines) and the data. Depths of investigation for AEM surveys are dependent on transmitter frequency, flight altitude and ground electrical conductivity.
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(Beamish, 2004). Hodgson and Ture (2014) indicate maximum apparent depths of investigation of > 50 m for the 3.005 kHz low-frequency data; the high frequencies are sensitive to conductivity in the upper 20 m.

The gamma ray data were collected at intervals of 70 m (TEL) and 60 m (TELB), and consisted of potassium, K (%), thorium, Th (ppm) and uranium, U (ppm) concentrations and the total count rate (counts sec\(^{-1}\)), which combines the Th, U and K emissions using standard corrections (Hodgson and Ture, 2013; IAEA, 1991). The gamma rays originate from the upper ~ 0.3 m of Earth’s surface and attenuate exponentially with increased flight elevation at a rate of ~ 0.5 per 100 m. Radiometric data above 100m were discarded.

The AM data were collected at spatial intervals of 7 m (TEL) and 6 m (TELB). Standard processing techniques (Blakely, 1995), including horizontal derivative filtering, have helped to distinguish variations in bedrock structures (e.g. granites, basaltic dykes and sediment), associated with variable magnetic mineral content and magnetisation.

2.3.3 Description

Airborne Magnetic

Figure 2.2a shows the AM first horizontal derivative map with interpreted magnetic lineaments. The igneous complexes of the Slieve Gullion, Carlingford and Mourne Mountains are all distinguished. The basaltic dykes associated with the magnetic lineaments corroborate the Cooper et al. (2012) and Anderson et al. (2016) studies demonstrating four dyke swarms in this region, regularly displaced by strike-slip faulting and reactivated ENE-WSW Caledonian faults.
Figure 2.2: Airborne data sets of (a) 1st horizontal derivative of airborne magnetic data (nT km\(^{-1}\)) with magnetic lineaments and known geological contacts; lithological boundaries and igneous complexes are indicated (Figure 2.1a), (b) apparent conductivity from 3.005 kHz electromagnetic data and (c) ternary image of K (%), Th (ppm) and U (ppm) concentrations from airborne radiometric data; note 5km by 5km area (black box) from Figure 2.1c, and (d) total count (of gamma radiation) sec\(^{-1}\) encompassing the area of the Dromiskin Gravels groundwater body (white), overlain on CORINE agricultural pastures boundaries (see text).
**Airborne Radiometric**

Natural gamma ray emissions originating from the upper ~ 0.3 m of the Earth’s surface can be used to infer K (%), Th (ppm) and U (ppm) concentrations in a standard element ratio map or ternary image (Figure 2.2c). Dempster et al. (2013) have shown that the geochemical composition of the upper surface of tills shows a close relationship to the local bedrock, a correlation observed within the radiometric data (Figure 2.2c), e.g. the high K, Th and U emissions over the Mourne Mountains. Increased moisture content in soils increases soil density, leading to increased attenuation of gamma rays (Carroll, 1981), so that near-surface water, rivers and lakes, and saturated peat appear black in the ternary image.

The Dromiskin Gravel (DG) groundwater body is the only groundwater body in the region characterised by low gamma ray emissions (Figure 2.2c) and low total counts (Figure 2.2d). These low values are probably associated with a lack of gamma ray sources in clean sands (< 1% clay content) in the top few metres (Section 2.5). Across the DG groundwater body, total count rates < 1175 counts sec\(^{-1}\) are generally associated with higher elevations of 4 - 8 m OD and correlate with areas designated as pasture land within the CORINE (EEA and EPA, 2007) land classification (Figure 2.2d). Higher count rates generally correlate with non-irrigated arable land and topography < 4 m OD, e.g. near the coast, possibly due to surface water and groundwater deposition of natural radioactive sources (Wilford et al., 1997) or clay eluviation and movement (Dickson and Scott, 1997) in well-drained surface soils.

**Airborne Electromagnetic**

The AEM apparent conductivity data are poor at distinguishing bedrock geological structures (compare Figures 2.1b and 2.2b) but do discern conductive soils, such as alluvium and peat. They show hitherto unmapped, NE-SW Caledonian-trending lineaments, identified by higher apparent conductivities (Figures 2.2b, 2.3a, 2.3b), related to conductive bands of bedrock (e.g., shale, mudstone), similar to NE-SW trending conductive (>30 mS m\(^{-1}\)) bands across the predominantly unexposed Moffat Shale Group to the northwest (Cooper et al., 2014; Hodgson and Ture, 2014).
The peat, alluvium and the Irish Sea Till (Teagasc et al., 2006) are characterised by conductivities > 10 mS m\(^{-1}\) (Figure 2.3b). The DG groundwater body exhibits a large degree of heterogeneity across all frequencies (Figure 2.3c). Increased apparent conductivities (30 to > 500 mS m\(^{-1}\)) occur along the coast and are indicative of saline influence in the coastal sediments, extending ~750m inland along the eastern edge of the DG groundwater body (Figure 2.3c).

Figure 2.3: (a) apparent conductivity derived from 11.962 kHz airborne electromagnetic data; lithological boundaries are from Figure 2.1b; locations of areas in Figures 2.3b and 2.3c are shown, (b) apparent conductivity showing northeast-southwest conductive lineaments, and values > 10 mS m\(^{-1}\) associated with peat and alluvium soils, and (c) apparent conductivity across the Dromiskin Gravels showing the locations of three electrical resistivity tomography (ERT) profiles (1, 2 and 3) and a conductive linear feature (X) at ERT 1, extending inland for ~2 km, with an absence of data (grey area) due to the presence of a motorway and railway.
An electrical resistivity tomography profile (ERT1, Figure 2.3c), was recorded through this feature as part of a ground truthing campaign (O’Connell and Daly, 2013), coincident with a Tellus flight line. The profile transverses a linear, conductive (~ 20 mS m^{-1}) feature (X, Figure 2.3c), (most prominent in the 24.510 kHz data and unrelated to cultural noise, extends ~ 2 km inland from the coast). The ERT data (Figure 2.4d) were inverted using Res2dinv 3.58 (Geotomo Software, 2010) to produce a two-dimensional (2-D) electrical conductivity model with a root mean square (rms) error between the data and the theoretical response (Figure 2.4e) of < 8% after 7 iterations. The model suggests three conductivity layers (Figure 2.4f). A shallow (0-5 m) layer with conductivities <25 mSm^{-1}, an intermediate layer 10-20m thick with conductivities of ~50 mSm^{-1} and underlying low conductivities <5 mSm^{-1}, suggesting dry silty sands in the top layer, increasing clay content in the middle layer and bedrock below 20m. Preliminary interpretations indicate these values are consistent with results from other similar regions in Ireland (e.g., O’Connor, 1998; O’Connor, 2012; Pellicer et al., 2012). Therefore ERT1 indicates that the linear feature observed on the AEM data is restricted to the sediments above bedrock as there are no conductive features below ~ 15 m and may represent a palaeo-inlet prior to reclamation from the sea.

The inversion of the AEM data coincident with ERT1 was carried out using a well-established inversion technique (Beamish, 2012; Farquharson et al., 2003; Viganotti et al., 2013). The 1-D inversion algorithm (University of British Colombia, 2000, 2006) uses an a priori conductivity starting model and iterates to the simplest model that fits the observations within an uncertainty, assumed to be equivalent to system noise of 20-30 ppm (Leväniemi et al., 2009), with additional errors to account for modelling uncertainties equivalent to 5% of the in-phase and quadrature components. Preliminary AEM inversions with a starting model of a uniform sub-surface electrical conductivity demonstrated that data from the three highest frequencies could not resolve conductivity structure > 40 m below surface, in agreement with the interpretation of Hodgson and Ture (2014). The inclusion of noisy 0.912 kHz data did not improve this so these data were
not used in subsequent inversions. The AEM data (Figures 2.4a and 2.4b) were inverted using four variants of the inversion algorithm (University of British Colombia, 2006) to test model parameters. Appendix 1 demonstrates the misfits for dipole moment factors of 1.0, 1.8, 2.0 and 2.2. The models required the half-space to be > 30 m below ground level. Misfit levels between the AEM data and the theoretical responses ranged from 7 to 35 ppm with the highest misfits in the conductive zone adjacent to the coast. Models along a profile over seawater > 40 m deep estimated seawater conductivities from 2350 to 3900 mS m\(^{-1}\) for the upper 8 m of the water column and decreasing at greater depth, in line with results from (Beamish, 2004). The misfit values over the DG groundwater body are typical of those in Figure 2.4, i.e. 70% of all values are less than the average value of 17 ppm. The average rms error between all observed AEM data and their theoretical responses over the DG groundwater body is ~8-10%.

Figure 2.4: Airborne electromagnetic (AEM) and electrical resistivity tomography (ERT) inversions comprising (a) observed and modelled AEM in-phase components, (b) observed and modelled AEM quadrature components, (c) electrical conductivity model inferred from AEM data along ERT1, (d) measured ERT1 apparent conductivity pseudosection, (e) theoretical ERT1 apparent conductivity pseudosection, and (f) inverted ERT1 conductivity model. The ERT1 consist of data from a Wenner-Schlumberger configuration using an Iris Syscal Pro system with 4 multicore cables, each with 12 electrode connections and 45 electrodes with separations of 10 m.
A smooth conductivity model consisting of 14 layers with thicknesses increasing from 2m to 4m with increasing depth to 42 m, a fixed trade-off parameter of 6, a reference conductivity model of 0.5 mS m\(^{-1}\) (i.e. half-space conductivity below 42 m) and a transmitter dipole moment factor of 2.0 provided the lowest misfit between the theoretical responses and observed data. The effect on the inferred electrical conductivity from a combination of the model misfit, due to a violation of the 1-D conductivity assumption, and noise levels in the AEM data was tested by randomly perturbing the inverted conductivity estimates. This suggested that inverted conductivities are probably accurate to within ±7% in the depth ranges 0 – 5 m and > 20 m, and within ±10% in the intermediate range.

The agreement between the AEM and ERT1 conductivity models is shown in Figures 2.4c and 2.4f. The models demonstrate a standard result that the conductivity-thickness product of the conductive layer is the best resolved parameter with the main difference between the AEM and ERT models in the upper 5m. Rainfall data, obtained for the nearest weather station i.e. Dublin Airport (Met Éireann, 2016) indicated that the AEM data, acquired in January and February of 2012, was collected following average rainfall of ~60 mm/month and low average evapotranspiration (~18 mm/month) for the months prior to the survey. The area around the Dromiskin Gravel water body was partially covered with surface water so it is likely that the uppermost layers were fully saturated during the survey period.

In conjunction with additional ground-based ERT surveys and sparse hydrogeological information, AEM modelling was extended across a 5 x 5 km area containing the DG groundwater body to examine the potential to use the electrical conductivity distribution to examine hydraulic properties within the gravel aquifer, specifically targeting a commercial farm, located to the northeast of ERT1. One-dimensional (1-D) electrical conductivity versus depth models were constructed from the in-phase and quadrature components at 15,726 of the 21,974 AEM sounding locations (after the elimination of data at altitudes > 100m and data contaminated by cultural noise). ERT2 was acquired coincident with an AEM flight (for locations see Figure 2.3c) line in July 2015 after a prolonged period of low rainfall (~40 mm/month) and high evapotranspiration (~85 mm/month). A borehole was
cored in June 2016 and will be discussed in more detail in section 2.3.5. ERT3 (Figure 2.3c) was recorded 10m west of the borehole at the time of coring, running ~60 m east and parallel to ERT2, mirroring the flight lines of the AEM survey and still with the zone of influence of the downward propagating electromagnetic signal. ERT3 was acquired after rainfall of ~50 mm/month and evapotranspiration of ~ 80 mm/month.

Figure 2.5: Inverted conductivity-depth profiles for AEM and ERT profiles near BH1. (a) AEM flight line, (b) ERT2, (c) ERT3. The lithology at BH1 can be summarised as predominantly fine sand in the top 7 m, clay between 7 m and 11.3 m, and fine-medium gravel below (Table 1).

The inverted electrical conductivity as a function of depth from the AEM data, with the collinear ERT2 and offset ERT3 acquired with 48 electrodes at 4 m separations, are shown in Figure 2.5. ERT2 (Figure 2.5b) confirmed that the 14 layer smooth conductivity model described above provided the lowest misfit between the theoretical responses and observed data for the 5 x 5 km area. All three profiles indicate varying low conductivities in the upper 5m, over 10-15 m of conductive material with underlying low conductivities indicative of bedrock similar to ERT1 (Figure 2.4). The
variations in rainfall and ground saturation for each ERT profile are evident in all models in the upper 5m.

2.3.5 Petrophysical Data

Laboratory data for the effective electrical conductivity of sediment and soil demonstrate that electrical conductivity is a function of pore fluid salinity, temperature, inter-connected pore space and its saturation, volumetric water content, specific surface area at the grain-fluid interface, surface electrical conductivity due to the distribution and composition of clay minerals, compaction and other geotechnical parameters (Archie, 1942; Kalinski and Kelly, 1993; Revil et al., 1998; Rhoades et al., 1976; Shah and Singh, 2005; Waxman and Smits, 1968). Therefore in order to model the electrical conductivity with depth function a borehole (BH1) was also completed at the farm site, from which soil samples were recovered for analysis, and falling and rising head tests according to the BS5930 (1999) were completed in the borehole to assess the vertical distribution of hydraulic conductivity.

A 200mm diameter shell and auger borehole was completed to a depth of 12.3m (EOH) in July 2016 encountering three distinct sediment layers. The particle size distribution of samples recovered (Table 3) indicates that from the surface to a depth of roughly seven metres is dominated by fine sand with < 1% clay, a thick unit of tightly packed clay (~ 50%) with significant amounts of fine-medium silt lies between seven and 11.3m, beneath which is fine-medium gravel with limestone pebbles to EOH at 12.3 m (bedrock was not encountered). These layers are in good agreement with ERT3 (see Figure 2.5c) which indicates resistive values concurrent with the sand and gravel layers, with conductive values associated with the clay layer. Sediment samples (~0.5kg) were recovered at one-metre intervals from the surface to the EOH. A sub-set of these samples was used to measure gravimetric water content, cation exchange capacity (CEC) and particle size distribution as a function of depth (Table 3).
Table 3: Moisture, cation exchange capacity and particle size distribution in borehole 1. The gravimetric water content is defined by the mass of water divided by the mass of sediment (water + sediment particles); the mass of water was determined from the difference in mass before and after drying the sample in an oven for more than 4 hours between 105 – 110°C until all water was removed. The particle size distribution is estimated from the percentage of the mass of solid sediment particles after drying was obtained using a combination of wet sieving and sedimentation procedures. Water content and particle size distribution measurements followed BS 1377 (1996). The CEC measurements are expressed as C kg⁻¹ of dry solid particles following procedures described in Ross (1995) based on determination by the BaCl₂ Compulsive Exchange Method (Gillman and Sumpter, 1986).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Water (%)</th>
<th>CEC (C kg⁻¹)</th>
<th>PSD (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>mm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.063-0.125</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.020-0.063</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.006-0.020</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.002-0.006</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>&lt;0.002</td>
</tr>
</tbody>
</table>

Dry, sieved samples with effective diameter < 100 μm and mass of ~ 1 g from borehole sample depths of 3 m, 10 m and 11.8 m were powdered for use in a Bruker D5000 X-ray Diffraction (XRD) instrument to obtain their bulk composition. Powdered samples with a mass of ~ 3 g from each depth were also placed in cylinders filled with ~50 ml of de-ionised water and 10 drops of a dispersant. These were shaken, placed in an ultrasonic bath for 5 minutes, left to settle for 3 hours, and ~ 1 ml of slurry was then extracted from 39 mm below the surface. This contains the clay fraction < 2 μm which
was then pipetted onto a low-background Si-wafer mount, and air-dried overnight before using the XRD in steps of 0.01° from 3 – 30° (2θ) to obtain information on the concentrated clay minerals.

Each of the 1 g samples was subsequently heated to 550 °C for 1 hour and allowed to cool so that the XRD experiment could be repeated to determine if kaolinite, which is not stable at 550 °C, is present. Each of the samples on the Si-wafer mounts was solvated in ethylene-glycol vapour overnight at 60 °C and then re-run on the XRD to investigate if expandable clay minerals (smectite, montmorillonite, vermiculite) are present.

The XRD trace was interpreted using Bruker’s phase-matching software and the industry-standard (International Centre for Diffraction Data) Powder Diffraction File (4 3+). The traces from each bulk sample and interpreted phase information were compared with theoretical XRD traces whose parameters were optimised to fit each measured trace using Rietveld refinement software (Sietronics Siroquant, 4.0 (5), with database 3.0g) to give concentrations of minerals (mass percent), normalised to 100% (Table 4).

Table 4: Rietveld refined semi-quantitative results. The merit of the fit is quantified by $\chi^2$ and the R-Factor, which were 2.21 – 2.76 and 0.30 – 0.36, respectively. Their values suggest that the mineral concentrations are probably accurate to < 10%.

<table>
<thead>
<tr>
<th>Depth m</th>
<th>Quartz</th>
<th>Albite</th>
<th>Calcite</th>
<th>Mica/Illite</th>
<th>Chlorite</th>
<th>Orthoclase</th>
<th>Dolomite</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>3</td>
<td>51.7</td>
<td>13.7</td>
<td>10.1</td>
<td>9.3</td>
<td>8.8</td>
<td>4.6</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>28.3</td>
<td>13.4</td>
<td>8.5</td>
<td>21.0</td>
<td>26.1</td>
<td>-</td>
<td>2.8</td>
<td></td>
</tr>
<tr>
<td>11.8</td>
<td>32.6</td>
<td>14.5</td>
<td>8.0</td>
<td>22.1</td>
<td>19.1</td>
<td>-</td>
<td>3.8</td>
<td></td>
</tr>
</tbody>
</table>

The samples are not interpreted to contain expandable clays since the diffraction trace of the ethylene-glycol solvated <2 samples shows no significant peak shift from their air-dried state. A comparison of the traces for the bulk samples before and after heating shows small differences consistent with the presence of chlorite, although the presence of low
concentrations (< 1%) of kaolinite cannot be ruled out. The crystallinity of the 10Å (8.6° 2θ) mica peaks places the samples in the low to high anchizone and at this grade it is likely that the mica is a mixture of illite and muscovite. Illite is present (but in unknown quantity) as the samples show small (10Å) shifts in the peak position between the air-dried and glycolated traces; muscovite should show no change in peak position on glycolation.

2.4 Hydrogeophysical Interpretation

The approach to interpretation uses the 1-D electrical conductivity function of depth, $\sigma(z)$ (from ERT3), clay mass fraction, $\psi(z)$ and the electrical conductivity of groundwater at one location (BH1) to place bounds on the inter-connected porosity, $\varphi(z)$. The porosity so obtained depends on the choice of the electrical conductivity model relating connected porosity, clay content and pore fluid electrical conductivity to the macroscale electrical conductivity. All of these parameters can provide estimates of hydraulic conductivity as a function of depth (Section 2.5).

2.4.1 Theory

The electrical conduction is assumed to be ohmic in the pore volume and near the grain-fluid interface for the frequency range of the AEM (and ERT) data (Rinaldi and Cuestas, 2002). There are many models for the prediction of macroscale electrical conductivity for such porous media e.g. Glover (2010). We use two models with a minimum number of free parameters to account for the electrical conductivity contribution of the pore fluid and clay content. The first model uses an empirical addition law, established for shaly sediment by Waxman and Smits (1968), which assumes that small clay particles are dispersed in the pore space between coarser grains. The macroscale electrical conductivity (S m$^{-1}$) may be written as:

$$\sigma = (\sigma_f + BQ_v)\varphi^m = (\sigma_f + BQ_v)F^{-1}$$  \hspace{1cm} (1)$$

where $\varphi$ is porosity, $m$ is cementation exponent, $Q_v$ is the excess surface charge per unit pore volume (C m$^{-3}$), $\sigma_f$ is the fluid electrical conductivity (S m$^{-1}$), $F$ is the formation factor, $B$ is the equivalent mobility (m$^2$ s$^{-1}$ V$^{-1}$) of
counter-ions close to the grain surface given by Revil et al. (1998), Equation 19, as:

\[ B = 4.78 \times 10^{-8} \left[ 1 - 0.6 \exp \left( \frac{-\sigma_f}{0.013} \right) \right] \]  

(2)

This simple addition law has been verified for clayey soils by Rhoades et al. (1976) and other porous media (e.g. Johnson and Sen (1988)), and corroborated with numerical modelling by Tabbagh et al. (2002).

Some pore scale numerical models can also account for volume electrical conductivity in the pore fluid and surface electrical conductivity at the grains interface by assuming that the clay is distributed as a coating on the surface of synthetic spherical grains (Devarajan et al., 2006) or on sand grains (Zhan et al., 2010). This prompts the use of a second model, the differential effective medium model (DEM) established by Bussian (1983) and modified by Revil et al. (1998). It uses clay-coated spherical grains and the behaviour of positive and negative ions in the medium. Revil et al. (1998) present an expression for this macroscopic electrical conductivity:

\[
\sigma_+ = t_+^f \sigma_f \Phi^m [1 - t_+^s \sigma_s/t_+^f \sigma_f]^m [1 - t_+^s \sigma_s/t_+^m \sigma]^m
\]

\[
\sigma_- = t_-^f \sigma_f \Phi^m
\]

(3a)

\[
\sigma = \sigma_+ + \sigma_-
\]

(3c)

where \(\sigma_s\) is the surface electrical conductivity (S m\(^{-1}\)) representing conduction due to hydrated clay mineral exchange cations at different salinities in the electrical double layer coating the large grain, and \(t_{\pm}^f\) and \(t_{\pm}^s\) are Hittorf transport numbers for pore fluid and surface conduction, with \(t_{\pm}^f = 0.38, t_{\pm}^s = 0.62, t_+^s = 1, t_-^s = 0\). The surface electrical conductivity can be written as:

\[
\sigma_s = \frac{2}{3} \frac{\frac{\Phi}{1 - \Phi}}{Z_s \beta_s Q_v} = \frac{2}{3} \rho_g \beta_s CEC
\]

(4a)

\[
Q_v = \rho_g \frac{1 - \frac{\Phi}{\Phi}}{CEC}
\]

\[
\beta_s(T) = \beta_s(25) \left[ 1 + \nu_s(T - T_{25}) \right]
\]

(4c)
\[ CEC = \psi \sum_j \psi_j CEC_j \]  

where \( Z_s = 1 \) is the valence of counter-ions, assumed to be \( Na^+ \), \( \beta_s(25) = 5.14 \times 10^{-9} \) (m\(^2\) s\(^{-1}\) V\(^{-1}\)) is the surface mobility at temperature \( T_{25} = 25^\circ C \), \( \nu_s(Na^+) = 0.04 \) (m\(^2\) V\(^{-1}\) s\(^{-1}\)) is a temperature coefficient, \( T = 12 \) (C) is the \textit{in situ} temperature assumed in this paper, \( \rho_g = 2650 \) kg m\(^{-3}\) is the grain density for sand and clay minerals without their bound water, CEC is the cation exchange capacity (C kg\(^{-1}\)), the maximum number of exchangeable surface cations per unit mass of sediment, the mass fraction \( \psi \) of clay in a mixture of sand and clay, and \( \psi_j \) the relative mass fractions of the minerals with cation exchange capacity, \( CEC_j \) in the clay fraction. Approximate variations in \( CEC_j \) (Revil et al., 1998, Figure 1) for kaolinite, chlorite and illite are \( 3000 < CEC_k < 10000 \), \( 2000 < CEC_c < 10000 \), \( 10000 < CEC_i < 40000 \) (C kg\(^{-1}\)).

Equations 1 to 3 are appropriate for fully-saturated porous media. If the medium is partially saturated, these equations are modified using the transformations introduced by Waxman and Smits (1968):

\[ \varphi^m \rightarrow S^p \varphi^m; \quad Q_v \rightarrow \frac{Q_v}{S} \]  

where \( S \) is the fractional fluid saturation of the pore space and \( p \) is the second Archie exponent, \( 1.6 \leq p \leq 2.2 \) (Waxman and Thomas, 1974), and assumed in this paper to have the value \( p = 2 \). The second transformation implies \( \sigma_s \rightarrow \sigma_s/S \).

### 2.4.2 Porosity determination

Porosity is determined from the conductivities inverted from data along ERT3 (Figure 2.5c) using the techniques outlined in Section 2.4. The conductivities near the centre of ERT3, and closest to BH1, at the centres of 14 layers at depths 1, 3, 5...27 m corresponding to the layers used in the AEM inversion, are used in the petrophysical inversion. Each sample water content from BH1 was converted to a mass fraction, \( \vartheta_m \) and the \textit{in situ} CEC at each sample depth is estimated using \( CEC_is = CEC_d (1 - \vartheta_m) \), where \( CEC_d \) is the CEC of the dry particles in a sample. The water content and \textit{in situ} CEC were interpolated/extrapolated to depths corresponding to
the centre of the layers in ERT3. Using a non-linear least-squares curve fitting algorithm, porosity is estimated in each layer, given the moisture content and CEC (Figure 2.6a).

Figure 2.6: Petrophysical model for ERT3. (a) Interconnected porosity from inversion of ERT3 data using the DEM model with saturated upper layers (black), saturated W-S model (red), and interpolated/extrapolated borehole sample measurements of clay (magenta circles) and water content (blue circles), (b) surface electrical conductivity for W-S and DEM models (blue), electrical conductivity using a saturated DEM model (black dash) for the ERT3 conductivity (error bars). If the pore space estimated from this model is \(~ 80\%\) saturated in the top 5 m, electrical conductivity (black line) decreases by \(~ 20\%\) (see Discussion).

The theoretical electrical conductivity profiles from the W-S and DEM petrophysical models fit the inverted conductivity-depth profile almost perfectly (Figure 2.6b). Both models use a constant fluid electrical conductivity with depth of \(\sigma_f = 100 \text{ mS m}^{-1}\), measured on the same day as the acquisition of BH1 samples and ERT3. Both models assume that the exponent \(m = 2.2\) an average for chlorite and illite in the clay layer between \(~ 7-11 \text{ m}\) (Table 3) and \(m = 1.7\) in the relatively clay-free top and bottom layers; this is consistent with an increase in the exponent with surface electrical conductivity due to clay (Figure 5 in Revil et al., 1998). The observed clay mass fraction (Table 3) and CEC as a function of depth (Table 4), and the inversion of ERT3 conductivity data using the DEM model require the mica/illite fraction to be \(> 90\%\) illite with \(CEC_{\text{ill}} \sim 39850 \text{ C kg}^{-1}\), and a chlorite fraction with \(CEC_{\text{cl}} \sim 9850\), values which are at the upper
end of their range reported in the literature. The CEC in the top and bottom layers with < 1% clay is ~ 450 C kg\(^{-1}\). The estimated interconnected porosity (Figure 2.6a) correlates with BH1 and ERT3, indicating increased porosities in the upper fine sand layer, decreasing to <5% within the clay layer.

### 2.5 Hydrogeological Inference

Prior to this study, ground-based surveys (GSI and RBD Consultants, 2005) indicated that the DG groundwater body consists of marine sands, gravels and < 8% fine materials (clay), with a high water table close to the surface, variable saturation, variable hydraulic conductivity (\(K\)) and porosity of ~ 30% at the surface. Though surface porosity observations are similar to the modelled observations in Figure 2.6a, this study indicates ~15% clay fines in BH1 (Table 3). The only information (prior to this study) on clay-type in the region surrounding the DG body suggests that illite is dominant (O’Connor, 2012). This paucity of quantitative data is not unusual so predictions of hydrogeological parameters, particularly hydraulic conductivity, from electrical conductivity are potentially important. There are, however, difficulties in relating the microscale parameters (e.g. pore geometry, tortuosity, CEC, surface electrical conductivity, clay compaction) to the macroscale properties of electrical current flow and water flow (Abu-Hassanein et al., 1996; Hunt, 2001; Sanchez-Vila et al., 2006; Slater, 2007). Purvance and Andricevic (2000) discuss electrical and hydraulic conductivity (\(e-h\)) correlations based upon microscale network models with connected pore volume and pore surface area (Bernabe and Revil, 1995; Wong et al., 1984). The hydraulic conductivity, \(K\), can be related to electrical conductivity (Purvance and Andricevic, 2002) by a power law, \(K = a e^b\), where \(a\) and \(b\) are constants. Pore volume electrical flow in high-salinity or clay-free environments implies positive \(e-h\) correlations (\(b > 0\)); pore surface-dominated electrical flow in low-salinity clay-rich environments implies negative \(e-h\) power law correlations (\(b < 0\)). Recent microscale network models also support \(e-h\) correlations (Skaggs, 2011; Zhan et al., 2010). This section utilises three predictive models for estimating permeability, \(k\), and hydraulic conductivity.
Chapter 2

2.5.1 Theory

Model 1 is based on literature derived from critical path analysis (CPA) approaches linking water and electrical flow transport via a characteristic length scale because permeability has dimensions of area and depends on the absolute dimensions of pore space (Banavar J.R. and Johnson, 1987; Katz and Thompson, 1986; Le Doussal, 1989; Skaggs, 2011):

\[
K = \left(\frac{\rho_f g}{\mu}\right)_k = \left(\frac{\rho_f g}{\mu}\right) \frac{\sigma}{\alpha_f} c \delta_c^2
\]  

(6)

where \(\rho_f = 1000 \text{ kg m}^{-3}\) is the fluid density, \(g = 9.81 \text{ m s}^{-2}\) is acceleration due to gravity, \(\mu = 0.0014 \text{ kg m}^{-1} \text{s}\) is fluid dynamic viscosity and \(\delta_c\) is a characteristic length related to the mean pore size. The parameter \(c\) depends upon, \(\text{inter alia}\), pore geometry (cylinders, slits), network and pore size distribution so (6) can be computed with \(c = 1/87, 1/57, 1/27\) to estimate \(k\); for a regular network of slit-shaped pores \(1/27 > c > 1/57\) appears to satisfy experimental data better than a network of cylindrical pores (Skaggs, 2011).

Model 2 is based on a generalisation of the Kozeny–Carman equation for an isotropic, homogeneous binary sphere pack consisting of spherical grains where the smaller spheres represent the clay and fine silt fraction, assumed to be about two orders of magnitude smaller than the coarse grains (Leurer and Brown, 2008; Mavko et al., 1998). The permeability is:

\[
k = B \frac{\varphi_3^3}{1 - \varphi^2} d^2
\]  

\[
d = \left(\frac{(1 - c_s)}{d_l} + \frac{c_s}{d_s}\right)^{-1}
\]  

(7a)

(7b)

where \(B = 0.007\) (Hovem and Ingram, 1979), and \(d_s\) and \(d_l\) are the average grain diameters of the small and large spheres respectively, \(d\) is the effective grain size and \(c_s\) is the small sphere volume fraction.

Model 3 is based on relationships for clayey-sands between permeability, porosity, mean sand grain diameter and the CEC/electrical tortuosity exponent of clay minerals, the last of which acts as a proxy for clay grain diameter (Revil et al., 2002):
\[ k = (k_{sd})^{1-c_{cl} / \varphi_{sd}} (k_{cfs})^{c_{cl} / \varphi_{sd}}; \quad 0 \leq c_{cl} \leq \varphi_{sd} \]  
(8a)

\[ k_{sd} = \frac{d^2 \varphi_{sd}^{9/2}}{24} \]  
(8b)

\[ k_{cl} = k_{cl0} \left( \frac{\varphi_{cl}}{\varphi_0} \right)^{m_{cl}} \left( \frac{\varphi_0 - m_{cl} - 1}{\varphi_{cl} - 1} \right)^2 \]  
(8c)

\[ k_{cfs} = k_{cl} \varphi_{sd}^{3/2} \]  
(8d)

\[ c_{lv} = \left( \frac{c_{lv}}{1 - c_{lv}} \right) \left( \frac{1 - \varphi_{sd}}{1 - \varphi_{cl}} \right); \quad 0 \leq c_{lv} \leq c_{cl_{crit}} \]  
(8e)

where \( k_{sd} \) and \( k_{cl} \) are permeability in the sand and clay end-members, \( k_{cfs} \) is the permeability of a clay-filled sand, \( \varphi_{sd} \) and \( \varphi_{cl} \) are porosity in the sand and clay end members, \( m_{cl} \) is the clay electrical tortuosity (assumed equal to the exponent \( m \)), \( c_{lv} \) is the clay volume fraction (\( < \varphi_{sd} \)) and \( k_{cl0} \) is the permeability of the clay end-member at the reference porosity of 50\% (Figure 7 in Revil et al. (2002)), and \( k \) is the permeability of the clay-sand mixture.

For the sediment in BH1, the critical clay weight fraction is assumed to be 0.50 so that the top and bottom layers are in the clayey-sand domain while the middle layer is in the sandy-clay domain. The permeability of the clay end members at a surface porosity of 50\% are \(~9.87 \times 10^{-11} \) m\(^2\) and \(~4.7 \times 10^{-13} \) m\(^2\) for kaolinite and illite respectively (Revil et al., 2002, Figure 7). In the absence of an equivalent value for the permeability of chlorite, it is assumed to be \(~6.8 \times 10^{-12} \) m\(^2\), the geometric average of the kaolinite and illite permeability. As the illite and chlorite mass fractions are similar, the geometrically-averaged permeability for their mixture is assumed to be \(1.0 \times 10^{-12} \) m\(^2\) at a surface porosity of 50\%.

For the three layers observed in BH1, and imaged on the AEM and ERT inversions (i.e. an upper fine sand layer to 7 m, a clay and medium to fine silt layer to 11.3 m, underlain by a fine to coarse gravel layer), the CPA network model (Skaggs, 2011) with \( c = 1/57 \) fits the pattern of observed hydraulic conductivity for characteristic lengths of \( \delta_c = 0.15, \delta_c = 0.0001, \delta_c = 0.4 \) mm in the respective layers (Figure 2.7a). While the
The significance of the characteristic scale length in CPA models is not certain, the critical pore diameter is ~ 1/6 grain diameter for perfect spheres (Stoll, 1989). If this is valid for the CPA network, it implies a top layer of coarse sand, middle layer of clay and fine silt, and bottom layer of fine-medium gravel. The Revil et al. (2002) model requires a particle size of 0.9 mm for the top layer (coarse sand), 0.2 to 0.5 mm for the middle layer (medium sand) and ~ 6 mm for bottom layer (fine-medium gravel). The Kozeny-Carman binary grain size model requires large grain/small grain sizes of 1 mm/0.03 mm for the top layer (coarse sand/coarse silt), 0.35 mm/0.01 mm for the middle layer (medium sand/medium silt) and 32 mm/0.063 mm for the bottom layer (medium-coarse gravel/coarse silt). In general, all models are indicating coarse sand in the top 7 m consisting of sediment dominated by fine sand, with smaller amounts of medium sand. The CPA model fits the middle clay layer and the Kozeny-Carman binary grain size model predicts the lithology of the bottom layer most accurately.

Figure 2.7: (a) Hydraulic conductivity as a function of depth using porosity estimated from the saturated DEM electrical conductivity model, and Equation (6) (magenta), Equation (7) (black), and Equation (8) (blue). The hydraulic conductivities measured at BH1 (red circles) using falling head or rising head methods according to the BS5930 (1999) Code of Practice for Site Investigations. Saturated hydraulic conductivity range (error bar) of clay (e.g. Domenico and Schwartz, 1998) is shown for comparison. (b) Inverted electrical conductivity-depth profiles based on AEM and ERT data.
Typical saturated hydraulic conductivity values for fine, medium and coarse sands are in the ranges \(2 \times 10^{-7} - 2 \times 10^{-4} \text{ m s}^{-1}\), \(9 \times 10^{-7} - 5 \times 10^{-4} \text{ m s}^{-1}\) and \(9 \times 10^{-7} - 6 \times 10^{-3} \text{ m s}^{-1}\) respectively (Domenico and Schwartz, 1998). Falling head tests completed in the borehole give hydraulic conductivity values in the top 6 m is \(\sim 3.0 - 3.2 \times 10^{-4} \text{ m s}^{-1}\), consistent with the models and typical medium-coarse sand values and at the typical upper end of values for fine sand (see particle size distribution in Table 3). Hydraulic conductivity in the unsaturated zone varies with effective saturation (and pressure) so that the unsaturated hydraulic conductivity, \(K_u\) is related to the saturated hydraulic conductivity by \(K_u = K_r K_s\), where the relative hydraulic conductivity, \(K_r\), has a value between 0 and 1 (Mawer et al., 2015). If the upper roughly six metres of sediment is partially saturated, and if the falling head experiment is appropriate in these conditions, then it becomes more difficult to reconcile the particle size distribution with saturated hydraulic conductivity models.

The AEM and ERT data presented above show some interesting differences that are probably related to saturation in the uppermost layers. A comparison of inverted 1-D electrical conductivity (Figure 2.7b) with the modelled data from ERT2 and ERT3, suggests that the AEM data are more conductive to a depth of roughly five metres below the surface, suggesting that this zone may have been partially saturated during the acquisition of the ERT data, particularly the ERT2 data, as outlined in Section 2.3.4.

While the resolution of shallow (< 5 m) structure is relatively poor with these AEM data, forward modelling of saturated and partially saturated layers does indicate that a 50% reduction in electrical conductivity of the first few model layers will change the theoretical AEM responses by \(\sim 10\%\), a value greater than the noise threshold. Figure 2.7b demonstrates that an assumption of 80% saturation in the upper five metres of a fully-saturated ERT3 model results in a reduction of electrical conductivity of \(\sim 20\%\), a figure comparable with estimates of variations observed at sites in a climate similar to Ireland. The magnitude of the difference with the conductivity in the uppermost layers of ERT2 suggests lower saturation levels and/or different near-surface lithology than those observed at BH1/ERT3. The rainfall data discussed in Section 2.5.3 would suggest lower saturation
levels. This highlights the additional work required to calibrate ERT with AEM data when not acquired at the same time. The prediction of hydraulic conductivity in potentially unsaturated zones like these could be partially addressed using ground-based EM induction techniques (Brosten et al., 2011; Rezaei et al., 2016) and a practical power law relationship $K_r = \sigma_r^{2.1}$ between the relative hydraulic conductivity and relative electrical conductivity, $\sigma_r$ (Mawer et al., 2015).

### 2.5.2 Up-scaling to catchment level using AEM

In principle, the approach presented here can be integrated with ground-based *in situ* hydrogeophysical and hydrological data to provide a preliminary up-scaling of hydrological parameters to local or sub-regional spatial scales commensurate with the size of typical groundwater bodies (10s m$^2$ – 10s km$^2$). However, the Dromiskin Gravel water body is a particularly challenging environment for this as there are significant cultural artefacts within its boundaries (the prominent motorway/railway corridor, in particular, Figure 2.3c). The variation in apparent conductivity (Figure 2.3c) also suggests the influence of saltwater intrusion in the coastal zone.

In order to estimate this variation in fluid electrical conductivity (and therefore salinity, an important property when assessing groundwater resources), the variation in inverted electrical conductivity in the first layer (0-2m) (Figure 2.8a) is normalised by assuming the formation factor for the first layer at BH1 is constant across part of the aquifer away from the artefacts. The resulting variation in fluid electrical conductivity is presented in Figure 2.8b. Both data plots indicate electrical conductivity increasing from west to east indicating increased salinity towards the coast. The variation in AEM electrical conductivity at a depth of 9 m (i.e. within the clay layer encountered in BH1) assuming a constant fluid electrical conductivity (100 mS m$^{-1}$) with depth, is plotted in Figure 2.8c. At this depth, elevated conductivities (>25 mS m$^{-1}$) in the vicinity of BH1 and locally throughout the groundwater body suggest the presence of clay lenses within the aquifer. It uses clay-coated spherical grains and the behaviour of positive and negative ions in the medium. Revil et al. (1998) present an expression for this macroscopic conductivity. The hydraulic conductivity at a depth of 5 m,
normalised to hydraulic conductivity at BH1, is plotted in Figure 2.8d and is in agreement with the measured value of \(\sim 4 \times 10^{-4} \text{ m s}^{-1}\). These model outcomes, warrant further interrogation, as they show the value, even in a complex system such as the DG groundwater body, of the use of the AEM data coupled with \textit{in situ} field data (geophysical, geological and hydrogeological).

Figure 2.8: (a) variation in inverted AEM electrical conductivity in the 1st layer (0-2 m), (b) variation in fluid electrical conductivity assuming the formation factor for the 1st layer is constant across the area, (c) variation in AEM electrical conductivity at a depth of 9 m assuming a constant fluid electrical conductivity (100 mS m\(^{-1}\)) with depth and, (d) variation in hydraulic conductivity at a depth of 5 m after normalising with fluid electrical conductivity and calibrated with hydraulic conductivity (\(\times 10^{-4} \text{ m s}^{-1}\)) at BH1.

From this study, the uncertainty in the predicted hydraulic conductivity depth function is a combination of the inadequacy of models relating
electrical current flow and fluid transport to each other and to realistic heterogeneous porous media, and a lack of in-situ data on pore fluid, clay content and composition, soil and sediment properties across the entire aquifer. In order to limit this uncertainty, the acquisition of data with an appropriate temporal and spatial resolution will require integrated ground-based geophysical surveys (Binley et al., 2015), exemplar catchment-scale observation strategies (e.g. Sanchez-Vila et al. (2006)) instrumented with stable, accurate and distributed in-situ sensor systems (Craglia et al., 2012; Selker et al., 2006) satellite remote sensing, e.g. for surface soil moisture (e.g. van der Velde et al. (2012)) and biogeochemical surveys. Airborne geophysical data and ground-based data, e.g. AEM and borehole (He et al., 2014), airborne magnetic data (Dickson et al., 2015) and AR data e.g. K band (Rawlins et al., 2009), with soil surveys are already contributing to scaling from local (~ km²) to regional (~ 100 km²) catchment level.

2.6 Conclusions
High-resolution airborne geophysical remote sensing data have been used in a regional reconnaissance of groundwater resources in a coastal zone area with a variety of soil types and water bodies. The focus has been on the modelling of airborne electromagnetic and ground-based electrical resistivity data to infer the sub-surface electrical conductivity distribution and hydrological properties, particularly hydraulic conductivity, of a clay-sand-gravel aquifer. The approach has used the electrical conductivity function of depth to infer interconnected porosity with depth constrained by a single borehole that has provided information on particle size distribution, water content and cation exchange capacity of the clay fraction as a function of depth.

The methodology can assist with catchment-scale three-dimensional quantitative assessments of groundwater resources in such aquifers only if there is a spatially extensive in-situ hydrological observation network and complementary ground-based geophysical, geochemical and borehole data.
Chapter 3: Terrestrial and marine electrical resistivity to identify groundwater pathways in coastal karst aquifers

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3.1 Abstract
Groundwater movement in karst aquifers is characterised by high-velocity fissure and conduit flow paths and, in coastal karst aquifers, these act as pathways for saline intrusion and freshwater discharge to the sea. This paper examines groundwater movement in two neighbouring catchments in the west of Ireland that represent canonical coastal karst aquifers dominated by discharges in the intertidal zone and at offshore submarine springs. Terrestrial and surface-towed marine electrical resistivity tomography, coupled with ancillary hydrogeological data, identify the influence of faulting and conduits on groundwater egress/saltwater ingress. The on-shore and off-shore sub-surface geometry of major fault zones is identified and the tidal influence of seawater and groundwater flow is demonstrated in these zones and karst springs. Imaging of these subsurface structures is a pre-requisite for numerical modelling of current and future climate-driven freshwater-seawater interactions in karst coastal aquifers.

Keywords
Hydrogeophysics, submarine groundwater discharge
3.2 Introduction

While the magnitude of future sea level rises are uncertain (Prime et al., 2015) the projected changes in global surface temperatures of between 1.1° and 6.4°C (IPCC, 2007) will lead to changes in sea levels that will have major impacts on coastal communities and water resources. Understanding how freshwater and seawater interact in coastal karst aquifers is critical for current and future water management in the context of future sea level changes and in the context of various EU Directives (Water Framework Directive, Floods Directive, Marine Strategy Framework Directive).

Karst aquifers are globally pervasive, acting as a source of drinking water to ~25% of the population (Ford and Williams, 2007). They are defined as soluble carbonate rocks characterised by high-velocity groundwater movement through fissures and conduits (Drew, 1999; Prohic, 1989). Well-developed karst terrains are typified by thin soil cover and an absence of traditional surface drainage (Van Beynen and Townsend, 2005; White, 1988), allowing rapid percolation of surface water and anthropogenic contaminants to the aquifer. In coastal settings the fissures and conduits can act as pathways for both freshwater discharge into the marine environment and salt water contamination of the aquifer (Taniguchi et al., 2002). A decline in the water table towards sea level coupled with more dense saltwater sinking beneath the freshwater forms a characteristic saltwater wedge into the karst aquifer, forcing much of the groundwater to discharge to the nearshore (Ford and Williams, 2007). Submarine groundwater discharge (SGD) refers to any and all flows of groundwater upwards across the sea floor (SCOR/LOICZ, 2004) occurring diffusely through seepage or at discrete springs. SGD can also occur offshore via submarine springs that are expressions of karst conduits developed in periods of reduced sea level (Bonacci and Roje-Bonacci, 1997; Fleury et al., 2007; Land and Paull, 2000).

Electrical resistivity tomography (ERT) has been used effectively in many hydrogeological applications, e.g. aquifer characterisation (Comte et al., 2012), groundwater/surface water interactions (Nyquist et al. (2008), seawater contamination (Comte and Banton, 2007; Khalil, 2006; Martorana et al., 2014; Nguyen et al., 2009), salinity studies of lakes and reservoirs.
(Amidu and Dunbar, 2008) and the detection of conduits for groundwater movement through typically heterogeneous karst regions (Zhu et al., 2011). It may be combined with more traditional methods for the investigation of SGD, e.g. monitoring of freshwater outputs using seepage meters, piezometers and natural (geochemical) tracers, water balance approaches, hydrograph separation techniques and theoretical analysis with numerical simulations e.g., (Burnett et al. (2006). In the coastal marine environment, ERT surveys using static submerged electrodes arrays have been employed to examine groundwater/seawater interactions at the water/seabed interface (Henderson et al. (2009); Swarzenski and Izbicki (2009)); Taniguchi et al. (2007). Surface-towed ERT systems, allowing rapid collection of data, have been used for monitoring groundwater flow in wetlands (Mansoor and Slater (2007), mapping aquifer recharge pathways (Kelly B.F.J. (2009) and the examination of diffuse SGD through sediments in the coastal zone (Belaval (2003); Breier et al. (2005); Manheim et al. (2004)). Surface-towed ERT for the detection of SGD depends on a variety of factors including discernible resistivity contrasts, and acquisition and processing procedures (Day-Lewis et al., 2006). While it is often difficult to obtain anything more than qualitative information from these profiles (Swarzenski and Izbicki, 2009), it is possible to achieve optimum resolution by, for example, obtaining data at periods of low tide when outflow of groundwater is maximum and focusing of the current in conductive saline water is minimum.

This paper examines two neighbouring catchments within a canonical coastal karst aquifer on the southern coast of Galway Bay in the west of Ireland (Figure 3.1) where SGD plays an important role in the coastal hydrogeology (Cave and Henry, 2011; McCormack et al., 2014; Perriquet et al., 2014; Smith and Cave, 2012; Wilson and Rocha, 2012). This paper is the first to use surface-towed ERT to examine SGD in a coastal karst aquifer setting. Existing hydrogeological data is integrated with new static terrestrial and surface-towed marine ERT to examine how the structural geology of the region in one catchment and a well-developed conduit network in a neighbouring catchment influence SGD and saltwater ingress across the coastal zone. It provides new evidence for previously predicted
terrestrial and submarine pathways that contribute to the total groundwater flux across the aquifer.

Figure 3.1: (a) Regional setting of southern Galway Bay showing the Bell Harbour and Gort-Kinvarra catchments and associated topography, (b) parts of the Bell Harbour and Gort-Kinvarra catchments showing selected hydrogeological and near-surface features (GSI, 2007a). The subterranean and submarine extension of MacDermot’s Fault and subparallel faulting, and location of terrestrial ERT profiles T1 to T4 are shown.

3.3 Geology and hydrogeology of Southern Galway Bay
The geology of southern Galway Bay in the west of Ireland (Figure 3.1) consists of Devonian Old Red Sandstone overlain by Lower Carboniferous
(Dinantian) limestone across the Burren Uplands, capped by Namurian shales in the southwest (Pracht, 2004). The sequence was subjected to compressive deformation during the Variscan Orogeny and, during the Tertiary this led to the exposure of the sandstone in upland areas (e.g. Slieve Aughty) and erosion of the oldest sequences of limestone across the Gort-Kinvarra lowland areas. The Namurian shales were eroded progressively from east to west, resulting in a shorter period of erosion and karst development in the Burren uplands than in the lowlands (Simms, 2003).

The groundwater flow in the lowlands tends to be focused in larger and deeper conduits than those in the uplands (Simms, 2005). Surface drainage features are largely absent in the lowland portion of the study areas (Figure 3.1). Rivers draining Slieve Aughty sink underground on reaching lowlands limestone and rainfall percolates through fissures and joints in the limestone so creating caverns, often associated with surface depressions, and conduits through which water flows until it rises to the surface, often at large springs (GSI, 2015). These conduits are thought to be >5m to 25 m in diameter (Drew and Daly, 1993; GSI, 2004). The hydraulic gradient of the groundwater flow and the system’s storativity are so low that winter rainfall often results in flooding and the formation of temporary lakes (turloughs) before transmission of water to the sea (OPW, 1998), expressed by the high groundwater dispersivity ratios recorded locally (Perquito et al., 2014). Comparison of potential runoff and gauged discharges in surface rivers across the region (Cave and Henry, 2011) indicate that surface rivers discharge >25% of potential runoff, with intertidal and submarine discharge expected to account for the remaining runoff. SGD is therefore expected to provide a substantial contribution to discharge into southern Galway Bay (Cave and Henry, 2011; McCormack et al., 2014; Schubert et al., 2015).

The Bell Harbour catchment (Figure 3.1a) lies within a pre-glacial valley surrounded by low hills (~ 200-300m) but open to the sea to the north (Simms, 2005). The limestone bedrock is exposed over most of the catchment and consists of gently dipping (~3º south), pure bedded and massive limestone (Pracht, 2004). It is classified as a regionally important
aquifer dominated by regional scale conduit flow rather than diffuse flow (DELG/EPA/GSI, 1999). Soil cover is limited to the valley floor and is generally 3m to 10m thick (GSI and RBD Consultants, 2004a). There are three main groundwater-fed turloughs (Luirk, Gortboyheen and Turloughnagullaun) and a permanent, tidally influenced, brackish lake (Muckinish Lough, Figure 3.2a). Bathymetric lidar data for the coastal inlet in the north of the catchment indicate a shallow, flat-bottomed bay, typically -1.5m to -2.5m OD (Infomar, 2006; Figure 3.2a). Seven circular seabed depressions within the bays can be observed from aerial photographs and the lidar data, and are inferred to be expressions of submarine sinkholes or springs. Substrates are predominantly mud to muddy sand in the inner part of Bell Harbour, with some shelly sand in places (O’Toole, 1990). A NNE-SSW trending fault, MacDermot’s Fault, and minor parallel faulting, with <200m sinistral displacement (GSI, 2005), is exposed in the south of the catchment (Figure 3.1a). Projection of the fault northwards transects Muckinish Lough and six of the observed sinkholes or springs in the bay. Regional investigations of SGD from thermal infrared imagery (Wilson and Rocha, 2012) suggest the potential connection between onshore faulting and SGD.

Groundwater flow within the catchment is via well-developed near-surface (upper 5m to 10m) epikarst and through linked pathways of fissures, joints and large conduits which allow rapid vertical flow, probably ~ 20-150 m hr⁻¹ as in neighbouring catchments (Cronin et al., 1999; Perriquet et al., 2014). The region is characterised by a strong tidal range (Cave and Henry, 2011). A hydrogeological study (Perriquet et al., 2014) incorporating well and surface water logging indicated saltwater influence inland up to ~ 1 km from the shoreline into the lowest lying part of the catchment. Evidence of saltwater intrusion was observed in the vicinity of Muckinish Lough, located along the projection of MacDermot’s Fault. Specific conductivities (SpC) for this lake showed large variations linked to tidal movements, indicating high hydraulic connectivity between the lake and the bay, probably due to the development of conduits or weathered zones parallel to the fault core (Bense et al., 2013). Large hydraulic diffusivity ratios were also measured further inland (~ 3km) in the region of the fault.
Figure 3.2: The bathymetry of (a) the Bell Harbour and (b) Kinvarra inlets from lidar data (Infomar, 2006) coupled with digital echo sounder data, with locations of main intertidal springs, submarine springs/sinkholes and marine geophysical profiles described in the text. The subterranean and submarine extension trend of MacDermot’s Fault and a possible NNW trending fault in Bell Harbour (see Figure 3.4) are inferred from this study. Locations of marine ERT profiles (e.g. M4) are shown. Map coordinates (metres) are Irish National Grid.

To the east of the Bell Harbour catchment, groundwater flow in the Gort-Kinvarra lowlands is via epikarst from ~1-10m below ground level; there are solutionally enlarged conduits and cave systems, extending to ~ 30m below the epikarst, and small fractures and joints which are linked to the main conduit systems (GSI, 2004). There is also a suggestion of flow mechanisms ~ 70 to 80 m below ground level (OPW 1998; Drew and Daly, 1993) that may be associated with faulting or dolomitisation (GSI, 2004). At Loughcurra South, approximately 2 km southeast of Kinvarra (Figure 3.1b), a very high productivity (>65-70m$^3$/hour) local authority borehole is finished in weathered rock at greater than 60m depth. Regional bedrock density estimates (Armstrong, 1997) across the lowlands indicate lower bedrock
densities for the limestone in this area and may be indicative of increased porosity from extensive dissolution.

Petrunic et al. (2012) confirm a tidal seawater influence on the aquifer geochemistry ~ 2 km south of Kinvarra Bay at Loughcurra spring and Gill et al. (2013) observe a tidal influence on Caherglassaun Lough, 5.2 km southeast of Kinvarra Bay. Primary permeability is negligible in these lowland limestones (GSI, 2004) and this is similar in the younger limestone of the Bell Harbour catchment (Perriquet et al., 2014). Models constructed to define the hydraulic system for the region (OPW 1998; Drew 2003; Boycott and Bruce, 2003; Cave and Henry, 2011; Gill et al., 2013; McCormack et al., 2014) suggest a pipe or conduit flow path connecting the groundwater-fed Caherglassaun Lough to multiple intertidal springs at Dunguaire Castle, Kinvarra Harbour and the Archway within Kinvarra Bay (Figure 3.1b). No discharge points have been mapped offshore though a ‘hot spot’ of a $^{222}$Rn tracer off Tarrea Pier suggests a discrete discharge pathway (Schubert et al., 2015).

### 3.4 Methodology

A quantification of the different mechanisms contributing to SGD requires an understanding of sub-surface three-dimensional (3-D) structures using geophysics. The principal technique used in this study is electrical resistivity tomography (ERT). These data have been acquired along profiles in static terrestrial and (surface-towed) marine modes (3.3.1b and 3.2). The terrestrial ERT data were acquired using 48 electrodes at 3 m, 5 m and 10 m separations connected via multi-core cables to a resistivity meter (IRIS, 2009). The electrode array configurations included Dipole-Dipole (DD) and Wenner-Schlumberger (WS) arrays (Loke et al., 2013), and allowed depths of investigation of > 50 m below ground level. The data quality was improved by stacking $2 \leq n \leq 4$ measurements (assuming a quality factor of 3% using the same electrodes) to reduce random noise. The marine ERT profiles were recorded with a boat-mounted resistivity meter connected to a surface-towed multi-core cable incorporating 13 graphite electrodes at 5 m or 10 m separations. The meter was coupled to an echo sounder, providing a continuous record of the water depth, and a GPS system with a positional accuracy of ± 3m. A temperature/salinity meter provided water electrical...
conductivity values for inclusion in subsequent data modelling. The continual motion of the towed electrode cable does not allow stacking to reduce random noise as cable offsets due to wind and waves, and electrodes snagging vegetation vary with position (Day-Lewis et al., 2006). To assess noise levels (Rucker et al. (2011), resistances for all levels (n = 1, 2...10) were convolved with a low-pass filter (Fraser (1969) with a length of 25m equivalent to an average of 9 survey intervals.

The depth of investigation and horizontal resolution of sub-seafloor structure from surface-towed ERT is dependent on factors including water column thickness (Day-Lewis et al., 2006; Loke and Lane (2004)), salinity of the water column and saturated sediment (Befus et al., 2014), array configuration (Day-Lewis et al., 2006; Henderson et al., 2009; Manheim et al., 2004; Mansoor and Slater, 2007) and towing speed. Based on this literature and trials at the survey sites, the Wenner-Schlumberger electrode array configuration was deployed at Bell Harbour and a modified Wenner configuration, after Mansoor and Slater (2007), was deployed in Kinvarra Bay. Two marine surveys were carried out at Bell Harbour during medium to low tides and at varying survey speeds, typically ~ 3km hr⁻¹, depending on weather conditions and tidal flow influences. The water depth was < 3m, but increased to 11m over sinkholes; its resistivity was typically 0.28 Ωm. Using a 5m electrode separation and WS array, the depth of investigation below the seafloor was ~ 7m to 10m dependent on water depth and resistivity (Loke et al., 2013). In Kinvarra Bay the marine survey incorporated a seismic reflection sub-bottom profiler, requiring a larger boat and increased survey speeds, on average ~7 km hr⁻¹. The joint acquisition of marine ERT and seismic reflection data also limited acquisition in shallow water because of the size constraints of the vessel. Water depth was 2 - 7 m; its resistivity was typically 0.28 Ωm. Using a 10m electrode separation, the depth of investigation below the seafloor was ~ 15m to 18m. A sub-set of a total of 35km of ERT profiles is shown in Figures 2 and 6.

All ERT data have been inverted using a finite-element forward model (Coggon, 1971; Holcombe and Jiracek, 1984) using a robust L₁-norm least squares inversion algorithm (Claerbout and Muir, 1973) to produce sections of subsurface resistivity values. The least-squares approach minimised the
absolute difference between the observed and calculated apparent resistivities typically within ~ 7 iterations. As the frequency of the data acquisition guaranteed a distance between consecutive data points < the electrode separation for any survey speed, the forward model cell widths were chosen to be ~ half the electrode separation. The model layer thicknesses were continually increased by ~ 5% with depth, starting with a first layer thickness of 0.75m, for those layers that did not exceed the maximum pseudo-depth (Loke, 2010, 2012). The static terrestrial profile inversions employed standard model cell numbers equivalent to the number of data points (Geotomo Software, 2010). They achieved absolute errors between theoretical responses and data of ~ 1.2% to 2.7% for DD and WS configurations. Collinear arrays in mixed DD/WS configurations were merged to increase the number of data points and, therefore, resolution of the inverted resistivities (Dahlin and Zhou, 2004) while removing noisy artefacts (Kaufmann et al., 2012a).

The water depth was determined using a digital echo sounder. The sub-bottom seismic reflection profiling was acquired with a side-mounted digital profiling system (GeoAcoustics, 2009) using a swept frequency ‘chirp’ waveform in the frequency range 1.5 and 11.5 kHz on a 16 or 32 ms pulse length, allowing a vertical resolution of ~ 0.06m between layers of different acoustic impedance.

3.5 Results
Terrestrial profiles
The influences of fault and conduit flow and tidal driven seawater ingress are examined by ERT profiles T1 to T4 (Figure 3.1b). A mixed array inversion for ERT profile T1 (Figures 3.1b and 3.3a) shows soil ≤ 5 m thick to the west, assuming soil resistivities of 80 to 320 Ωm, and absent to the east above outcropping rock. Limestone resistivities vary from 320 to 5000 Ωm, typical of Irish settings (O'Connor, 1998; O'Rourke and O'Connor, 2009) and are similar on either side of the fault. Fault zones in carbonate rocks typically comprise a fracture-dominated zone flanking a fault core (Bense et al., 2013; Evans et al., 1997) resulting in a region parallel to the fault core with potentially increased porosities. As the bulk resistivity of a granular medium is a function of the porosity, pore fluid resistivity and clay
content (Archie 1942; Waxman and Smits, 1968), reduced resistivities suggest faults and/or karst zones. Inversions of a theoretical model (Figure 3.3d) that simulate the field acquisition procedure for the T1 DD array (Figures 3.3b and 3.3c), with added errors to account for random noise equivalent to 5%, approximate the observed DD (Figures 3.3e and 3.3f) and WS configurations (not shown). They suggest a 12 m wide, vertical low resistivity (~300 Ωm) fault zone for Zone A (Figure 3.2a). Zone B to the west may be a parallel fault and/or karst zone and to the east Zone C may be a karst zone or an edge effect of the inversion.

![Figure 3.3: Modelled resistivities for T1 at a ground level of ~ 40 m OD. (a) Mixed WS and DD array inverse model with topography and absolute error of ~ 1.1%. Zone A is indicative of MacDermot's Fault, B may be a parallel fault and/or karst zone and C may be a karst zone or an edge effect of the inversion, (b) measured apparent resistivities and (c) inverse model resistivities for T1 DD array (absolute error 1.5%) for comparison with the forward model, (d) incorporating a thin soil layer, horizontal lithological layering and vertical low resistivity zones to represent fault and karst zones and the resultant apparent resistivity pseudosection (e) and inverse model resistivities (f) (absolute error 3.7%).](image-url)
The Waxman-Smits (W-S) model (Waxman and Smits 1968) extends the classical Archie’s Law by assuming that the bulk resistivity of a rock is a function of the resistivity of pore fluid and small clay particles dispersed in the pore space between coarser solid grains: 

\[ \rho_b = (\sigma_f + \sigma_s)^{-1} \varphi^{-m}, \]

where \( \sigma_f \) is the pore fluid electrical conductivity, \( \varphi \) is the porosity, \( m \) is the cementation factor, and \( \sigma_s \) represents the electrical conductivity due to hydrated clay mineral exchange cations at different salinities in the pore volume. A pore fluid resistivity of \( \rho_f = 14.8 \, \Omega \, \text{m} \), was assumed in this study, which is the mean value recorded at a well within 200 m of the fault (Perriquet et al., 2014). The exponent \( m \) is related to the connectivity of the pore space containing the fluid with a lower bound for full connectivity of \( m = 1 \) and rising with increasing compaction and cementation. An upper bound of \( m = 2 \) is assumed for the competent rock beyond the fault zone, where a bulk resistivity \( \rho_b \approx 2500 \, \Omega \, \text{m} \) in the theoretical model (Figure 3.3c) implies a porosity of \( \sim 0.1 \) in the upper 20 m of a clay-free limestone. This is consistent with previous estimates for carbonate aquifers (Fetter, 2001; Worthington (1999)) and inferred from the gravity field (Armstrong, 1997).

In the fault zone, where \( \rho_b \approx 300 \, \Omega \, \text{m} \), a porosity of \( \sim 0.22 \) is implied if there is no clay present. As the fault zone is more important as a pathway for fluid flow than intergranular flow in the limestone matrix, its porosity must be \( > 0.1 \) as permeability generally increases with increasing porosity, e.g., following a Kozeny–Carman relationship (Mavko et al., 1998). A porosity \( > 0.1 \) implies a lower bound on the exponent of \( m \approx 1.3 \) to explain the modelled resistivities. The presence of conductive clay minerals will contribute to a reduction of resistivity and decrease the porosity required to explain the low resistivities in the fault gouge (Wildenschild et al., 2000).

The nearshore ERT profile T2 (Figures 3.2a and 3.4a) indicates soil thickness \(<1 \text{m} \) to 7 m and two vertical low bedrock resistivity zones (D and E). Above sea level, resistivities for D and E are typically 160 to 320 \( \Omega \, \text{m} \) reducing to \( \leq 30 \, \Omega \, \text{m} \) below sea level, indicating infiltration of seawater. Zones D and E therefore appear to be sub-surface faults linked to broader conductive zones at depths \( > 10 \, \text{m} \) below MSL associated with a saltwater wedge in the area (Perriquet et al., 2014). Time-lapsed surveying of Zone E over part of a tidal cycle during a prolonged period of low rainfall (T3,
Figure 3.1b, Figures 3.4b and 3.4c) show a low resistivity (~50 \(\Omega m\)) subsurface zone that became flooded with saltwater as tidal levels increased, reducing resistivities to < 10 \(\Omega m\) which requires an 80% reduction in pore water resistivity over the tidal cycle; this indicates significant saltwater flow in the fault zone.

![Diagram showing resistivities over time](image)

Figure 3.4: (a) Modelled resistivities for DD configuration at T2 at ground level of ~ 8 m OD with possible fault zones D and E. Below the high water mark (white dashed line) the resistivities drop to 20 Ohm-m suggesting brackish water in the fault zone. Modelled resistivities for T3 recorded with DD configuration across zone E at (b) low tide and (c) higher tide show a decrease in resistivities from ~ 60 \(\Omega m\) to 10 \(\Omega m\). (d) Percentage decrease between (b) and (c) highlighting other flow-paths in addition to Zone E.
Water levels in Muckinish Lough, 100m south of T2 and T3, show a tidal influence and display a 95% variation between minimum and maximum water resistivity of 0.35 and 10.75 Ωm (Perriquet et al., 2014). Zone E, and possibly D, is therefore considered to be an extension of faulting through Muckinish Lough facilitating seawater movement into the lake. By comparison, the electrical resistivity variations measured over an eighteen month period at Lough Luirk - to the south-west (Figure 3.1) - ranged from 12 Ωm to 23 Ωm, suggesting little seawater influence (Perriquet et al., 2014). The observed decrease in resistivity (Figure 3.4d) significantly increases the resolution of the subsurface structures and suggests multiple flow-paths in addition to Zone E.

Figure 3.5: Resistivity of major conduits between Caherglassaun Lough and Kinvarra with (a) & (b) observed and calculated apparent resistivity pseudo-sections for DD array T4 and (c) the resultant inverse model resistivities, (d) a synthetic model containing large diameter, freshwater (20 Ωm) conduits, and variable rock resistivities suggesting karstification and/or increased clay/water content (e) the resultant simulated apparent resistivity pseudo-section and (f) inverse model resistivities.

The suggestion of large diameter conduit flow, from tracer testing and modelling (Gill et al., 2013; McCormack et al., 2014; OPW, 1998), between
Caherglassaun Lough and Kinvarra via a conduit or conduits at elevations of 0 to -10 (m OD) (Drew and Daly, 1993) is confirmed by T4 (Figure 3.1b, Figure 3.5a to 3.5c). A forward model (Figure 3.5d and 3.5f) suggests the presence of two large diameter (~10 to 25 m), water filled conduits (assumed = ~20 Ωm water resistivity similar to inland values from (Perriquet et al., 2014)).

**Marine profiles**

Submarine structural geology along with diffuse and discrete SGD are examined by marine ERT profiles (Figure 3.2). The inversions of Wenner-Schlumberger marine resistivity sections from the Bell Harbour inlet (Figure 3.6a to 3.6e) achieved absolute error values of ~ 1.2 - 5.6% after 7 iterations. The inverse modelled resistivities increase from ~0.1-1.0 Ωm with depth, characteristic of unconsolidated seawater-saturated sediment (Befus et al., 2014). The limestone bedrock resistivity of ~500 Ωm is lower than its terrestrial equivalent, possibly due to damping of large resistivity contrasts toward the mean in the inversion process and poor sensitivity due to current focussing in the saline water column (Day-Lewis et al. (2006) in addition to saltwater saturation.

Sediment resistivities of ~0.15 Ωm (M1, Figure 3.6a) increasing to ~0.45 Ωm in thicker sequences in Pooldoody and Poulnaclogh Bays, (M4, Figure 3.6d) may reflect a change from mud in Bell Harbour to muddy sand in Pooldoody and Poulnaclogh Bays (O'Toole, 1990). Generally, the sediments are < 3m thick in the south and along edges of the inlet, and up to 8m thick in Pooldoody and Poulnaclogh Bays. Water resistivities ($\rho_w$) can be determined as a function of the total dissolved solids (TDS) through the relationship $TDS = A/\rho_w$ (Freeze and Cherry, 1979), where the conversion factor $A$ is typically between 0.5 and 0.8, depending on the ionic composition of the water and normalised at a temperature of 25°C (Table 5). In the sediments near intertidal springs BH1 and BH2 (Figure 3.2a), a change in the bulk resistivity from ~ 0.15 Ωm to 0.75 Ωm (Figure 3.6a) requires an increase in pore water resistivity to 1.3 Ωm (assuming seawater resistivity of 0.27 Ωm), indicative of brackish water. At the Bell Harbour Quay (BHQ) intertidal spring, water resistivity of ~1.0 Ωm at low
tide decreased to ~ 0.33 Ωm at high tide (Perriquet, 2014). Adjacent to this spring, sediment resistivities up to ~ 4 Ωm were recorded at low tide (Figure 3.6b) reflecting increased freshwater input from the adjacent spring.

<table>
<thead>
<tr>
<th>Type of water</th>
<th>TDS (g/l)</th>
<th>( \rho_w ) (Ωm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fresh</td>
<td>0-1</td>
<td>&gt; 7</td>
</tr>
<tr>
<td>Brackish</td>
<td>1-10</td>
<td>0.7 – 7</td>
</tr>
<tr>
<td>Seawater</td>
<td>10-100</td>
<td>0.07 - 0.7</td>
</tr>
</tbody>
</table>

Table 5: Typical TDS values and corresponding water resistivities based on a conversion factor of 0.7.

Vertical low resistivity zones in the bedrock (dashed black line on M3 and M4, Figure 3.6c and 3.6d) are interpreted as the northward extension of MacDermot's Fault towards the submarine springs. Similar low resistivity zones (dashed white line on M4 and M5, Figure 3.6d and 3.6e) also cut through a submarine sinkhole and appear to indicate a subparallel fault that may extend to the terrestrial Zone D on T2 (Figure 3.2a, 3.4a). Increased sediment resistivities above this fault zone (feature 1, Figure 3.2a, 3.6e) suggest possible groundwater discharge. Other sinkholes/seabed depressions have associated low bedrock resistivities e.g. the sinkhole south of Oyster Spring (M1, Figure 3.6a), and a localised karst zone beneath a seabed depression (feature 2, Figure 3.2a, 3.6c) with notable increased sediment resistivities.
Figure 3.6: Marine ERT profiles; estimated base of unconsolidated sediments (≤ 1 Ωm) indicated by black dotted line (grey where the base is unclear). MacDermot’s Fault extension indicated by black dashed line with subparallel faulting indicated by white dashed line. Depth in metres below water level. Features 1 - 5 discussed in text.
To the east in Kinvarra Bay, the inversions of modified-Wenner marine resistivity sections (Figure 3.6f and 3.6g) achieved absolute error values after 7 iterations of 1.8% and 9%. Deeper water necessitated a 10 m electrode separation to investigate 15 to 20m below the seabed. A vertical low bedrock resistivity zone on M6 (feature 3, Figure 3.2b, 3.6f) recorded ~70m northwest of Tarrea Pier has been interpreted as a karst zone that is likely the SGD source for the $^{222}$Rn tracer 'hot spot' (Schubert et al. (2015). The modelled sediment resistivities are variable suggesting a change in sediment type and/or a groundwater influence. A similar low bedrock resistivity zone (feature 4, Figure 3.2b, 3.6f) is located ~ 600m northwest of the pier. The deepest part of the bay is underlain by a low bedrock resistivity zone (feature 5, Figure 3.2b, 3.6g) which may also be an active groundwater discharge point. Notably, active shellfish farms have been positioned above this zone. Other vertical low resistivity potential karst zone discharge points without associated resistivity increases in the overlying sediments were observed in the larger data set (not shown), and may act as pathways during periods of increased rainfall as suggested by Cave and Henry (2011).

Examination of inverse model resistivity relative sensitivities in both Bell harbour and Kinvara bay, assuming a cut-off limit of 0.1 to 0.2 for the effective depth of investigation of the data as described by Oldenburg and Li (1999), indicate reliable results for the interpreted fault and karst zones below the conductive seawater and unconsolidated sediment layers. A comparison of inverted marine ERT resistivity sections with co-located sub-bottom seismic sections (Figure 3.7) suggests that much of the sediment layering did not have a corresponding electrical contrast, apart from the sediment-rock interface. Blurred sediment/rock interfaces, often with gaps due to poor reflectivity e.g., at features 3 (Figure 3.7a) and 5 (Figure 3.7c), are typical of eroded karst limestone surfaces (Evans and Lizarralde, 2003) while undulating bedrock topography is typical of karstified limestone. Bedrock depressions suggestive of sinkholes underlie a potential SGD point at feature 4 (Figure 3.7b). Enhanced reflectors and localised plumes (Figure 3.7c) at feature 5 are suggestive of potential groundwater seepage (Hovland and Judd, 1988; Taylor, 1992).
3.6 Discussion and Conclusions

Groundwater discharge pathways in two adjacent catchments located in a coastal karst aquifer on the southern coast of Galway Bay in the west of Ireland have been examined with terrestrial and surface-towed marine electrical resistivity tomography. These geophysical techniques have provided evidence for both structural and dissolutional influences on groundwater pathway development, complementing ancillary geological and hydrogeological evidence for fault controlled groundwater pathways in one catchment and large conduit flow in the other. Towed marine resistivity data allowed the examination of SGD at intertidal and offshore locations and confirmed the offshore extension of faulting and its influence on groundwater movement. Jointly acquired sub-bottom profiling allowed comparison of sediment and rock interfaces with inverted resistivity boundaries highlighting the undulating karst bedrock topography in suspected SGD zones.
In the Bell Harbour catchment, a northward subterranean and submarine extension of MacDermot’s Fault and additional subparallel faulting have been identified. A time-lapsed ERT survey over part of a tidal cycle across the subterranean extension of MacDermot’s Fault demonstrated tidally-influenced saltwater intrusion into the fault through the low lying part of the catchment. The time-lapsed survey significantly enhanced the spatial resolution of the sub-surface structures responsible for the seawater intrusion. Combined with the marine ERT, the geophysical investigation provided evidence that the sub-surface structural controls on groundwater movement appear to be limited to the west of the inlet while discharge along the east of the catchment appears to be dominated by epikarstic intertidal springs.

In the Gort-Kinvarra catchment, the geophysical investigation confirmed large diameter conduit flow from a central, subterranean-fed lake to intertidal springs in the north. Though intertidal discharge is known to dominate the Gort-Kinvarra catchment (Boycott and Bruce, 2003; McCormack et al., 2014), marine ERT provided confirmation of discrete SGD off Tarrea Pier in the east of the inlet and other potentially ephemeral discharge points within Kinvarra Bay. Offshore submarine springs are generally expressions of karstified flow paths that developed in periods of reduced sea level (Ford and Williams, 2007) such as the mid-Holocene (~6000 BP) where a sea level of -5m (+/- 1m) is considered to have persisted for ~ 2000 years for Galway Bay (Lambeck and Purcell (2001); Williams and Doyle (2014)) . White (2002) suggests that dissolution is a rapid process requiring only thousands of years for the development of significant groundwater flow paths. The -5 mOD contour in Kinvarra Bay (Figure 3.2b) places the potential offshore SGD points identified by the geophysical survey within the intertidal zone, implying that these were the main intertidal springs prior to the Holocene marine transgression.

The characterisation of SGD and groundwater pathways in coastal karst aquifers is critical for understanding climate-change and anthropogenic impacts such as sea level rise with its potential to increase the volume of saltwater entering coastal aquifers, and extreme flood events with their
potential to contaminate groundwater from rapidly percolating surface pollution (Kløve et al., 2014; Sweeney et al., 2008). The pre-requisite for the modelling strategy required to forecast the impacts on groundwater is an assessment of the contribution to groundwater fluxes from, *inter alia*, diffuse sources and discrete pathways. The geophysical techniques deployed for this study have imaged the subsurface structures that control freshwater-seawater interactions and have shown the importance of water movement in discrete zones. The deployment of these methods enhances the quality of the models used to assess the hydraulics of karst aquifers and can be used in coastal zone management planning.
Chapter 4: Characterisation of karst hydrogeology in western Ireland using geophysical and hydraulic modelling techniques

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Study region:
Bell Harbour. A sub-catchment of karst landscape, The Burren, in Western Ireland.

4.1 Abstract
Bell Harbour is difficult to investigate using traditional hydrogeological techniques due to its complex mixture of upland, lowland and coastal karst, with ephemeral lakes and submarine/intertidal discharges. This study uses electrical resistivity tomography and discrete conduit network modelling to characterise the hydrogeology of the catchment by determining flow pathways and their likely hydraulic mechanisms. Results suggest two primary pathways of northwards groundwater flow in the catchment, a fault which discharges offshore, and a ~2 m diameter karst conduit running underneath the catchment lowlands against the prevailing geological dip. This conduit, whose existence was suspected but never confirmed, links a large ephemeral lake to the coast where it discharges intertidally. Hydraulic modelling indicates that the conduit network is a complex mixture of constrictions and bypass channels with multiple inlets and outlets. Two ephemeral lakes are shown to be hydraulically discontinuous, either drained separately or linked by a low pressure channel.
4.2 Introduction

Methods of hydrogeological investigation for karst catchments have long been established (Goldscheider and Drew, 2007). Traditional techniques include tracers, spring hydrograph analysis and hydrochemical sampling. More recently, techniques such as numerical modelling and geophysics have grown in importance and capability. Each method has its own benefits and drawbacks and can be more or less applicable to particular type of catchments. The Bell Harbour catchment in western Ireland has thus far proven difficult to investigate using traditional techniques due to its complex mixture of upland, lowland and coastal karst, with the added complexity of submarine/intertidal discharges.

The Bell Harbour catchment is located in the northern part of the Burren in Western Ireland. The karst landscape of the Burren is relatively well-understood hydrogeologically (Drew, 2003) but the Bell Harbour sub-catchment remains somewhat enigmatic. As the coastal outlet springs for the catchment are situated at an intertidal or wholly submerged elevation, their use for sampling or gauging is impractical. As a result, tracer studies which are normally a useful technique in karst regions (e.g. Baedke and Krothe (2000)), have thus far proven inconclusive. Other techniques such as spring hydrograph analysis (Fiorillo, 2009) or hydrochemical sampling (Moore et al., 2009) are equally impractical in this catchment. As such, this region requires a more physical approach in order to characterise the hydrogeology of the catchment.

While the upland karst remains difficult to investigate (although it is well understood due to its characteristic geology), the lowland karst (mostly <30 m above sea level), which forms the locus of the catchment, allows for use of shallow surface geophysical investigation techniques. In tandem with these geophysical methods, the presence of ephemeral lakes (turloughs) and conduits within the catchment allows for the application of hydraulic modelling using a discrete conduit network.

Electrical resistivity tomography (ERT) has been widely used for hydrogeological applications (e.g., Khalil (2006), Nyquist et al. (2008), Nguyen et al. (2009), Comte et al. (2012), Martorana et al. (2014), Amidu
and Dunbar (2008)). This non-invasive technique allows the operator to ‘see’ into the earth (Bechtel et al., 2007) by determining lateral and vertical variations in subsurface resistivity (or its inverse, conductivity) indicative of the underlying geological features. The technique is particularly sensitive to the presence of sub-surface clay deposits and water, especially saline water. Its use in karst regions around the world is well documented (Ismail and Anderson, 2012; Kaufmann et al., 2012b; Satitpittakul et al., 2013). It has also been widely used to detect zones of karstification and conduits in the examination of groundwater movement through highly heterogeneous karst regions (e.g., Zhu et al. (2011), Meyerhoff et al. (2012)).

Discrete conduit network modelling is well-suited for conduit-driven karst catchments and has been used successfully in a number of studies (Chen and Goldscheider, 2014; Peterson and Wicks, 2006), including the nearby Gort Lowlands Catchment (Gill et al., 2013). The technique is typically used to provide an output (e.g. discharges, flood levels) based on previous investigative work to build the model. In this study, however, the modelling process is used as an investigative technique to better characterise the hydraulic linkages between surface water features. The method is particularly useful in a coastal catchment such as Bell Harbour as it does not technically require discharge information from the intertidal/submarine springs for calibration. The system can instead be calibrated using volume measurements from upstream surface water features. An accurate conduit network model could thus provide a detailed estimate of discharge from unobservable springs (McCormack et al., 2014).

Thus, the objective of this study was to use investigative techniques which take advantage of the catchment’s particular karst characteristics. ERT was used to validate the initial conceptual model of the catchments in areas where shallow karst or fractures were expected. Following this, conduit network modelling was used to further investigate the system, particularly the two ephemeral lakes and the conduit linkages between them and the sea. The study marks the first known combined application of these techniques in a karst region.
4.3 Area Description

4.3.1 Location and Climate Conditions

The Burren plateau is an upland limestone landscape in western Ireland of approx. 360 km$^2$ in area (McNamara and Hennessy, 2009). It consists of a broad plateau which rises up to 300 m and is one of only two upland limestone landscapes in Ireland as 90% of limestone areas in Ireland are below 150 m, (Drew, 2008). The climate is oceanic, with approximately 1500 mm annual precipitation and 980 mm effective rainfall (Drew, 1990). Rainfall is common throughout the year, although it tends to be drier in spring and early summer.

![Geology and hydrogeology of the Burren region, western Ireland.](image)

The Bell Harbour sub-catchment is approx. 56 km$^2$ and is located in the north eastern corner of the Burren (see Figure 4.1). It consists of a lowland valley surrounded on three sides by hillsides of exposed karstified limestone. The eastern and western extents of the catchment are relatively easy to estimate and have been delineated using previous tracer studies. The southern catchment boundary is more poorly constrained.
4.3.2 Geology

The Burren plateau forms a large and gently inclined (dipping 2-3 degrees to the south) limestone plateau and is dominated by a pure-beded carboniferous limestone of several hundred meters in thickness (Figure 4.1). It is bordered to the west and north by the Atlantic Ocean and Galway Bay respectively. To the east is the low-lying limestone plain of the Gort Lowlands and to the south lie Namurian sandstones and shales. Stratigraphically, the Burren is comprised of a thick succession of relatively pure Viséan limestones bounded by two thick clastic sequences above and below (Figure 4.2). The region is underlain by Devonian Old Red Sandstones. This is unconformably overlain by approx. 400 m of impure limestones (Tubber and Ballysteen Formations), which are overlain by the Burren and Slievenaglasha Formations consisting of pale grey and thickly to massively bedded limestones with occasional cherty intervals and clay horizons. These clay horizons (known as wayboards) are highly influential and have given the upper Burren plateau its characteristic terraced appearance. They are considered to represent fossil soils (palaeosols) that developed on paleokarst surfaces in periods of sea level regression during the deposition of the formation (Pracht, 2004). In the Aillwee Formation, 12 such horizons are known to exist, occurring at intervals of approximately 10-20m with typical thicknesses of 30-50 cm. Finally, the Burren is partially capped by Namurian shales and sandstones (Gull Island and Clare Shale).

These relatively resistant Namurian siliciclastics were stripped away from northeast to southwest across the region (Simms, 2003) allowing for the unroofing of the limestones of the Gort Lowlands while protecting the limestones of the Burren plateau. They have likely only been stripped away from the north Burren within the last million years. Bell Harbour Valley (and Ballyvaughan Valley to the west) predate this stripping and were likely to have developed during the Tertiary by rivers draining to the north, downcutting the shale cover, creating valleys and exposing the limestone underneath. The exposure of the limestone would then have rapidly increased the enlarging of the valleys due to dissolitional process (Simms, 2003).
Figure 4.2: Bell Harbour conceptual cross section displaying stratigraphy, groundwater table and likely groundwater flow paths.

The Burren limestones are relatively undeformed and only two major faults are mapped in the area, one of which is in the western part of the Bell Harbour catchment (Figure 4.3). This fault, known as MacDermot’s Fault, runs approximately north–south and shows a slight (<200 m) sinistral displacement of the Burren and Slievenaglasha Formations (Pracht, 2004). Subvertical features such as mineralised veins and joints are very common in the Burren and play a significant role in the development of underground cave systems (Gillespie et al., 2001).

The Burren is highly karstified with weathered limestone pavement occurring over 20% of its area and a limestone-rendzina combination occurring over an additional 30% (Plunkett Dillon, 1985). This exposed limestone surface was particularly susceptible to dissolution and weathering, resulting in a well-developed shallow epikarst layer of approximately 5-10 m in thickness. As with many coastal karst regions, a multi-level karst network is known to exist, owing to the changes in sea level during the previous ice age (Brooks et al., 2008). Indeed a borehole drilled by the Geological Survey of Ireland (GSI) in early 2015 found karstified features at depth including a highly karstified zone at 90-110 m.
below sea level which corresponds to the sea level during the most recent ice age (Edwards and Brooks, 2008)

![Map of Bell Harbour catchment](image)

**Legend**
- Turboughs
- Faults
- Bell Harbour Subcatchment Boundary
- ERT Transects (T1, T2 etc)
- Intertidal Springs
- Submarine Springs
- Depth Logger Locations
- Raingauge
- ERT Low Resistivity Zones

**Figure 4.3:** Map of Bell Harbour catchment showing MacDermot’s Fault, turloughs, springs, raingauge location ERT transect lines and low resistivity zones.

### 4.3.3 Hydrogeology

Diffuse, autogenic recharge dominates the hydrological regime of the Burren. The exposed limestone offers rapid recharge into the groundwater network. Once in the saturated zone, the predominant path of water
through the Burren is via a well-developed karst conduit network. Tracer studies (Drew, 2003) have revealed flow paths and have allowed for the delineation of much of the Burren (Figure 4.1). However, the Bell Harbour catchment has not yet been successfully traced. This is largely due to the difficulty in recovering tracer from submarine springs. Nevertheless, due to the considerable discharge from the springs, water is known to move northwards through Bell Harbour; but the boundary with the southern catchment (which drains towards the Fergus Springs, see Figure 4.1) is quite uncertain due to the topography, the southwards dip of the limestone and the presence of impermeable clay horizons. These layers obstruct vertical flow and can transmit water significant distances southwards through the unsaturated zone.

In the Bell Harbour Lowlands (approx. 10-20 m above the saturated zone), the hydrodynamic environment has been shown to include matrix, fissure and conduit flow systems (Perriquet, 2014). During extended wet periods, these underground systems exceed their capacity and discharge into two ephemeral lakes (or turloughs), Luirk and Gortboyheen. A third lake within the catchment, Muckinish Lake is located near the coast. However, it is a tidally impacted and brackish lake that is not linked with the active conduit network. As is typical of turloughs, Gortboyheen and Luirk Lakes have no surface water inlets or outlets. Recharge and drainage of these lakes occurs entirely via groundwater through estavelles (or exposed limestone pavement at higher elevations). Drainage eventually emerges at a series of intertidal springs along the coastline, the largest of which is Pouldoody Spring. Within the Bay itself, a number of cavities are visible on bathymetric Lidar and Aerial photos. These springs appear to be associated with MacDermot’s Fault as they lie along a straight path extending from the known fault line (as does Muckinish Lake). Evidence has also been found (such as the 2015 GSI borehole) which indicates a deep paleokarst system beneath the primary active karst system. It is possible that this deep system discharges water some distance offshore.

4.4 Methodology

In this study, the hydrogeological behaviour of the catchment was elucidated by building on previous research (Perriquet et al., 2014; Petrunic...
et al., 2012; Smith and Cave, 2012) and applying a combination of ERT and hydraulic modelling. Specifically, the dynamic behaviour of the lakes was targeted as a means to understand the catchment as a whole.

### 4.4.1 Fieldwork and Data Collection

A variety of different field data were collected for the project in order to develop and calibrate the modelling approach.

**Rainfall**

High-resolution (15 minute) rainfall data were collected using a tipping bucket ARG100 rain gauge (Environmental Measurement Ltd) installed in the centre of the catchment (Figure 4.3) at an elevation of 32 m above sea level (ASL). Daily rainfall and evapotranspiration data were obtained from climatic weather stations run by the national meteorological service, Met Éireann.

**Turlough Water Level and Volume**

Turlough water level time series were collected at hourly resolution between June 2014 and June 2015 using Schlumberger Mini-Diver® DI501 and DI502 pressure transducers. Compensation for the variation in prevailing air pressure was made using a BaroDiver® (DI500) which was installed at ground level near the centre of the catchment. Depth-area relationships were required to convert turlough water levels into volume data. Thus Luirk and Gortboyheen turloughs as well as Muckinish Lake were surveyed during summer 2014 using a Trimble 4700 GPS system which provided accuracy of 0.01 m horizontally and vertically. Depth area relationships were then computed using Surfer 3D software and a Visual Basic script which computed the volume and planar area at given depth intervals.

**Offshore Spring Discharge**

Discharge from an offshore spring was monitored using an INW Aquistar® CT2X water electrical conductivity/temperature datalogger between December 2014 and July 2015. The logger was fixed in place using a
concrete platform and placed on the sea floor and the platform was connected to the surface via a rope and buoy to aid recovery (Figure 4.3).

4.4.2 Electrical Resistivity Tomography

Electrical resistivity tomography was used to image underground flow paths in the shallow and active karst network within the lowlands of Bell Harbour. The ERT work consisted of four transects covering suspected active karst zones. Two transects (T1, T2) were carried out over 2012 and 2013 field campaigns in conjunction with accompanying research (detailed in Chapter 3 which forms the paper currently in submission by O‘Connell et al. (2016)) while the other two transects (T3, T4) were carried out in January 2016. The purpose of T1 and T2 was to investigate the hydrogeological functioning of MacDermot’s Fault and its potential linkage with Muckinish Lake and the offshore springs. T3 and T4 were carried out to identify the suspected underground conduit (or conduits) linking Gortboyheen turlough with the coastal springs, which would thus validate the modelling approach.

The ERT profiles were acquired with a 10 channel IRIS Syscal Pro resistivity meter coupled to a 48 electrode multicore cable using a standard 2D approach (Dahlin and Zhou, 2004) employing the Dipole-Dipole (DD) array configuration to optimise the detection of near vertical karst features (Satitpittakul et al., 2013). Electrode separations of 5 m for T1, 3m for T2 and 10 m for T3 and T4 achieved depths of investigation of up to 50 m below ground level. The DD array employs combinations of 4 electrodes at varying separations along the profile to determine subsurface bulk resistivity. These are inverted with Res2dinv software (Geotomo Software, 2010) using a standard least squares inversion algorithm (Loke and Barker, 1996) to produce sections of subsurface resistivity values.

4.4.3 Conduit Network Modelling

Unlike the majority of the Burren which is characterised by active conduits and accessible cave systems, the degree of conduit flow in Bell Harbour is more difficult to ascertain. This is partially due to the low-lying nature of much of the catchment which results in flooded and inaccessible caves.
Perriquet (2014) however, observed conduit hydrodynamic behaviour in a number of boreholes using recession analysis. Furthermore, the presence of a deep, relatively quick-flooding turlough (Gortboyheen) also suggests a surcharge-tank and conduit system rather than an epikarst flow-through type system (Naughton et al., 2012). On the basis of the ERT transects which also indicated the presence of a conduit system, hydraulic modelling was used to interpret the hydrological functioning of the turloughs. This was carried out by testing a number of hypotheses of how the system is connected based on karst hydraulic principles.

The hydraulic model was built using Infoworks ICM version 6.5 (Innovyze software). This software package incorporates the Hydroworks modelling engine and is designed for management of urban and river drainage networks. The software simulates the hydraulic behaviour of a pipe network under varying conditions of rainfall, land use, inflows etc. As the model is capable of modelling the hydraulic conditions in both open channel and pressurised flow channels, it is highly suitable for modelling a well-developed karst conduit network, as has been proved in a neighbouring karst system (Gill et al., 2013).

The model represents the main conduit flow system in the Bell Harbour as a complex network of pipes (representing conduits) and tanks (representing turloughs), discharging at an outfall (coastal springs). Recharge is incorporated into the model using a conceptual epikarst fracture system represented by sub-catchments draining into the main conduit system. A realistic diffuse autogenic recharge signal was achieved using a combination of rainfall-runoff routing, Groundwater Infiltration Module (GIM) and SUDS (Sustainable Urban Drainage) applications in the Infoworks modelling suite. Rainfall in each sub-catchment is subjected to evapotranspiration and initial wetting losses. Following this, the recharge signal is transmitted through the soil through using the GIM and onward towards the conduit network via a series of permeable pipes which obey Darcy’s Law. Two varieties of sub-catchments were used in this model to distinguish between the differing recharge characteristics of the upland exposed limestone and the lowland soil-covered limestone. For a tank (turlough) to flood, the flow must be constricted in the downstream conduit.
This is facilitated using a ‘throttle’ pipe which can be altered in size to create the required pressure to flood the turlough. Additionally, bypass channels can be added around the turlough connection. These bypass connections enable further control on the flood pattern, particularly the recession curve (see Figure 4.4 for a conceptual diagram of the model).

The only means of calibrating the model was using the water levels in the turloughs (Gill et al., 2013). Water level data from Gortboyheen and Luirk turloughs between summer 2014 and summer 2015 was used for this purpose. The aim of the modelling exercise was to establish which conduit network combinations could best replicate the water levels observed in the turloughs over the 2014-2015 period. The hypothetical network combinations are discussed further in section 4.5.3.2.

Figure 4.4: Conceptual model displaying sub-catchments, permeable pipes, turlough, mainline karst conduit and throttle.

4.5 Results and Discussion

4.5.1 Turlough Water Levels

Water levels from the turlough loggers are presented in Figure 4.5. Gortboyheen turlough fills relatively rapidly to a considerable depth of 9 m (25.3 mASL) which is indicative of a surcharge tank system (Naughton et
During the 2014-2015 flood season the turlough stored over 1.51x10^6 m^3 of water. Luirk turlough however, only flooded to 2.16 m (3.56 mASL) and stored just 0.16 x10^6 m^3. The flooding dynamics of Luirk indicate a smaller catchment and substantially less apparent hydraulic head beneath the turlough. Due to Luirk’s low elevation and its proximity to the coast (<500 m), the water level was seen to oscillate with the tide. Oscillations of 10 cm were seen during low periods with oscillations becoming less distinct at greater depths and disappearing altogether at depths above 2.1 m (3.5 mASL).

Figure 4.5: Turlough depth data. Note: results are shown as depth, not stage (Gortboyheen is 14.8m higher than Luirk).

4.5.2 ERT Results

O’Connell et al. (2017) (see Chapter 3) confirmed the extension of faulting through Muckinish Lake offshore to submarine springs/sinkholes and tidal ingress of seawater along MacDermot’s Fault while offshore ERT provided evidence for diffuse groundwater discharge in the intertidal zone and pathways for groundwater movement through offshore springs/sinkholes. This allowed a comparison of increased sediment resistivity at offshore locations, identifying potential submarine discharge associated with bedrock faulting/karst zones.

T1 & T2 (MacDermot’s Fault)

The DD array for T1 from O’Connell et al. (2017), recorded 2 km inland from the shore at an elevation of 30-40 m above mean sea level, is presented in Figure 4.6. Expected ground conditions included glacial till deposits (GSI and RBD Consultants, 2004a) overlying pure-bedded and massive limestone (Pracht, 2004). Soil and rock resistivity is a function of the porosity, pore fluid resistivity and clay content (Archie, 1942; Waxman
and Smits, 1968) so lower resistivity would be expected for clay rich till than the limestone bedrock. However, fracturing and dissolution of the limestone would reduce resistivity values through increased water and/or sediment infill. Near surface resistivity values ranged from 100-320 $\Omega$m indicative of glacial till (O'Connor, 1998), thinning to the east where outcrop was noted during surveying. The resistivity of the underlying limestone bedrock is typical of values observed elsewhere in Ireland (O'Rourke and O'Connor, 2009). A vertical zone of reduced resistivity (A) indicated MacDermot’s Fault, while vertical zone (B) suggested a parallel fault and/or karst zone. Both lie above the saturated zone (6-14 m AOD (Perriquet, 2014) as outlined in Figure 4.6) and theoretical 2D forward modelling of A (Geotomo Software, 2002) with added errors to account for random noise equivalent to 5% where constructed (e.g. O'Connell et al. (2017)) suggests a 10 to 15 m wide, vertical low resistivity (~250 $\Omega$m) fault zone. Zone B resistivity values (> 640 $\Omega$m) suggest a minor feature with less water-filled porosity.

![Figure 4.6: T1 - Inshore ERT Profile of MacDermot’s fault from O'Connell et al. (2017). Saturated zone at high and low water table levels indicated by dashed lines.](image)

The T2 modelled resistivity from O'Connell et al. (2017) is presented in Figure 4.7. It is located ~7 m above mean sea level, at a distance of 0.2 km from the shore, and (a) displays the low tide profile while (b) presents the high tide profile. This time-lapsed survey across the fault over part of a tidal cycle during a prolonged period of low rainfall shows a low resistivity (~50 $\Omega$m) subsurface zone (C) at low tide that floods with saltwater as tidal levels increase, reducing resistivity to < 10 $\Omega$m. The observed resistivity in (C) is lower than observed for (A) which lies above the saturated zone. The low
resistivity signals from profiles T1 and T2 along MacDermot’s Fault (both inshore and along the coast) indicate that the weathered rock adjacent to the fault core is extremely permeable, connecting the uplands with the coast.

To further characterise MacDermot’s Fault, O’Connell et al. (2017) carried out offshore ERT profiles along strike of the fault and found evidence of its continuation offshore, running directly underneath a number of offshore depressions suspected of being submarine sinkholes/springs. To confirm the findings of O’Connell et al. (2017), and determine whether discharge was occurring from these springs, an electrical conductivity datalogger was installed on the sea floor within one such depression. The results (Figure 4.8) show a drop in salinity during periods of heavy rainfall indicating the discharge of freshwater near the logger. These results, combined with the findings of T1, T2 and O’Connell et al. (2017) indicate that MacDermot’s Fault likely acts as a flow pathway for much of the western edge of the catchment and continues out to sea where it discharges via the offshore springs.
Figure 4.8: Salinity data at offshore spring and daily rainfall.

T3 & T4

ERT profiles T3 (950m) and T4 (480m) are presented in Figure 4.9. Assuming thin soils based on observed outcrop during surveying, the bedrock resistivity is quite variable, especially at shallow depths. Low resistivity in the upper ~10 m (Figure 4.10) is indicative of the shallow epikarst. At greater depths, a number of anomalous low resistivity features (D, E, F and G in Figure 4.10) suggest either highly fissured/fractured rock zones or water-bearing conduits. Lowest resistivity can be observed at (E) suggesting that the suspected conduit crosses the profile in this location. A mean groundwater resistivity value of ~14 \( \Omega \text{m} \) (700 \( \mu \text{S/cm} \)) was observed for a borehole in the vicinity of T3 (Perriquet et al., 2014). Applying this value to a theoretical 2D forward model of feature (E) (Geotomo Software, 2002), with added errors to account for random noise equivalent to 5% (e.g. O’Connell et al. (2017), suggests a ~2 m water-filled conduit which may be a tributary conduit or a zone of highly fractured/fissured rock. Forward modelling of feature (D) suggests a 15-20 m diameter zone with a resistivity of ~ 150 \( \Omega \text{m} \) which, assuming little or no clay content, implies a 28% increase in secondary porosity suggesting significant fissuring of the limestone. This corroborates with the findings of Perriquet (2014) who assessed the response to recharge events in a nearby borehole and determined that the area is dominated by conduit flow. Furthermore, point E is a hydraulically coherent location for the Gortboyheen-coast linkage as its centre point lies between Gortboyheen estavelle (15.2 m) and the coastal springs (0 m). The Perriquet (2014) results for the aforementioned borehole indicated that the subsurface in this area was fracture/matrix flow driven.
Thus the anomalies at (F) and (G) may be the result of a localised fractured zone which may not be linked with the active conduit network.

Figure 4.9: ERT Profiles T3 and T4 within the catchment lowlands. Dashed black line indicates ~10 m epikarst.

4.5.3 Modelling

The ERT survey provides laterally continuous observations of resistivity variations in the limestone. Providing evidence for fault and conduit drive preferential pathways, an outline conduit network model could be built based on the ERT survey. The catchment was split into two drainage networks, MacDermot’s Fault to the west and the conduit network in the lowlands and the east.

4.5.3.1 MacDermot’s Fault Network

MacDermot’s Fault crosses the catchment north to south lining up with Muckinish Lake and the offshore depressions. ERT surveys carried out in two locations have indicated reduced porosities along the fault with associated resistivity variations indicative of survey positions above and below the saturated zone. The fault continues offshore and discharges at a number of submarine springs as seen by electrical conductivity measurements offshore (Figure 4.8). Based on these findings, the fault has been conceptualised as a stand-alone drainage channel fed by recharge from the western extent of the catchment and discharging at the submarine springs. The model includes a hydraulic linkage to Muckinish Lake which showed tidal oscillations with occasional rises of up to 1.5 m during wet periods which corresponds to the findings of Perriquet (2014). Mean discharge from the modelled fault system was estimated as 0.34 m³/s with peaks of 3.3 m³/s.
4.5.3.2 Lowland Network

Following on from the ERT investigations which indicated the presence of a ~2 m conduit in the lowlands, a number of iterations of the hydraulic model were constructed and tested to determine which hypothetical conduit network configuration would best match the observed behaviour (based on the water level in the turloughs). This process focussed on two elements:

- The relationship of Gortboyheen turlough with the underlying conduit network
- The relationship of Luirk turlough with the conduit network, and with Gortboyheen turlough.

**Gortboyheen Turlough**

The modelling process consisted of starting with a simple surcharge tank turlough with realistic pipe dimensions underneath (based on explored phreatic caves in the vicinity). From this point, new elements were added or removed in an iterative process with the conduit dimensions being regularly re-sized. These elements included a throttle, a bypass and additional inflow/outflow pipes. Each element brings an individual control to the calibration process, either altering the peak flood depth, the flood recession curve or both. The implementation of these elements was based on previous experience with the Gort Lowlands hydraulic model (Gill et al., 2013).

An overview of the Gortboyheen calibration process and the impact of each additional element is shown in Figure 4.10. At first (configuration A-1), the simple surcharge tank turlough model underestimated the flooding in the turlough so a throttle pipe was added just downstream of the turlough with a diameter of 40% less than the original pipe (B-1). The result of this throttle was a major increase in max flood depth and recession curve. To reduce the flood level and add additional control to the recession curve, a conduit bypass was added (C-1). This combination of a throttle and a bypass allows for more nuanced control over the peak flood and recession curve. However, as can be seen in C-2, the flood peak was still too low while the recession curve was too slow. The flood peak was corrected with the addition of a surface input (D-1, based on the presence of an actual spring feeding the turlough at approx. 28 m). The recession curve could
then be adjusted using a series of narrow multi-level outlets (between 20-23 m) which represent diffuse type drainage. These conduits and their elevations are based on the presence of exposed karst limestone in the turlough which would offer additional drainage capacity once the turlough reaches the required depth.

The most accurate calibration was found using an optimised conduit pipe diameter of ø=0.74 m for the bypass channel with an ø=0.53 m throttle (see E-1, Figure 4.10). On the surface, the elevated inlet was approx. ø=1000 mm and multiple outlets were approx. ø=300 mm each.

Luirk Turlough

Following the modelling of Gortboyheen, the hydraulic connection between the two turloughs was tested. Three hypotheses were tested:

- Luirk and Gortboyheen are connected in a continuous hydraulic head system (Configuration A)
- Luirk and Gortboyheen are connected in a discontinuous hydraulic head system (Configuration B)
- Luirk and Gortboyheen have separate drainage systems (Configuration C)

The conduit configurations and their optimised conduit diameters (ø) are shown in Figure 4.12. It should be noted that while the Gortboyheen bypass channel is not shown in these plots, it is included in the calibrations.

Configuration A comprises of a fully surcharged system during flooded periods due to a constriction (ø=0.5m) downstream of Luirk which impacts the entire system (additional constriction at Gortboyheen is still required to generate realistic hydraulic head in that turlough). In configuration B, the constriction at Luirk is less severe (ø=1.02m) which reduces hydraulic head in Luirk and the Gortboyheen-Luirk conduit. Gortboyheen turlough however is still surcharged due to the constriction immediately downstream of it. In configuration C, the systems have separate drainage systems and their hydraulic heads are independent.
Figure 4.10: Gortboyheen model schematics (column 1) and associated simulations (column 2).
Chapter 5

Figure 4.11: Model configuration schematics during flooded periods (not to scale & bypass/inlets are not shown).

Results for these configurations are shown in Figure 4.12 which shows that a continuous hydraulic head system (Configuration A) provides a reasonable estimate for Gortboyheen turlough but severely overestimates the water level in Luirk. Configurations B and C, however, provide higher precision for Gortboyheen and far more accurate results for Luirk. These results clearly indicate that configuration A is implausible while configurations B and C are possible. The underground constriction required to build hydraulic head and flood Gortboyheen is evidently not continuous within the conduit network. Thus while the water drained from these
turlough may or may not end up at the same outlet system reaching the sea, the constrictions influencing each turlough are hydraulically separate. This system dynamic is in contrast with the neighbouring Gort Lowlands catchment where the turloughs are linked in a continuous pressurised network. In the Gort system, the flooding of one turlough directly influences the behaviour of at least four others (Gill et al., 2013).

Figure 4.12: Simulation results for Gortboyheen and Luirk turloughs for configurations A, B and C.

Model efficiency, $R^2$ (goodness of fit – i.e. the Nash-Sutcliffe criterion) was calculated for configurations B and C using Equation 1.

$$
R^2 = \frac{\text{sum} (\text{obs}_i - \text{mod}_i)^2}{\text{sum} (\text{obs}_i - \text{mean}_i)^2}
$$

where $\text{obs}_i$ is the observed flow and $\text{mod}_i$ is the modelled flow at each timestep. For configuration B, efficiencies of 0.868 and 0.574 were obtained.
for Gortboyheen and Luirk respectively while configuration C displayed efficiencies of 0.878 and 0.589. The reduced precision for Luirk is due to its small sub-catchment towards the north of Bell Harbour, far from the rain gauge in the south. This makes it more susceptible to localised changes in rainfall.

Gortboyheen turlough is thus conceptualised as a surcharge tank for a large catchment, providing additional storage when the underground network has reached capacity. The primary mechanism causing this flooding in Gortboyheen is an underground constriction causing a significant build-up of pressure. During flooding, Gortboyheen turlough fills at a rate of approximately 0.65 m$^3$/s. This rate is generally consistent at all flood periods irrespective of the initial water level. In comparison, Luirk turlough serves a smaller catchment and its potential storage is required less often. It floods at a rate of approximately 0.1-0.15 m$^3$/s. However, during the wettest day of the 2014-2015 flood season, the inflow rate rose to 0.66 m$^3$/s. This implies that Luirk and Gortboyheen have a similar degree of constriction beneath them. However, Luirk turlough is rarely subject to the same hydraulic head as Gortboyheen turlough due to its smaller catchment.

The similarities between the turloughs are not reflected in their outflow rates. Gortboyheen drains at approximately 0.3 m$^3$/s whereas Luirk drains at approximately 0.15 m$^3$/s. The relatively slow outflow is a result of the proximity of Luirk to the sea. At high tide, the sea level is often higher than the water level in Luirk, reversing the hydraulic gradient, resulting in subtle flooding events reoccurring steadily over a 12.42hr cycle (during 2014-2015, the hydraulic gradient between Luirk and the sea ranged between 1.6% and -0.4%). A schematic illustration of the model and the locations of its major elements are shown in Figure 4.13. Discharge for the network was estimated at 1.2 m$^3$/s with peaks of 3.2 m$^3$/s.

The hydraulic model worked successfully in this study as a means to model the turloughs and investigate the underground linkages. However, it should be noted that the model is currently limited in scope due to the uncertainty of the southern catchment boundary. The model is built according to Drew
(2003) who delineated the surrounding catchments based on topographic divides and tracer tests in neighbouring Burren catchments. The southern boundary uncertainty derives from the 2-3° southwards dip of the Burren bedrock and the presence of impermeable clay horizons. While these horizons are thin (30-50 cm) and likely discontinuous, they obstruct vertical flow to a certain extent. The hydrogeology is further complicated by mineralised veins which can provide vertical flow paths as evidenced by the frequent occurrence of springs at the vein-wayboard intersections on the surface. In some instances, a subvertical cave system can develop (e.g. Pol Gonzo cave (Figure 4.1), in which a waterfall with discharge of up to 100 l/s is present (Bunce, 2010). Therefore, the ratio of vertical-horizontal and north-south flow within the unsaturated zone is uncertain.

![Figure 4.13: Conduit network model schematic based on Configuration B.](image)

Between 2014 and 2016, a series of tracer tests were carried out on behalf of the Geological Survey of Ireland in an attempt to better constrain the
southern boundary of the catchment (Drew and Bunce, 2016) The focus of this study was Carran Turlough (Figure 4.1) which was suspected to be draining north (Bell Harbour) and south (Fergus Springs) via separate outlets. The tracer study successfully connected both outlets to the south which indicated that the north-south boundary lies between Carran and Pol Gonzo Cave (which is understood to drain northwards as the hydraulic gradient between Pol Gonzo Cave and the Fergus Springs is insufficient). It is thus likely that the southern catchment boundary of Bell Harbour spans a region of up to 2-3 km in which recharge can flow north or south depending on localised geological features and aquifer saturation.

4.6 Conclusion

Previous studies in the Bell Harbour Catchment have often been hampered by the catchment’s complex mixture of upland, lowland and coastal karst with submarine and intertidal discharge. In this study, however, alternative techniques were used which suited the catchment’s particular characteristics. Through a combination of geophysics and conduit network modelling, the hydrogeology of the Bell Harbour Catchment, particularly its turloughs, has been elucidated.

In the Bell Harbour Lowlands, the shallow water table and its associated shallow karst features are highly suited for effective application of electrical resistivity tomography (ERT). These conditions, which are widespread in Ireland, allowed for the determination of porous, low resistivity zones which are probably linked with groundwater flow pathways. Using this approach, the initial conceptualisation of a karst conduit linking Gortboyheen to the sea was demonstrated to be true. Following this, with the addition of measurements from a number of surface water features, the subterranean flow network could be investigated by developing and simulating a number of theoretical karst conduit combinations.

The results of the conduit network models indicated that the hydraulic connection between Gortboyheen turlough and the underground network is likely to be a complex mixture of multiple inlets and outlets, a downstream constriction (throttle) and a bypass channel. With regards to the connection between Gortboyheen and Luirk turloughs, the model simulations indicated
that the turloughs are not connected via a continuous hydraulic head system. Instead, they are either drained separately or they could be linked in a discontinuous hydraulic head system. To determine which situation is occurring, future ERT work in Bell Harbour should focus on Luirk turlough and the locations of any low resistivity zones in its vicinity which could potentially link with the mainline conduit system from Gortboyheen. Further research into the southern catchment boundary will allow for a more accurate input signal to the model. This will enable greater accuracy in the estimation of discharge, which not only enters the turloughs, but also bypasses beneath them.

Overall, the combined ERT and conduit network modelling approach was successful as it allowed for the conceptual model of the catchment to be demonstrated and then further developed. The approach was well suited to the conditions within the catchment and could potentially be used in other lowland karst regions (e.g. the Yucatán Peninsula or west-central France).
Chapter 5: Discussion & Conclusions

5.1 Discussion

This thesis demonstrates how geophysical techniques including terrestrial and marine electrical methods and remotely sensed airborne techniques can be employed across the coastal zone to examine the hydrogeological properties of aquifers. From the determination of the microscale electrical properties of the sediments and rocks, to the macroscale examination of groundwater flowpaths, this work augments our understanding of groundwater movement in two very different aquifers in the coastal zone.

Though widely used for applications in groundwater resources (Bedrosian et al. 2015; de Souza Filho et al., 2010; Gondwe et al., 2012; He et al., 2014; Siemon et al., 2011; Wilson et al., 2013; Beamish, 2012; Fitterman and Deszcz-Pan, 1998; Kirkegaard et al. 2011), this research incorporates petrophysical modelling to provide a new application for AEM in the examination of the hydraulic properties of effective porosity and hydraulic conductivity on a regional scale, an approach not exercised to date by other authors. This work also explores the limits of quantifying the properties of near-surface clay-sand-gravel aquifers with sparse ground-based hydrogeological information.

The hydrogeophysical examination of the Dromiskin sand and gravel aquifer in Co. Louth demonstrates the use of the modelled electrical conductivity depth function derived from ground-based electrical resistivity data, to infer effective porosity with depth through the application of predictive petrophysical models (Bussian, 1983; Revil et al., 1998), constrained by a single borehole that has provided information on particle size distribution, water content and cation exchange capacity of the clay fraction as a function of depth. To date petrophysical modelling work has mostly been lab based (Kalinski and Kelly, 1993; Revil et al., 1998; Rhoades et al., 1976; Shah and Singh, 2005) or specific to oil exploration in the examination of reservoirs (Archie, 1942; Bussian, 1983; Waxman and Smits, 1968). Applied to the modelled electrical conductivity distribution across the aquifer, derived from remotely sensed AEM, these properties
have been up-scaled to infer hydrological properties, particularly hydraulic conductivity, across the entire aquifer.

Helicopter based TDEM systems (e.g., SkyTEM) typically provide the best approach for subsurface mapping. However, with the availability of the Tellus airborne FEM dataset (which was designed and driven more by exploration potential) this thesis provides an approach to maximise the use of the available data for hydrogeophysical applications. With limited coastal hydrogeophysics carried out in Ireland (Gibson et al., 2012), typically focused on the offshore environment (e.g., Garcia et al., 2014; Sacchetti et al., 2012) this work provides methodologies employing the Tellus dataset for the examination of similar sites across Ireland.

Future work should focus on an area with a spatially extensive in-situ hydrological observation network and complementary ground-based geophysical, geochemical and borehole data. Large-scale hydrogeological observatories have been established e.g., in the US (http://www.criticalzone.org), HOBE in Denmark (http://www.hobecenter.dk) and TERENO in Germany (Bogena et al., 2006), to determine processes that contribute to water and land management at catchment scale. Integrating geophysical techniques including airborne (e.g., SkyTEM and ground based approaches) with satellite remote sensing and soil moisture mapping, they provide low-cost, low-disruption strategies that can upscale limited measurements to quantify catchment-scale groundwater resources for sustainable abstraction rates and the mitigation of flood or drought risk. The creation of an Irish hydrogeological observatory combining the wealth of hydrogeological information established to date through the WFD integrated with Tellus airborne data, satellite imagery, existing monitoring points and any other existing relevant data e.g. the Irish Soil Information System, would be instrumental in informing future work, allowing focussed planning based on in-situ monitoring networks requiring sparse ground based geophysical data.

Research into groundwater movement and SGD in the karst aquifer on the south coast of Galway Bay has, to date, focused on groundwater chemistry, biogeochemical reactions, nutrient and contaminant loading, thermal
imaging, saltwater intrusion, groundwater flow paths, groundwater discharge and residence times (Cave and Henry, 2011; McCormack et al., 2014; Perriquet et al., 2014; Smith and Cave, 2012; Wilson and Rocha, 2012). In this thesis, terrestrial and marine ERT methods, integrated with existing hydrogeological information have confirmed existing hypotheses on groundwater movement and SGD in two adjacent catchments within the aquifer, providing new insights into the fault controlled groundwater pathways in one catchment and large conduit flow in the other.

Terrestrial ERT methods are routinely used in hydrogeological applications (e.g., Nyquist et al., 2008; Comte & Banton, 2007; Nguyen et al., 2009; Amidu and Dunbar, 2008) and karst areas e.g. Zhu et al. (2011). Marine applications typically allow qualitative investigation of diffuse SGD through sediments (Belaval, 2003; Breier et al., 2005; Day-Lewis et al., 2006; Manheim et al., 2004). No authors have employed the marine surface-towed ERT technique in a karst aquifer setting. Applied across this coastal karst aquifer, these techniques have provided evidence of how the structural geology and conduit networks influence SGD and saltwater ingress across the coastal zone.

In the Gort-Kinvarra catchment, large diameter conduit flow (Drew and Daly, 1993) has been confirmed from a central, subterranean-fed lake to springs in the intertidal zone of Kinvarra Bay. Though intertidal discharge is known to dominate this catchment (Boycott and Bruce, 2003; McCormack et al., 2014), the marine ERT has confirmed the presence of discrete and other potentially ephemeral SGD points within Kinvarra Bay. In the neighbouring Bell Harbour catchment, time-lapsed data has demonstrated tidally-influenced saltwater intrusion into the fault through the low lying part of the catchment, confirmed the offshore extension of faulting and its influence on groundwater movement and examined diffuse discharge from epikarstic intertidal springs.

This work has informed the conduit network modelling of the Bell Harbour catchment to characterise flow pathways and their likely hydraulic mechanisms, with particular emphasis on conduit linkages between two ephemeral lakes and the sea. An initial conceptualisation of a karst conduit
linking the main lowland turlough, Gortboyheen, to the sea was demonstrated to be true. With the addition of measurements from a number of surface water features, the subterranean flow network could be investigated by developing and simulating a number of theoretical karst conduit combinations. The resultant model is a complex mixture of multiple inlets and outlets, a downstream constriction and a bypass channel. The model simulations also indicated turloughs that are not connected via a continuous hydraulic head system. This approach can assist in coastal zone management planning and could potentially be used in other lowland karst regions (e.g. the Yucatán Peninsula or west-central France).

5.2 Conclusions
This thesis has utilised current and novel practices in airborne, terrestrial and marine geophysical techniques to examine groundwater characteristics in coastal aquifers, investigating the link between geophysical and hydrogeological properties in two very different coastal aquifer types.

High-resolution AEM and ground-based electrical resistivity data have been modelled to infer the sub-surface electrical conductivity distribution and hydrological properties, particularly hydraulic conductivity, of a clay-sand-gravel aquifer. The approach has used the electrical conductivity function of depth to infer interconnected porosity with depth constrained by a single borehole that has provided information on particle size distribution, water content and cation exchange capacity of the clay fraction as a function of depth.

This methodology can assist with catchment-scale three-dimensional quantitative assessments of groundwater resources in such aquifers only if there is a spatially extensive in-situ hydrological observation network and complementary ground-based geophysical, geochemical and borehole data.

The geophysical techniques deployed in the coastal karst aquifer have provided evidence for both structural and dissolitional influences on groundwater pathway development, complementing ancillary geological
and hydrogeological evidence for fault controlled groundwater pathways and conduit flow.

The characterisation of SGD and groundwater pathways in coastal karst aquifers is critical for understanding climate-change and anthropogenic impacts such as sea level rise as an assessment of the contribution to groundwater fluxes from, inter alia, diffuse sources and discrete pathways is essential for the modelling strategy required to forecast the impacts on groundwater. By imaging subsurface structures and discrete zones that control freshwater-seawater interactions existing hypotheses on groundwater movement and SGD in the Bell Harbour and the Gort-Kinvarra catchments have been confirmed and expanded on, informing hydraulic modelling on a catchment level. The deployment of these methods enhances the quality of the models used to assess the hydraulics of karst aquifers and can be used in coastal zone management planning.

Through a combination of geophysics and conduit network modelling, the hydrogeology of the Bell Harbour Catchment, particularly its turloughs, has been elucidated. The approach was well suited to the conditions within the catchment and could potentially be used in other lowland karst regions (e.g. the Yucatán Peninsula or west-central France).

This research augments the understanding of groundwater movement in both the sand and gravel and karst limestone aquifers to inform models that provide a baseline for future investigations of groundwater regime changes and increased saline intrusion as a function of climate-driven sea-level rise and increased flood events.

The work presented in this thesis demonstrates innovative approaches to the understanding of the physical characteristics of coastal aquifers including their geology, hydrogeology, linked groundwater and surface water systems and hydraulic properties.

The methodologies developed within this PhD provide transferrable practices to assist in catchment-scale assessments of coastal aquifers.
5.3 Future Work

Based on the findings of this PhD thesis, the following actions are recommended for future work:

- TELLUS airborne surveying continues across the island of Ireland under the governance of the GSI with 50% of the island completed to date. This extensive resource of radiometric, magnetic and AEM data should be interrogated, particularly in coastal settings and in aquifers with known hydrogeological characteristics and in-situ monitoring programs to further refine the remote sensing approach described in this thesis. With its continued acquisition across the island of Ireland the approach presented in this thesis provides a significant new application to the Tellus AEM dataset for the regional reconnaissance of groundwater resources.

- The potential integration of airborne AEM and AR data with satellite remote sensing data e.g. the new European Space Agency Copernicus programme, utilising AR data from earth observation satellites, for soil moisture and soil mapping for catchment scale management could be investigated.

- The airborne data currently being acquired across Co. Galway extends to Kinvarra Bay and across the coastal zone of the karst aquifer examined in this thesis. The acquired AEM data should be modelled in conjunction with the ERT data contained in this thesis and possibly ground based electromagnetics, to examine the electrical conductivity distribution in the coastal zone, demonstrate the saline influence in to the karst aquifer with the possibility of examining the contribution of large dimension conduit flow paths as observed in Chapter 3.

- Locally, additional terrestrial ERT should be acquired in the Bell Harbour catchment to establish a broader/pseudo 3D resistivity distribution in order to confirm the northward continuation of conduits, to establish any connection between Mac Dermot’s fault and the
lowland network and to further refine the hydraulic model contained in Chapter 4.

- Any future ERT work should also focus on Luirk turlough and the locations of any low resistivity zones in its vicinity which could potentially link with the mainline conduit system from Gortboyheen. This will enhance the flow regime model, leading to improved estimation of discharge.
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Appendix 1: Summary of Geophysical Methodologies

The theory of the geophysical methods employed for this research are discussed in detail in various texts e.g. (Keller and Frischknecht, 1966; Nabighian, 1988; Parasnis, 1990; Reynolds, 1997; Telford, 1990).

**Airborne Electromagnetics (AEM)**

Electromagnetic techniques operate by measuring the response of the ground to the propagation of electromagnetic waves. Electromagnetic waves consist of coupled orthogonal electrical and magnetic fields. Geophysical methods employing electromagnetic waves operate by generating a primary electromagnetic field through a transmitter coil which propagates both above and below the ground and is recorded by a receiver coil. As the field propagates through the ground, the magnetic field induces eddy currents in the presence of conductive subsurface media, generating a secondary field. The response measured by the receiver coil is a combination of this secondary field in addition to the primary field transmitted through the air and will differ in phase and amplitude to the transmitted primary field. The variations in these components imparts information on the electrical properties of the subsurface.

Electromagnetic induction techniques are classified as either frequency-domain (FEM) or time-domain (TEM). FEM uses a fixed frequency or range of frequencies and records the primary and secondary field as detailed above. TEM employs a transient signal rather than a fixed frequency signal and measures the secondary field as a function of time after the transmitter is turned off. The depth of exploration of FEM depends on the transmitter/receiver coil separation, orientation, transmitting frequency and subsurface conductivity and is typically limited to the near surface (<100 m). The depth of exploration of TEM is a function of the coil size and can achieve depths of hundreds of meters (dependent on subsurface conductivity). As only FEM data were used in this research the following discussion relates to FEM acquisition only.

Wave amplitudes of primary fields decrease exponentially with depth with secondary fields also attenuated as they propagate to the surface. Wave propagation is predominantly governed by the transmitting frequency and
the conductivity of the subsurface. For airborne surveys, the transmission frequencies are typically in the very low frequency range (1 - 25 kHz). The depth of penetration or ‘skin depth’, δ, is defined as the depth at which the electromagnetic plane wave attenuates to 1/e (or 37%) of the transmitted signal and is represented by the following equation:

\[ \delta = \left( \frac{2}{\sigma \omega \mu_0} \right)^{1/2} \quad (1) \]

where \( \sigma \) is the conductivity in Siemens per metre (S/m), \( \mu_0 \) is the permeability \( (= 4\pi \times 10^{-7} \text{ Henries/m}) \) and \( \omega \) is the angular frequency of the plane wave \( (2\pi f) \), where \( f \) is the operating frequency in Hertz (Hz) such that:

\[ \delta = 503(\rho/f)^{1/2} \quad (2) \]

where \( \rho \) is the resistivity in Ohm metres (Ωm). For airborne surveys, increased flight altitude also affects the depth of investigation by decreasing coupling with the ground, reducing the signal to noise ratio.

The AEM data used in this research was collected during the Tellus (TEL) and Tellus Border (TB) surveys, with acquisition and initial processing detailed in Hodgson and Ture (2013) and SGL (2012). Specifications for each survey are detailed in Table A1.1. Four frequencies of AEM data were acquired; 0.9 kHz (912 Hz), 3Khz (3005 Hz), 12 kHz (11962 Hz) and 25 kHz (24510 Hz). Each recorded frequency comprised two components; a real or in-phase component (P) and an imaginary or quadrature component (Q), both measured as secondary to primary field coupling ratios in parts per million (ppm):

\[ P = \text{Re} = \frac{H_z}{H_0} \quad (3) \]
\[ Q = \text{Im} = \frac{H_q}{H_0} \quad (4) \]

where \( H_z \) is the vertical anomalous field and \( H_0 \) is the horizontal primary field.

\( P \) and \( Q \) were transformed to apparent resistivity and apparent depth for each frequency using a half-space model (Beamish et al., 2006). This transformation process used a program based on a GTK version of TRANSAEM. Apparent resistivity was calculated using a look-up procedure which employs the in-phase and quadrature data components at each frequency (Hodgson and Ture, 2013). The ground was modelled as a single layer half-space assuming a constant lithology.
Skin depths, calculated using equation (2), for each of the four frequencies used, at varying ground conductivities, are plotted in Figure A1.1 to give an indication of potential depths of investigation. The data used for this research included depth calculations for the Co. Louth and Co. Down coastal areas for which the 900 Hz data had a mean depth of 85m, the 3 kHz data had a mean depth of 73m, the 12/14 KHz had a mean depth of 43m while the 25 kHz had the shallowest depth of investigation with a mean depth of 27m.

Figure A1.1: Skin depths calculated for Tellus Border and Tellus frequencies of 0.912, 3.005, 11.962 and 24.510 kHz for ground conductivities up to 1000 mS/m.

The airborne electromagnetic data was initially examined in a report by O'Connell and Daly (2013). SGL (2012) suggest the AEM data quality becomes degraded above a survey altitude of 75m while Hodgson and Ture (2013) indicate that the signal cannot be resolved from background noise at altitudes above 180m. For the work presented here, a cut-off flight altitude of 100m was chosen similar to Beamish and White (2012). Much of the cultural noise from urban areas was removed in this cut-off. Filtering also included the elimination of cultural effects associated with road
networks, powerlines, rail tracks and urban areas, etc.. To determine the electrical conductivity distribution as a function of depth across a coastal sand and gravel aquifer in Co. Louth, the AEM P and Q components were modelled in conjunction with a ground-based electrical resistivity survey and sparse hydrogeological information, using a well-established inversion technique (Beamish (2012); Farquharson et al., 2003); Viganotti et al. (2013). The inversion algorithm (University of British Colombia, 2000, 2006) uses an *a priori* conductivity starting model and iterates to the simplest model that fits the observations. The modelling of the electrical conductivity distribution was constrained with direct measurements of electrical conductivity from ground based geophysics.

**Airborne Radiometrics**

K, Th and U are naturally occurring radioactive isotopes that are present in small amounts in soils and rocks. The energy and intensity of the gamma rays that these isotopes emit are sufficient to enable detection by gamma ray spectrometry. The gamma rays originate from the upper ~ 0.3 m of Earth’s surface and attenuate exponentially with increased flight elevation at a rate of ~ 0.5 per 100 m (IAEA, 1991). Radioactive decay produces alpha, beta and gamma rays of which the gammy rays are high energy electromagnetic waves (frequencies > \(10^{19}\) Hz). The principal naturally occurring radioactive isotopes are potassium (K), thorium (Th) and uranium, (U). Water saturation is the primary cause of attenuation of emissions with density and porosity being secondary variables.

The radiometric data used in this research was collected during the Tellus (TEL) and Tellus Border (TB) surveys, with acquisition and initial processing detailed in Hodgson and Ture (2013) and SGL (2012). Specifications for each survey are detailed in Table A1.1. The data included a total count of emissions (counts/s) with individual concentrations of K (%), Th (ppm) and U (ppm) emissions. Standard radiometric survey calibrations and corrections were applied to the data by the contractor (SGL, 2012) based on the International Atomic Energy Agency report (IAEA, 1991) and the Geological Survey of Canada processing report (Grasty, 1972). A cut-off survey height of 100m was applied. A ternary plot (RGB colours) of the radiometric data was created from the K, Th and U concentrations. The areas in the plot that are white indicate high concentrations of K, Th and U.
while areas that are black indicate low concentrations of all three. Near-surface water such as lakes and rivers and peat deposits (due to their saturation) attenuate the signal (black). Total count values were also plotted. Low Total count values indicate low emission zones. The ternary images and contour plots of total count values (ms\(^{-1}\)) were visually examined.

<table>
<thead>
<tr>
<th></th>
<th>TEL</th>
<th>TB</th>
</tr>
</thead>
<tbody>
<tr>
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</tr>
<tr>
<td>Line direction</td>
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<td>345 degrees</td>
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<tr>
<td>Min. survey altitude (rural)</td>
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<td>59m</td>
</tr>
<tr>
<td>Min. survey altitude (other)</td>
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<td>240m</td>
</tr>
<tr>
<td>Typical survey speed</td>
<td>70m/s</td>
<td>60m/s</td>
</tr>
<tr>
<td>Electromagnetic Sampling</td>
<td>0.25 sec</td>
<td>0.1 sec</td>
</tr>
<tr>
<td>Radiometric Sampling</td>
<td>1.0 sec</td>
<td>1.0 sec</td>
</tr>
</tbody>
</table>

Table A1.1: Specifications for TEL and TB airborne geophysical surveys.

**Electrical Resistivity Tomography (ERT)**
The fundamental law governing electrical resistivity is Ohm’s Law which determines the resistance, \( R \), for an electrical circuit as

\[
R = \frac{V}{I}
\]

(5)

where \( V \) (measured in volts, V) is the potential difference across a resistor, \( I \) (measured in amps, A) is the current passing through the resistor and \( R \) is measured in Ohms (\( \Omega \)). The resistance, \( R \), is proportional to the resistor length, \( L \), and inversely proportional to the cross-sectional area, \( A \)

\[
R \propto \frac{L}{A}
\]

(6)

This can be written as:

\[
R = \rho \frac{L}{A}
\]

(7)

where \( \rho \) is the constant of proportionality which is the ‘true’ resistivity measured in \( \Omega \text{m} \).

To measure \( \rho \), a generalised electrode configuration comprising two current electrodes C1 and C2, two potential electrodes P1 and P2, with current I and voltage V is used. Where the current electrodes (C1 and C2) are separated by a finite distance, the potential distribution at P1, being affected by the current flow from C1 and C2, is equal to the sum of the
voltages from both electrodes. In non-homogenous ground the calculated resistivity is no longer a 'true' resistivity but is an 'apparent' resistivity, $\rho_a$, determined as a product of a geometric factor $k$ dependent on the electrode array configuration and the measured resistance. The geometric factors for the standard Wenner-Schlumberger array and Dipole-Dipole arrays are $k = \pi a (n + 1)(n + 2)$ and $k = \pi a (n + 1)(n + 2)(n + 3)$ respectively where $a$ is the dipole length and $n$ is the dipole separation factor.

The main factors considered in determining which electrode array to use include depth of investigation, required resolution, array sensitivity and whether the target is principally localised or extends laterally or vertically e.g., localised features in a karstic environment, faults or sand lenses. The Wenner-Schlumberger array is generally considered to be good at resolving horizontal features while having relatively good depth resolution while the Dipole-Dipole array is good at resolving vertical and dipping structures though it suffers from poor depth resolution (Binley and Kemna, 2005; Dahlin and Zhou, 2004).

In the towed system, each array configuration records ten resistance values through the water column and beneath the sea bed at each survey interval through combinations of a current electrode pair (C1 and C2) and 10 potential electrodes pairs (P1, P2, ……P11) (see Figure A1.2). 1D soundings, comprising 10 resistance values, are continuously recorded at intervals determined by the electrode spacing and survey speed, building up a 2D pseudosection of the water column and sub-seabed apparent resistivities. Errors in typically result from factors including cable offsets due to wind and wave action, electrodes snagging vegetation and random system noise (Day-Lewis et al., 2006).
Each array configuration results in a different response as the pseudo-section is not a true vertical cross section (Binley and Kemna, 2005) while the spatial sensitivities of each array type will also vary. To illustrate this, sensitivity values were calculated for a homogenous medium (Figure A1.3) using the Res2Mod software (Geotomo Software, 2002) assuming the configurations in Figure A1.2 and an electrode separation of 5m. The area of the subsurface that influences the apparent resistivity most is indicated by the red and orange range of sensitivity values. The sensitivity patterns for the Wenner-Schlumberger and modified Wenner arrays in which the current electrodes are central to the potential electrodes indicate a broad, centralised sensitivity pattern while the Dipole-Dipole array (Figure A1.3c), indicates that the highest sensitivities are offset.
ERT surveys produce ill-posed and non-unique resistivity inversions (Loke et al., 2013) due to the limitations on the number of electrodes used and the surface coverage achievable. Constraining the model goes some way towards addressing this problem. For marine towed arrays, the addition of the water depth and water layer resistivity improves the model resolution by removing the water layer from the inversion. For dynamically acquired data, the frequency of recorded data points is dependent on survey speed such that the distance between measured data may be much less than the electrode separation. The measured data is unlikely to resolve structures smaller than half the electrode separation, (Loke, 2010, 2012).
Appendix 2: Summary of Papers & Thesis Outputs

Papers:
The work presented in this thesis forms three papers currently in submission or in prep.;

- Chapter 2 forms a paper submitted to Water Resources Research. As lead author, I was responsible for analysis of the airborne data, planning, acquisition and modelling of the supplementary ground-truth geophysics, modelling of the 1-D electrical conductivity distribution, assistance with the petrophysical modelling and hydrogeophysical interpretation, all maps and figures and a major part of the writing.

- Chapter 3 forms a paper accepted by the Near Surface Geophysics journal. As lead author, I was responsible for planning and acquisition of all marine and terrestrial ERT, forward and inverse modelling, time-lapse analysis, interpretation of ERT modelling and sub-bottom profiling outcomes, all maps and figures and all of the writing.

- Chapter 4 forms a paper published in the Journal of Hydrology: Regional Studies. As co-author, I was responsible for planning and acquisition of geophysical data, forward and inverse modelling, geophysical interpretation, maps and figures relating to the geophysical data and a major part of the writing related to the geophysics.

Articles:

Presentations:
Appendices


O’Connell Y., Geophysics in Karst, IAH Karst Fieldcourse, Lisdoonvarna, May 2014.

Posters:


Appendix 3: Paper Acknowledgements

Chapter 2 Acknowledgements
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Appendix 4: Misfit Analysis of Dipole Moment Variants

The misfit obtained using dipole moments of 1.0, 1.8, 2.0 and 2.2 has been plotted in Figure A4.1 below. Very little variation was observed between dipole moments of 1.8, 2.0 and 2.2.

![Figure A4.1: Misfit values across 5 x 5 km area for dipole moment factors of 1.0, 1.8, 2.0 and 2.2 to determine optimum dipole moment.](image)

The misfit variation between dipole moments of 1.8, 2.0 and 2.2 did not result in broad increases or decreases. This is illustrated by a histogram analysis of the misfits (Figure A4.2) in which a dipole moment of 2.0 was determined to produce the best outcome.
Figure A4.2: Corresponding histogram of misfit values for dipole moment factors of 1.0, 1.8, 2.0 and 2.2 to determine optimum dipole moment.