

Investigating the Geothermal Energy Potential and Permafrost Structure in Nunavut with
Magnetotelluric Data

by

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Abstract

To reduce reliance on fossil fuels, Canada is investigating the development of alternative energy systems that have lower carbon emissions than those currently in use. This is particularly important for communities in the Canadian Arctic, where the vast majority of electricity and heat is currently generated by burning fossil fuels. Geothermal energy systems have the potential to provide an alternate source of heat in this region, and their feasibility in remote Northern settlements is being investigated.

Knowledge of subsurface rock type and groundwater conditions is an essential part of exploration for geothermal energy. In summer 2023 a group from the University of Alberta collected magnetotelluric (MT) data at two communities in Nunavut to investigate the subsurface conditions. The MT method can determine the subsurface electrical resistivity to depths of several kilometers. Resistivity is a parameter that is sensitive to the temperature and porosity of the subsurface, and to the quantity and salinity of groundwater. At Cambridge Bay, MT data was collected at 26 stations in July 2023 in an east-west array that covered an area of 17 x 8 km. At Resolute Bay, 33 stations were recorded in August 2023 in a region that was 8 x 5 km. Data from an additional 11 stations were previously recorded at Resolute Bay in 2022. The time series data were processed to give frequency domain data in the frequency band 500 – 0.001 Hz. The data were then used to generate models of subsurface resistivity using a combination of 1-D and 3-D inversion methods. Interpretation of the resistivity used knowledge of subsurface temperatures and experiments on frozen / unfrozen sedimentary rocks that contained saline ground water.

At Cambridge Bay the resistivity model was characterized by two layers: (1) a 100 m thick surficial layer that has a low-resistivity due to the presence of limestone saturated with partially frozen saline pore water and (2) a deeper high-resistivity layer interpreted as limestone containing low

salinity pore water. Knowledge of the geothermal gradient suggests that the upper part of this surface layer was frozen and the lower part unfrozen. At Resolute Bay, the resistivity model was characterized by three layers: (1) a high-resistivity limestone layer containing partially frozen saline pore water (2) a low-resistivity layer containing unfrozen saline pore fluids, and (3) a deeper high-resistivity layer containing low salinity pore water. Porosity estimates of the subsurface below Cambridge Bay were low, indicating poor natural reservoir potential, whereas the subsurface below Resolute Bay may have greater porosity. These results also indicate that the geothermal gradients at both communities are relatively low, suggesting that direct use geothermal installations are more feasible at each location, and engineered geothermal systems (EGS) may be required to produce the reservoir characteristics required for geothermal use.

The low resistivity of the 100 m thick surface layer at Cambridge Bay may have originated in the diffusion of seawater from the surface when the land surface was submerged after the retreat of the Laurentide Ice sheet. Simple 1-D diffusion modelling is consistent with the timescales and layer thickness of approximately 100 m.

Preface

This thesis is an original work by Andrew Williamson. It has been written to fulfill the partial requirements for the degree of Master of Science at the University of Alberta.

Chapter 5 describes the investigation of geothermal potential at Cambridge Bay, Nunavut using magnetotelluric data. Content of this chapter was submitted as part of a technical report to RESPEC in March, 2024. Chapter 7 details a set of numerical experiments to expand upon these results. Together, this content is currently being prepared for publication.

Chapter 6 describes the investigation of geothermal potential at Resolute Bay, Nunavut using magnetotelluric data. Content of this chapter was also submitted as part of a technical report to RESPEC in March, 2024 and this content is also being prepared for publication.

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Chapter 1: Introduction

1.1. Geothermal energy in Canada

The Canadian government is exploring opportunities for alternative energy sources to reduce carbon emissions. The country's energy needs are currently being met largely by hydroelectric power, fossil fuels, nuclear, and a growing volume of solar and wind power. Among the available choices, research is underway to assess the feasibility of geothermal energy as a potential alternative in certain regions. There are several advantages that geothermal energy resources possess compared to other options (Grasby et al, 2012). Firstly, geothermal energy is capable of providing consistent baseload power. Secondly, a number of technologies exist that allow geothermal energy to be used for a wide range of subsurface conditions. Lastly, geothermal energy resources have a relatively low environmental impact compared to most other energy sources. Canada has sizeable and varied geological resources that provide many opportunities for incorporating geothermal energy into the nation's growing alternative energy portfolio (Figure 1.1).

Rapid adoption of renewable energy resources is motivated primarily by the desire to reduce greenhouse gas (GHG) emissions that are responsible for anthropogenic climate change. Warming temperatures are predicted to disastrously impact many aspects of life on Earth, including human health, water resources, agricultural production, and global forests and biodiversity (Hitz & Smith, 2004; Abbass et al., 2022). The implementation of alternative energy resources is required to pivot the global community into a position that is capable of weathering the most dramatic effects of climate change.

Globally, Canada has some of the highest energy use per capita in the world, primarily due to a high demand for heating in the cold Canadian climate, and the strong presence of energy-intensive industry (Ritchie et al., 2020; U.S. Energy Information Administration, 2025). The colder climate in Canada necessitates higher energy requirements for space heating across the country. Additionally, because of the abundant natural resources, Canada produces an enormous amount of energy, which lowers the cost of energy in Canada, driving energy use. In 2022, fossil fuels made up 77% of Canada's total energy supply (Natural Resources Canada, 2025), a vast portion of which was used for space heating and cooling, and electricity production. Thus, finding alternative

resources that can more cleanly produce electricity and provide heat for buildings could dramatically reduce the net carbon that is emitted by Canadians.

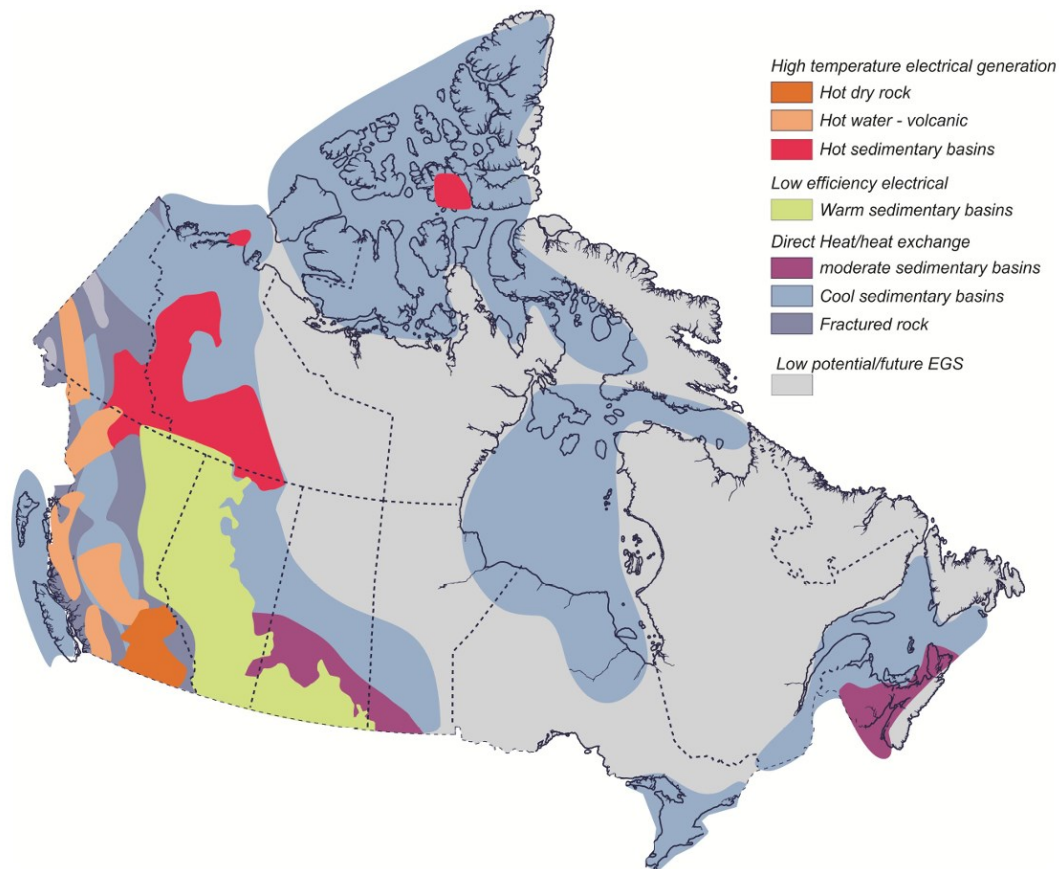


Figure 1.1: A map showing the distribution of geothermal potential in Canada based on end use. From Grasby et al., 2012.

Past research has established that Canada sits in a favourable location for geothermal development, partly due to high quality geothermal resources in British Columbia, Yukon, and the Northwest Territories, and partly because of vast sedimentary basins that could also be suitable for geothermal development (Figure 1.1) (Jessop, 2008; Grasby et al., 2012). Previous pilot projects, such as the Meager Creek geothermal project and the geothermal feasibility study at the University of Regina in the 1970s, demonstrate that geothermal resource development can be successful in Canada (Grasby et al., 2012). Extensive oil and gas exploration in the Western Canada Sedimentary Basin (WCSB) also have shown the region to contain significant reservoir potential that could be tapped into using geothermal technologies. A major challenge of developing geothermal resources in Canada is the often large distances between major population centers, and the low population

density of many regions with geothermal potential (Grasby et al., 2012). At the moment, the prospect of developing geothermal resources at more remote communities in the north are being studied, although a number of unique challenges exist.

1.2. Renewable energy sources and geothermal energy in Nunavut

Presently, Arctic communities in Canada are heavily reliant on fossil fuels for their heating and electricity requirements, with more than 170 remote indigenous communities almost completely reliant on diesel generators for electricity generation. In 2018, Nunavut residents consumed 79% more petroleum than the national average (Government of Nunavut, 2018). Arctic communities also lack interconnected energy infrastructure, with most communities belonging to independent power grids, which can make power outages dangerous in the extreme conditions of the Arctic. Energy production is entirely contained within the community-run facilities; petroleum fuel must be imported during fairweather seasons to use for the remainder of the year. The mass import and storage of these petroleum fuel sources pose heightened risks of fuel spills and environmental contamination (Poland et al., 2003). These factors present serious risks that could be avoided with the implementation of energy sources that do not require fossil fuels.

Instead, the Government of Canada and the province of Nunavut are looking towards investing in renewable energy sources to offset carbon usage and provide stable alternatives. Solar, wind, and hydropower currently dominate new renewables installations in Canada, with others such as geothermal and bioenergy trailing significantly (Canada Energy Regulator, 2020). Unfortunately, the Arctic poses serious restrictions on the efficacy of more popular alternative energy technologies. Limited daylight at certain times of year severely restricts the amount of power that can be generated using solar panels. As well, Arctic temperatures can become too cold for the operation of conventional wind turbines. Energy generation in the far north must be continuous and reliable in the extreme climate faced by those living in the Arctic.

In some Arctic communities, including Cambridge Bay and Resolute Bay, Nunavut, the potential of geothermal energy is being investigated. In considering the feasibility of harnessing geothermal resources, one must have knowledge of the local geology; understanding the local geological structure and heat flow is paramount to the successful engineering of geothermal systems. In northern latitudes, the presence of permafrost can significantly raise the cost of geothermal drilling, as can the implementation of enhanced geothermal systems, which may be required if ideal

geothermal conditions are not present. Most possibly, geothermal need not completely replace fossil fuels in northern communities, but could instead be integrated with other renewable energy sources to reduce carbon emissions and foster energy independence. A number of geothermal technologies exist for direct use, such as space heating or greenhousing that could be implemented without the need for exceptionally high subsurface temperatures (Miranda et al., 2022). Alternatively, geothermal systems could utilize waste heat that is produced by existing diesel generators to store heated water below the surface.

Information about the subsurface is generally scarce at the Arctic locations that are looking to develop geothermal resources. Further research is required to assess the geothermal potential at these communities in order to identify the economic feasibility of developing geothermal infrastructure. To this end, this thesis presents the results of preliminary study of the geothermal resources available at Cambridge Bay and Resolute Bay.

1.3. Geothermal exploration

The identification of economic geothermal resources often requires a multi-faceted approach to exploration. This includes the use of a variety of geological and geophysical techniques to study the efficacy of subsurface resources and form a model, which is eventually validated by drilling. Hickson and Yehia (2014) discuss the stages of geothermal exploration from a commercial perspective, whereby progressive exploration and assessment techniques are used to manage risk.

Typically, geothermal exploration programs begin with the use of broad, less expensive exploration techniques, such as geological mapping and sampling, geochemical analysis, and geophysical surveying. Geological methods involve the mapping of surface or near-surface features to gain an understanding of how the geothermal prospect was formed. Geochemical analysis can provide information about composition of geothermal reservoirs, which can be indicative of reservoir temperature, composition, and flow direction. Complementarily, geophysical surveying provides data to better constrain the deeper structure of buried targets, which can be used to identify subsurface targets, and to estimate properties of geothermal resources, such as the temperature, porosity, permeability, and volume. Together, these techniques are used to produce a robust model of geothermal prospects, which is essential to evaluate the economic viability of the resources of interest.

This thesis will focus on the use of geophysics for Arctic geothermal exploration, with a particular emphasis on electric and electromagnetic (EM) methods. These methods measure the electrical resistivity of the subsurface, which is sensitive to temperature and porosity, two properties that are characteristic of geothermal systems. Of this subset of geophysical techniques, natural-source magnetotellurics (MT) is a method that is well-suited for geothermal exploration in the Arctic, since it is capable of investigating to depths often required for geothermal exploration, and it can do so without requiring an artificial power source. Other electrical and EM methods are generally less capable of imaging to the depths required, and often involve more complicated or expensive data collection. Thus, the MT method is an appropriate technique for the preliminary investigation of geothermal potential.

1.4. Magnetotelluric investigations at Cambridge Bay

In July 2023, data were collected at 26 MT stations in a 17 km x 8km area at Cambridge Bay (CB), Nunavut with the goals of identifying and assessing potential geothermal resources in the subsurface. These data were collected as part of a geothermal feasibility study arranged by RESPEC on behalf of the Qulliq Energy Corporation, the sole energy provider in Nunavut. Analysis of the Cambridge Bay dataset suggests that the electrical structure below the surface is primarily 1-D in nature, that is, it largely varies only with depth. As a result, a series of forward and inversion models in both 1-D and 3-D were produced: the 1-D inversion models were produced using the OCCAM 1-D inversion algorithm (Constable et al., 1987), and 3-D inversion models were produced using the ModEM inversion algorithm of Kelbert et al. (2012). The 1-D and 3-D modelling results were consistent and were used to characterize the subsurface as: (1) CB1, a 100 m thick surficial layer that has a low resistivity due to the presence of limestone saturated with partially frozen saline pore water and (2) CB2, a deeper high resistivity layer interpreted as limestone containing low salinity pore water. Knowledge of the geothermal gradient suggests that the upper part of this layer was frozen and the lower part unfrozen. The thin layer of low resistivity is inferred to be a result of seawater that had diffused through the rocks from the surface, and additional exploration of this hypothesis is performed using numerical modelling in Chapter 7. These results require additional evidence to confirm, but ultimately suggest that Cambridge Bay is poorly suited for the development of conventional geothermal systems, however, direct use or engineered geothermal systems could provide alternative options.

1.5. Magnetotelluric investigations at Resolute Bay

In August 2023, data were collected at 33 MT stations in an 8 km x 5 km area at Resolute Bay (RB), Nunavut with the same goals as the Cambridge Bay survey. Previous temperature log measurements at Resolute Bay suggested that the local geothermal gradient is anomalously high. Dimensionality analysis of the Resolute Bay dataset revealed that the resistivity structure of the subsurface is somewhat complex, signalling that 3-D inversion modelling was the most appropriate technique for producing a resistivity model for this location. A series of 1-D and 3-D inversion models of the resistivity structure at Resolute Bay were generated, again using the OCCAM and ModEM inversion algorithms (Constable et al., 1987; Kelbert et al., 2012). Again, knowledge of the resistivity of frozen and unfrozen materials was used to interpret these models. At Resolute Bay, the model was characterized by three layers: (1) RB1, a high resistivity limestone layer containing partially frozen saline pore water (2) RB2, a low resistivity layer containing unfrozen saline pore fluids, and (3) RB3, a deeper high resistivity layer containing low salinity pore water. The resistivity structure of RB1 matches that which would be observed of a layer of continuous permafrost, the thickness of which corresponds to a lower geothermal gradient than predicted by Misener (1955). A low geothermal gradient at Resolute Bay would restrict the types of geothermal systems that could be implemented in the community, however, estimates of elevated porosity values at depth suggest that there may be reservoir potential for direct use geothermal applications. Additional data is required to confirm the results obtained from the MT survey.

1.6. Thesis overview

The following chapters will explore how geothermal energy resources can be utilized in Canada, and Nunavut in particular. Next, certain methods that can be used to assess geothermal potential in the Arctic will be explored. Electromagnetic geophysics provide a powerful tool for the exploration of geothermal resources, however, knowledge of the electrical properties of rocks is essential for the interpretation of data collected using these techniques. Regions with significant permafrost near the surface can display unique thermal profiles, and can pose challenges for the interpretation of electromagnetic data and the assessment of geothermal resources. Finally, the MT data collected at the communities of Cambridge Bay and Resolute Bay in Nunavut will be discussed, as well as the methods used to interpret these datasets in the context of geothermal energy resources. The overlying structure of this thesis is provided below:

Chapter 2 provides background on the types of geothermal energy resources that exist, their uses, and how these resources can be identified. A brief outline of the geothermal exploration process is discussed, including some geological and geophysical techniques that are commonly used. Finally, this chapter will explore the current state of geothermal development in Canada, including some of the major challenges.

Chapter 3 introduces the concept of electrical resistivity, which is fundamental to the exploration methods described in later chapters. The mechanisms of electrical conduction in rocks, and how these properties vary amongst geological materials will be emphasized. Notably, the electrical behaviour of rocks changes drastically when they are frozen. An overview of permafrost structure and the frozen state of shallow rocks in Arctic environments is also presented.

Chapter 4 compares several electrical and electromagnetic geophysical methods that are commonly used for investigations below the Earth's surface. An emphasis is placed on magnetotellurics, which provides the primary method of investigation in Chapters 5 and 6.

Chapter 5 describes the geothermal investigation at Cambridge Bay, Nunavut using magnetotelluric data. This chapter includes a brief summary of the geological conditions at Cambridge Bay, an overview of data collection, processing, and a discussion of the 1-D and 3-D modelling. These results are interpreted and discussed in the contexts of geothermal resource development and permafrost.

Chapter 6 describes the geothermal investigation at Resolute Bay, Nunavut, similarly to the process completed for Cambridge Bay. Again, a summary of the data collection, processing, modelling, and interpretation is discussed.

Chapter 7 expands upon the investigation at Cambridge Bay, using additional numerical modelling of salt-water diffusion to assess the interpretations of the MT models. A number of numerical simulations will be presented and the results compared with conclusions obtained through MT modelling.

Finally, Chapter 8 will serve to conclude this discussion, providing a concise summary of the topics stated above, as well as potential for future study.

Chapter 2: Geothermal Energy Resources

2.1. Introduction

Heat energy contained in the interior of the Earth is known as geothermal energy. This heat originates from the decay of radioactive elements contained in rocks, and heat remaining from the formation of the Earth several billion years ago. While the total heat content of the Earth is enormous relative to global energy needs, the vast majority of this energy is located at depths that are too deep for extraction to be economically feasible. Nevertheless, geothermal energy resources provide a promising avenue for energy development amid a growing push towards alternative energy sources. The key challenge to adopting geothermal energy is locating where the heat energy is close enough to the surface to be extracted economically with existing technology.

The development of geothermal infrastructure requires the discovery of attractive geothermal resources that are cost-effective to develop, and generally requires a multi-faceted approach to exploration (Hickson & Yehia, 2014; Dobson, 2016). The rate at which the temperature increases with depth in the Earth is called the geothermal gradient, and is dependent on the flow of heat energy from below. Most economic geothermal resources are associated with regions of high heat flow and elevated geothermal gradients, making the heat energy more accessible. Geothermal resources are generally located using a combination of geological and geophysical methods. Drilling is then used to confirm the resource, with a focus on determining the temperature, porosity and permeability. These exploration methods each have their own strengths and weaknesses, which must be assessed in order to efficiently explore for geothermal resources, as is the case in the exploration of other resources.

Geothermal heat originates in two primary sources: the decay of radioactive elements within the Earth, and residual heat that remains from the planet's formation (Barbier, 2002). The deeper parts of the Earth are much hotter than regions near the surface. For instance, regions near the Mohorovičić discontinuity generally attain temperatures far less than 1000°C (Jain, 2014; Schutt et al., 2018), whereas temperatures at the deeper lithosphere-asthenosphere boundary (LAB) can exceed 1300°C (Anderson, 1981). This distribution of heat energy in the Earth allows for heat transfer from the hot interior towards the cooler exterior by three processes: conduction,

convection, and radiation. These processes result in a geothermal gradient of temperature that increases with depth.

In the shallow Earth, heat transfer is dominated by conduction, where energy is transferred between neighboring materials due to a difference in temperature. The rate of conductive heat transfer is largely dependant on the thermal conductivity of the materials that are conducting the heat. Rock is a relatively good thermal insulator, making the conduction of heat through the Earth a very slow process. This conductive regime is relatively inefficient at distributing energy and dominates from the Earth's surface down to depths of 100 - 200 km (Morgan, 1989).

At greater depths, heat flow is primarily driven by more efficient convective and advective processes, and dominate in the rheologically distinct mantle. Heating of deeper mantle material causes convection due to variations in density. In this thermal regime, the bulk movement of material primarily transports the heat energy, which is far more efficient than conductive heat transfer. These mechanisms are responsible for the formation of a thermal boundary layer at the LAB, and lower geothermal gradients in the mantle (Figure 2.1) (Artemieva, 2006). Convection in the Earth's mantle is also responsible for the movement of tectonic plates at the Earth's surface, which has ultimately shaped the surface of the Earth that is seen today.

The geothermal gradient in the Earth is approximately 25 – 30 °C (Paixach, et al., 2025) on average, but varies significantly with depth, and laterally due to the local geology. Heat flow also varies across the Earth's surface, with average continental crust heat flows of 65 mW/m², and average oceanic crust heat flows of 101 mW/m² (Pollack, et al., 1993). Variations in surface heat flow are largely due to lateral changes in subsurface characteristics. The types of rock, fracturing, and water content are generally related to regional plate tectonics and strongly control the subsurface thermal regime (Paixach, et al., 2025). The presence of elevated concentrations of radioactive isotopes in local rocks can elevate heat flow and is therefore another factor affecting the spatial variation in geothermal heat flow and gradients (Jones & Majorowicz, 1987).

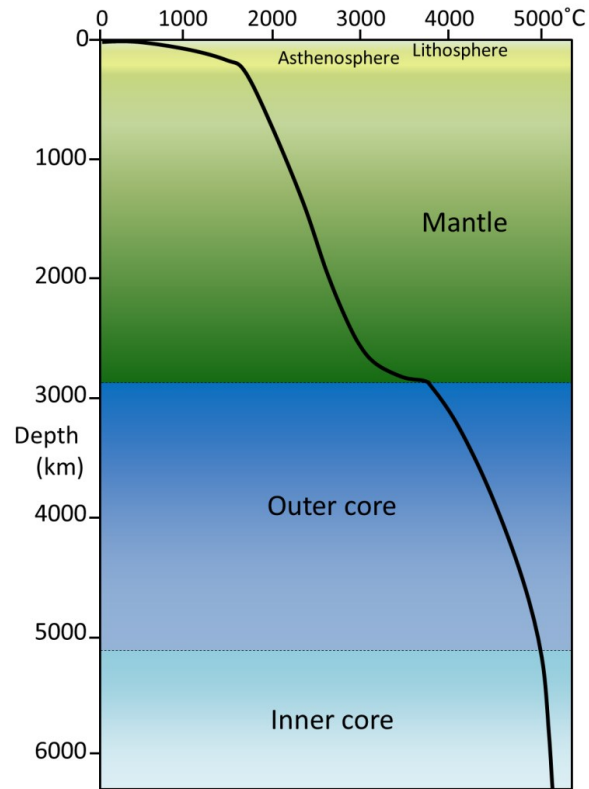


Figure 2.1: A diagram showing a generalized temperature profile from the Earth's surface to its core. The geothermal gradient is variable with depth throughout the Earth. From Earle (2015).

With the mean global heat flow of 82 mW/m^2 , Barbier (2002) estimates the global heat loss at the surface to be greater than 40 TW, or $1.26 \times 10^{26} \text{ J}$ in one year. This is greater than the total global energy consumption in 2017 (Ahmad & Zhang, 2020), showing that geothermal heat flow is a sizeable source of energy on Earth. While geothermal systems can be challenging and costly to install, they provide a constant source of energy, independent of the weather, and with a relatively small environmental impact (Paixach, et al., 2025). Global geothermal energy capacity has exceeded 16 GW and continues to supply a growing proportion of the world's energy (Gutiérrez-Negrín, 2024; Paixach, et al., 2025). The rate of implementation of geothermal energy is expected to increase with the global push to move away from fossil fuels. Although geothermal energy is present across the Earth's surface, it is generally only economic to extract it in areas where the heat flow is high.

Geothermal energy can be exploited for various purposes including space heating, electricity generation, and a wide range of industrial applications (Lund et al., 2005; Lund et al., 2022). When

subsurface temperatures are high enough ($> 100 - 150^{\circ}\text{C}$) it can be efficient to generate electricity. Geothermally generated electricity is useful as it is not dependant on the time of day or surface conditions, and can be transported for use elsewhere by power lines or batteries. Electricity generation using geothermal energy typically requires high reservoir temperatures, which may be found at several kilometres depth in regions with relatively low geothermal gradients.

Where geothermal temperatures are not high enough, or are inefficient for electricity production, heat capture for direct use applications may be possible. Some direct use applications include electric heat pumps, district heating, agriculture and aquaculture (Lund et al., 2022). Direct use applications generally require lower geothermal temperatures and can be implemented in a greater number of regions. Unfortunately, geothermal energy that is stored in subsurface fluids cannot be easily transported, and must be used locally.

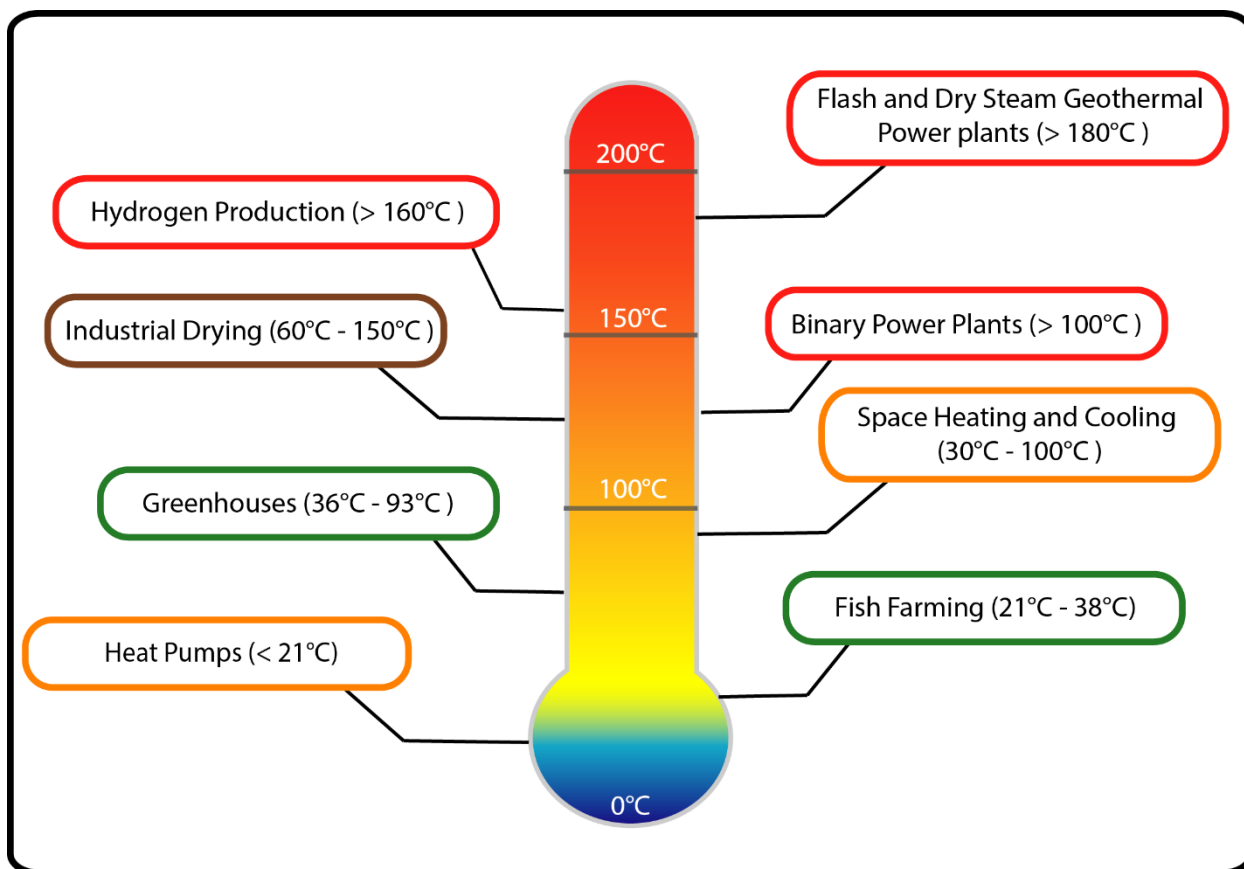


Figure 2.2: A diagram showing a range of uses for geothermal energy, adapted from (U.S. Department of Energy, 2019). A range of geothermal temperatures is available for direct use scenarios, and for conventional geothermal power generation.

On a local scale, three basic requirements are needed for the efficient extraction of geothermal energy: these are (1) heat, (2) a fluid to transport the heat, and (3) permeability to allow fluid to flow. Heat energy constitutes the resource to be extracted and is the most important variable in geothermal exploration. A working fluid, often water in the form of a brine, is required to act as a body for transport of the heat to the surface. Finally, a reservoir is required to store and circulate the working fluid beneath the surface to absorb the heat. This reservoir is usually a body of rock with enhanced permeability, capable of storing the fluid. In the absence of a natural working fluid or reservoir, engineered alternatives can be used at higher costs. Figure 2.3 shows an example of how geothermal fluid can be circulated through the Earth to extract heat energy.

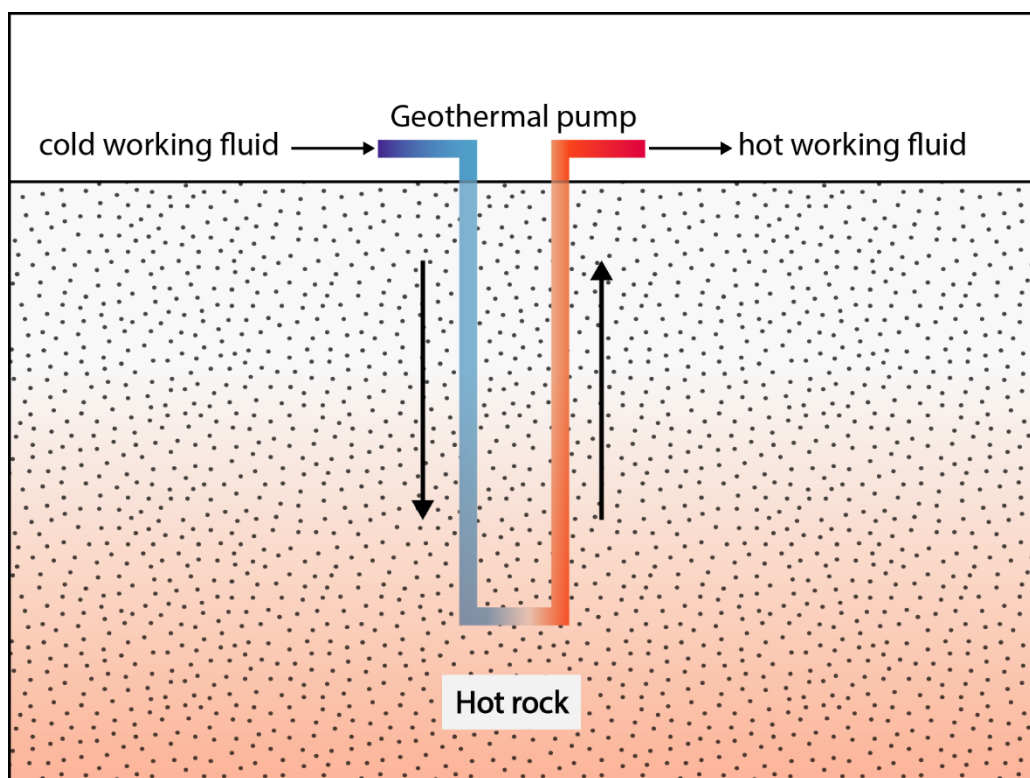


Figure 2.3: A simple illustration of a geothermal system to describe how geothermal energy can be extracted from rocks below the Earth's surface. A working fluid is circulated through a loop, where it can be heated from hot rocks below the surface. Once pumped to the surface, the heat energy can be extracted from the working fluid for a variety of purposes. The fluid can be stored directly in a reservoir to collect energy, or circulated through a closed-loop system.

Identification of the presence of heat, porosity, and permeability is often the primary step in geothermal exploration and can be accomplished using surface feature identification, geology and

geochemical analysis, and geophysical exploration. Direct drilling can be used to directly determine the geothermal feasibility of a location, but can be prohibitively expensive and is often used to confirm the validity of other results. Ultimately, geothermal resources are most often characterized by their temperature, with high-temperature ($T > 150\text{ }^{\circ}\text{C}$) and medium-temperature ($80\text{ }^{\circ}\text{C} < T < 150\text{ }^{\circ}\text{C}$) capable of generating electricity, and low-temperature ($T < 80\text{ }^{\circ}\text{C}$) potentially useful for other direct use systems (Grasby, et al., 2012).

2.2. Types of geothermal systems

Geothermal systems that store heat energy in subsurface rock and fluids within the Earth can be categorized in different ways. The following categorizations are commonly used: (1) magmatic and (2) non-magmatic geothermal systems, (3) engineered geothermal systems (EGS), and (4) supercritical geothermal systems (Muñoz, 2014). Magmatic geothermal systems are characterized by the presence of magma within the system and include convective hydrothermal systems, hot dry rock, and partial melt systems (Muñoz, 2014). Non-magmatic systems typically involve hot fluids that are entrained in geothermal reservoirs such as porous rock or fault systems. EGS systems are a third man made system type that involves artificial fracturing of impermeable rocks to construct a reservoir in regions that would otherwise be unsuitable for the circulation of geothermal fluids. The distribution of these systems is dependent on local geology, tectonism, and the depths at which they can be found and are sometimes suitable for different uses.

Magma derived hydrothermal systems are defined by the emplacement of melted or partially melted rocks in subsurface regions of high temperature. These systems are most often related to geological expressions of plate tectonics such as magmatic arcs and volcanism where subsurface temperatures are high enough to melt rock. Such systems are of the most commonly utilized geothermal systems due to their obvious surface expressions and high geothermal temperatures. Typically these systems are composed of a magmatically derived heat source, a reservoir of geothermal fluids with high porosity and permeability, and an impermeable barrier to immobilize the fluids – most commonly a clay cap (Muñoz, 2014). These geothermal fluids often contain significant amounts of dissolved salts.

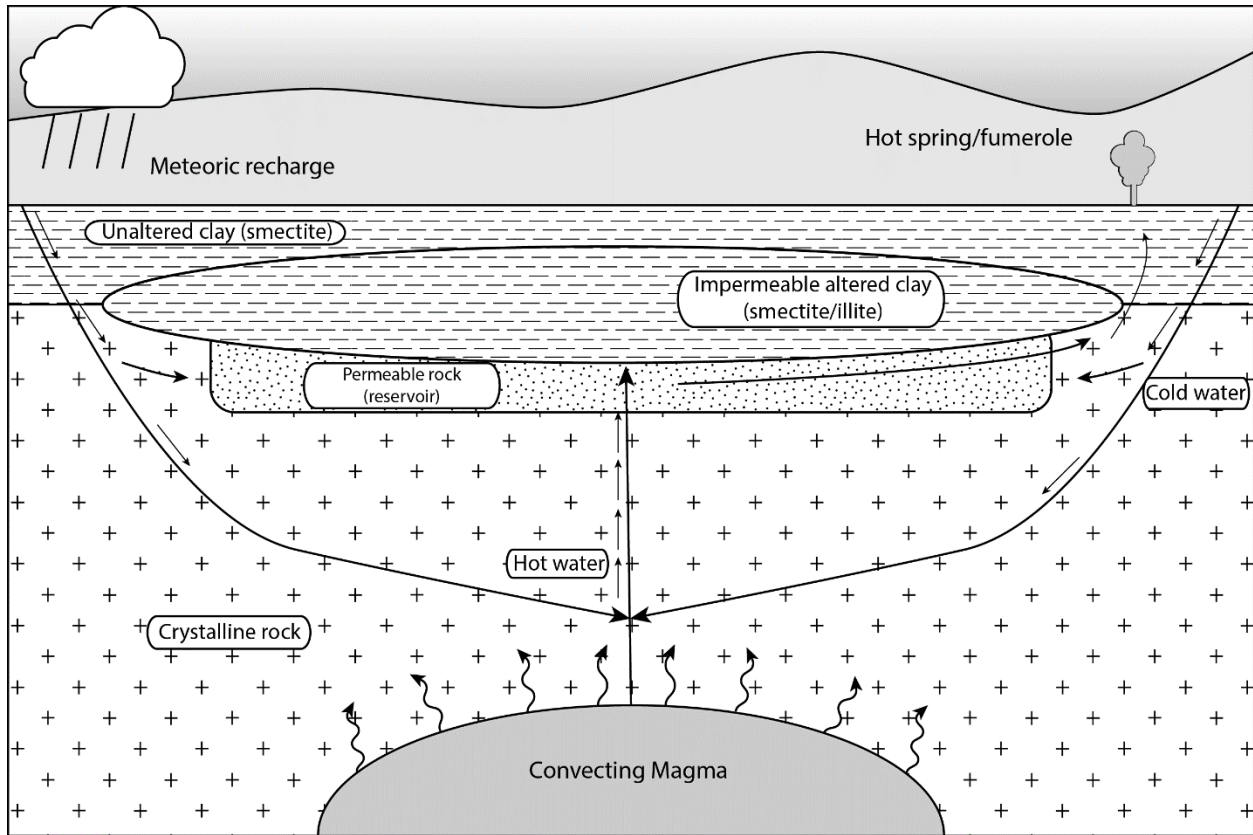


Figure 2.4: A schematic of a magmatic hydrothermal system. Heightened temperatures produce a characteristic impermeable altered clay cap, which overlies the permeable reservoir. Fluid flows through the host rock via fault and fracture pathways, and is recharged by subsurface and surface sources. Adapted from White (1967).

Clay minerals are sensitive to changing temperatures and as such a variety of altered clay minerals may be found across the temperature gradient of a hydrothermal system. Lower temperature smectite exist at lower temperatures (70 - 150°C) and are typical of the overlying cap (Hersir et al., 2013; Muñoz, 2013). Other mineral species, illite, chlorite, and epidote minerals, are formed at greater temperatures. The clay cap and geothermal fluids comprise regions that have low electrical resistivity, whereas the deeper clays are more resistive. These differences in electrical properties allow for successful imaging of hydrothermal systems using geophysical methods that measure resistivity (Hersir, et al., 2013; Hanneson & Unsworth, 2023) and will be described in more detail in Chapter 3. Magmatic systems often present temperatures that are suitable for high-enthalpy geothermal systems, such as those that produce power.

Non-magmatic systems are those where heat is not derived directly from molten rock, but instead occur generally in regions where hot saline fluids are entrained in rocks, whether in subsurface aquifers, or along extensive fracturing or faulting. These systems may occur in regions with tectonic-related faulting, but do not necessarily have significantly elevated thermal gradients. The overall structure of non-magmatic systems can occur in a wide range of regimes and so can vary significantly in size and shape. Because of this, a single conceptual model for these types of systems cannot be used. Fluids in non-magmatic systems often exhibit heightened salinity, with dissolved species being sourced from the dissolution of rocks beneath the surface and surface or meteoric waters connected to the fault or fracture system (Moeck, 2014). In coastal regions, seawater can infiltrate subsurface reservoirs through connected pathways or permeable rock (Barlow, 2010). Saline geothermal fluids entrained in fault systems tend to have low-resistivity compared to the surrounding host rock, creating an ideal target for certain geophysical methods. (Ucok et al., 1980).

Engineered geothermal systems gained traction in the early 1970's as a means of adapting geothermal energy systems to regions that only contain hot, dry rock. In this context, dry rocks are those that do not contain sizeable fluid or steam content, and are thus not useful for conventional heat extraction. This method involves the hydraulic fracturing of subsurface to artificially construct a subsurface reservoir for the circulation of geothermal fluids (Potter et al., 1974). By circulating fluid through the region of enhanced fracturing using two separate geothermal wells, heat energy can be extracted from an otherwise unusable geothermal source. Since the inception of EGS systems, dozens of such systems have been implemented worldwide (Breede et al., 2013). While EGS technologies can be developed in a greater range of locations, their uses are offset by the heightened costs of drilling requirements and the costs of reservoir stimulation. As EGS is typically employed in regions with lower geothermal viability, deeper drilling is often required to reach temperatures necessary for economic recovery of energy. Continued advances in drilling technology and EGS lower costs and provide incremental improvements to the efficacy of adopting EGS systems (Tester et al., 2007). Closed loop technologies are also undergoing development, where geothermal fluids are cycled through a two well system without exiting the system.

Supercritical geothermal systems are extremely high-enthalpy systems that involve extremely high temperatures, useful for electricity production. These systems can be characterized by water stored

with high energy due to particularly high temperatures and pressures (Reinsch et al., 2017). These systems are capable of providing the extremely high temperatures for electricity production, but are generally restricted to regions of unique tectonic or volcanic structure. Greater efficiency per well for supercritical systems could lead to greater energy output with smaller footprints. The high temperatures can make drilling for these systems particularly challenging.

2.3. Geothermal exploration

The modern exploration process for geothermal energy resources can be challenging, and typically involves a multi-pronged approach to identify economically viable targets. As with most other resources under the Earth's surface, geothermal development involves a significant amount of financial risk. This risk that can be mitigated with well-planned exploration. Currently, an assortment of tools are available for locating and assessing geothermal resources, including geology, geochemistry, geophysics, and drilling (Hickson & Yehia, 2014). Typically, an exploration program begins with the least expensive (low-risk) techniques, and more expensive techniques are applied as information is gathered and uncertainty about the resource decreases. This is applied in practice through the broad characterization of geothermal systems using geological and geochemical observations at the surface, followed by model refinement using a variety of geophysical methods. Finally, the geothermal model can be tested through deep drilling and well analysis. Complete geothermal projects involve both the exploration of geothermal resources, and their exploitation. This section will focus on finding and assessing these resources.

Exploration in regions that lack prior extensive information often begin with large-scale studies that can cover several hundred square kilometres (Wollenberg, 1982). In many cases, active geothermal manifestations are found at the surface. These include hot springs, sulphurous odors, and precipitates, can signal the presence of a connected geothermal pathway to the surface. Oftentimes, primary reservoirs are not coincident with their surface expressions and additional study must be done to ascertain the location and geometry of the system in the subsurface.

Once a potential target has been identified, surficial geological methods are used to gather the information needed to infer properties of the subsurface. Mapping of geology and study of rocks at the surface help provide a clearer picture on how the local geology, and by extension geothermal systems, have evolved. Identification of mineralization that is associated with hydrothermal systems can be used to verify reserves and study the composition. Petrological observations may

provide information about the geologic history or help to locate regions where previous geothermal have flowed. As previously discussed, the chemical constituents present in geothermal rock and dissolved in fluids provides information about their provenance. In using these methods, information about the reservoir lithology can be ascertained, including information about the cap rock and reservoir. At this point, information of the geothermal system is collated to form a preliminary conceptual model of the system. Possible targets might be chosen, but are likely not acted upon at this stage.

Once a preliminary model has been developed and the nature of the resource is better constrained, additional, more reliable and expensive methods can be used to refine the model. Geophysical data are widely used in geothermal exploration as a way to indirectly gather more precise information about the properties of a geothermal system, such as the depth and geometry, extent of geothermal activity, temperature, and permeability. Seismic methods can be used to detect and model important features of geothermal areas and fault systems (Hayakawa, 1970). Gravimetric methods can map regions of varying rock type, temperature, and porosity. As well, precise gravity measurements may also be implemented over time to monitor fluid reservoir flow and recharge during the later stages of geothermal production (Sugihara & Ishido, 2008). Electrical and electromagnetic (EM) methods are capable of producing resistivity models of geothermal systems that can be used to estimate reservoir temperatures, porosity, permeability, and volume (Hersir et al., 2022; Hanneson & Unsworth, 2023). The types of electrical and EM techniques, and their applications to geothermal exploration will be described in more detail in Chapter 4. Geophysical data can be quite expensive to collect. However, geophysical survey costs are only a fraction of the total exploration, drilling, development, and operating costs of geothermal wells and significantly reduce the risk of drilling in non-viable locations (Hickson & Yehia, 2014). For these reasons, geophysical exploration is an integral part of geothermal exploration.

The exploration techniques discussed thus far are often integral to the construction of an accurate conceptual model. To further assess the geothermal resources and verify the current working model, wells must be drilled. Drilling wells is the most reliable method for geothermal assessment, and is also the most costly by far. In fact, production drilling and well testing can comprise 50% or more of total geothermal exploration costs (Stefánsson, 2002). Estimates made by geological, geochemical, and geophysical evidence can be confirmed from well logging and flow testing at

well sites. If preliminary exploration does not yield promising results, site development may be abandoned in favour of other prospects. The enormous costs of developing geothermal energy highlight the risks involved and the importance of using a thorough, mixed approach to exploration to reduce uncertainty in projects of considerable cost.

In recent years, Canada has shown renewed interest in developing geothermal energy resources in the search for alternative energy sources. It may now be clear that a variety of techniques must be used to ensure that funding towards exploring geothermal feasibility is used effectively. Next, the past and present opportunities for geothermal in Canada are discussed.

2.4. Geothermal energy in Canada

The geothermal resource potential of Canada was largely unknown before 1973. Uncertainty during the energy crises of the 1970s spurred rapid exploration into alternative sources of energy to petroleum in Canada, and elsewhere in the world. Currently, there is no electricity being directly produced with geothermal energy sources in Canada; however, some smaller scale projects are ongoing, such as the Swan hills hybrid gas-geothermal power plant, and geothermal heating from abandoned mines in Springhill, N.S. (Jessop et al., 1995; Grasby et al., 2012).

The geology of Canada is varied, with several distinct geologic regions. The Canadian Shield is composed of exposed ancient igneous and high-grade metamorphic rocks of Precambrian origin. The absence of active tectonic activity and elevated rock porosity, along with low geothermal gradients means that this entire region is generally poorly suited for geothermal development (Jessop, et al., 1991). In the west, the geology reflects the dynamic orogenic history of the Canadian Cordillera and volcanic belts. The Canadian Cordillera region extends through British Columbia and Yukon and into the western regions of Alberta and the Mackenzie Mountains regions of the Northwest Territories. This region boasts the highest geothermal gradients in the country and is considered to host the most economically feasible geothermal prospects. The Mount Meager Volcanic Complex has shown some promise over the years for successful geothermal development (Grasby et al., 2012). High geothermal potential in British Columbia is related to the Western Canadian Volcanic Belt and other extrusive features derived from subduction along the country's western margin.

In addition to traditional geothermal plays used to generate electricity using high-grade thermal resources, developments have been made in utilizing lower heat geothermal resources. Significant research has been done into the feasibility of geothermal in sedimentary basins. While Canada hosts a number of sedimentary basins, knowledge of these regions varies greatly; the Western Canadian Sedimentary Basin (WCSB) is perhaps the most well studied of these environments. The WCSB lies between the Canadian Cordillera and Canadian Shield regions and consists of a wedge of layered sediments of Pre-Cambrian – Quaternary sedimentary units. Significant oil and gas exploration in this region has led to an extensive record of the geology throughout this region, including widespread logging of temperature and permeability. These regions have been shown to contain sizeable fluid content, with the majority being water. Typical temperatures of sedimentary basins are on the lower end of geothermal resources due to the relatively low thermal conductivity of sedimentary rocks; however, higher radiogenic heat in sedimentary basins produces greater heat flow than the shield. (Jones & Majorowicz, 1987). Sedimentary rocks of the WCSB extend to maximum depths of approximately 7 km, and cover 1.4 million km² (Mackay & Pedersen, 2022), potentially providing suitable conditions for heat extraction. Larger population density within the WCSB region also makes it a more attractive option for geothermal development as resources can be developed more easily near large population centers.

Geothermal data in the Territories is often scarce, but temperature gradients are generally low (Minnick, et al., 2018). Evidence of high geothermal gradients in isolated locations has been presented; however, connecting these resources to consumers is challenging due to isolated supply lines and low population density. Geothermal energy has some attractive qualities in the North; existing diesel power production is subject to logistical challenges, and communities are often isolated from new supply for several months of the year, while geothermal energy is capable of providing reliable heating or electricity production. Additionally, geothermal production is unaffected at times when there is no wind, or little daily sunlight.

The challenge of developing geothermal energy in the north is further complicated by the fact that the costs of drilling and installing geothermal infrastructure are considerably higher. Supply and transportation costs increase substantially for northern projects and permafrost increases the complexity and costs of drilling and construction for geothermal installations. While the up-front cost is substantial, many of these costs can be recovered over time, and there are a multitude of

social benefits to bringing alternative energy to the Arctic. (Kinney et al., 2019) concludes that there are financial, health, and climate advantages of using direct-use geothermal technology to support sustainable food production in Northern communities. There is some precedence to consider geothermal energy as a cost effective alternative in the Canadian North, despite the unique challenges (Majorowicz & Grasby, 2014).

The following chapters will set the stage for geophysical exploration of geothermal resources in the far north. An understanding of how continuous permafrost affects the study of geothermal is needed. Then, an overview of electrical and electromagnetic methods that can be used for exploration in the Arctic will be explored.

Chapter 3: Electrical Properties of Frozen and Unfrozen Rocks

3.1. Introduction

When geophysical methods are used to study the subsurface, it is paramount to understand the physical properties that are measured. Electrical and electromagnetic geophysical methods are primarily used to obtain information about the electrical resistivity, a property that describes the ability of a material to resist the flow of electric charge. Mathematically, the electrical resistivity is the reciprocal of the electrical conductivity. Electrical resistivity is widely used in the study of geothermal systems and permafrost, since it is sensitive to the features that are often characteristic to these systems, such as fluid-filled reservoirs or frozen rocks. This chapter will provide an overview on the property of resistivity, including how it varies in frozen and unfrozen conditions. Then, the structure of permafrost will be discussed to provide a background for using electromagnetic methods for geothermal exploration in regions with permafrost.

The propagation of electromagnetic (EM) fields in the Earth depends on the material properties of electrical conductivity (or resistivity), magnetic susceptibility, and dielectric permittivity. The magnetic susceptibility of a material describes how it responds to an applied magnetic field. Geologically, magnetic susceptibility varies in different rock species to a degree where differences may be detectable if they are large, such as at large ore deposits, or in electrically resistive settings. Dielectric permittivity of a material describes the electrical polarizability of a material under an external electric field. This property is generally most important for EM systems that operate at sufficiently high frequencies, such as ground penetrating radar. For most EM techniques, electrical conductivity is the most important property.

3.2. Electrical resistivity of Earth materials

In studying the electrical resistivity of rocks, it is very important to understand the mechanisms through which materials are able to conduct electricity. A variety of mechanisms exist, some of which are more effective than others. Through these differences, the ability of Earth materials to conduct electricity varies significantly.

Here, materials with only a single phase will be considered first, because of their simplicity. Then, the behaviour can be extended to multi-phase systems, which are more common in the Earth.

3.2.1. Electrical resistivity

Electrical conductivity is a material property that describes how well a material allows electric current to flow through it. Ohm's Law describes the resistance of a dimensionless resistor, but can be reformulated to describe current flow through a continuous medium as:

$$\mathbf{E} = \sigma \mathbf{J} \quad (3.1)$$

where $\mathbf{E}$ is the electrical field in V/m, $\mathbf{J}$ is the current density in A/m², and σ is the electrical conductivity, given in S/m. The electrical resistivity, ρ , is the reciprocal of σ , such that

$$\rho = \frac{1}{\sigma} = \frac{\mathbf{J}}{\mathbf{E}} \quad (3.2)$$

and is measured in Ωm . Both conductivity and resistivity are commonly used, depending on the topic.

Electric current flows through a continuous medium under the effects of an applied electric field. The current is facilitated by the movement of charged particles, such as electrons and ions. The mobility of a charge carrier is defined as the drift velocity divided by the electric field, $\mu = v/E$. The mobility describes how easily the charge carrier can move through the medium. Using μ and Eq. 3.2, an alternative description of the electrical resistivity is given by

$$\rho = \frac{1}{nq\mu} \quad (3.3)$$

where n is the number of charged particles and q is the charge of the particles, showing that a material with many mobile charge carriers will have a low resistivity. This result will extend to geophysical exploration, where the presence of mobile charge carriers in geological materials will produce structures with characteristically low resistivity.

3.2.2. Conductivity in single-phase systems

The simplicity of a system consisting of only a single material is a natural place to start discussing the conductivity of rock. Electrical conduction can occur through several mechanisms: electronic conduction, semi-conduction, and through solid or aqueous electrolytes. The overall conductivity

of a material depends not only on the modes of conduction, but also properties such as temperature and pressure (Roy, 2020). The resistivity values of a number of materials are shown in Table 3.1, illustrating the wide range that exists.

Electronic conduction occurs commonly in metals and metallic materials where plentiful valence electrons act as efficient charge carriers. Some resistance arises from crystal defects and imperfections, however, despite these effects, most metals are good conductors. Pure, elemental metals such as copper (5.81×10^7 S/m), gold (4.10×10^7 S/m), and silver (6.80×10^7 S/m) are among the most conductive naturally occurring materials and consequently are valuable for a wide range of uses in machinery, and modern electronics and electrical distribution.

Materials that are semi-conductors also have electrons as charge carriers. However, the number of free electrons is much lower than in metals, since the energy required to raise electrons from the valence band to the conduction band in these materials is much greater than in metals. This conduction mechanism is strongly controlled by temperature, and as such the conductivity has a high temperature dependence (Roy, 2020). Many common geologic minerals behave as semi-conductors, and so the temperature dependant behaviour plays an important role in the behaviour of these minerals in different contexts.

Electrolytic materials are those that conduct electricity through the movement of ions. A clear difference between the conductivity of seawater and water without a significant number of dissolved ions (Table 3.1) demonstrates the role of the charge carrier in enabling electrical conduction. Saline water is a commonly found electrolytic conductor that is often found as a saturating fluid in geologic settings. Conductivity of electrolytes is also dependent on factors such as temperature and pressure. The conductivity of seawater on the Earth varies according to these properties, as well as the concentration of dissolved ions (Tyler et al, 2017). Conductivity of salt water is approximately proportional to the dissolved salt concentration (Schlumberger, 2013). Electrolytes, both in liquid and solid states are essential components in batteries, functioning as an important medium for ion transport.

Material	Type of Conductor	Conductivity at 20°C (S/m)	Resistivity at 20°C (Ωm)	Reference
Silver	Conductor	6.80×10^7	1.47×10^{-8}	Freedman & Young (2016)
Copper	Conductor	5.81×10^7	1.72×10^{-8}	Freedman & Young (2016)
Gold	Conductor	4.10×10^7	2.44×10^{-8}	Freedman & Young (2016)
Silicon	Semi-Conductor	4.35×10^{-4}	2300	Freedman & Young (2016)
Germanium	Semi-Conductor	1.67	0.60	Freedman & Young (2016)
Seawater	Electrolyte	~ 4.00	~ 0.25	Schlumberger (2013)
Pure water	Insulator	4.00×10^{-6}	2.5×10^5	Light et al. (2004)
Graphite (parallel to basal plane)	Semi-Conductor	3.33×10^2	3×10^{-3}	Pierson (1993)
Graphite (perpendicular to basal plane)	Insulator	$2.0 - 4.0 \times 10^5$	$2.5 - 5.0 \times 10^{-6}$	Pierson (1993)
Mica	Insulator	$10^{-15} - 10^{-11}$	$10^{11} - 10^{15}$	Freedman & Young (2016)
Sulfur	Insulator	10^{-15}	10^{15}	Freedman & Young (2016)
Quartz (fused)	Insulator	1.33×10^{-18}	75×10^{16}	Freedman & Young (2016)

Table 3.1: *Electrical conductivity and resistivity of a variety of materials at 20°C. Resistivity values can differ by many orders of magnitude between different types of materials.*

The concept of electrical anisotropy is demonstrated by the significantly different values in electrical conductivity for graphite (Table 3.1). The sheet-like layered structure of graphite strongly controls its electrical characteristics, leading to vastly different electrical properties depending on the orientation of the material. This behaviour can exist in geological settings, and may be the result of rock composition or rock fabric.

3.2.3. Resistivity in multi-phase systems

Most materials found in the Earth are a mixture of multiple phases of materials with different resistivity. Some examples of these mixtures include (1) magma bodies that contain a mixture of molten rock and unmolten grains, and (2) sedimentary rocks that are a mixture of resistive grains with aqueous fluids in their pore space. It is therefore important to consider relationships that can

be used to calculate the bulk electrical properties of mixtures based on the properties of the individual phases. These relationships are called mixing laws.

Knowledge of porosity and permeability is imperative when discussing multi-phase systems that include fluid-saturated rocks, as these parameters strongly control the bulk conductivity of the material. Porosity and permeability provide pathways for charge carrier movement within an otherwise immobile material. The key mechanisms of electrical conduction in both frozen and unfrozen rocks are related to conduction through the pore space. Understanding the relationship between the conductivity of rocks and their porosity is essential to successfully interpreting electromagnetic survey results, as will be shown later. Porosity is the measure of the empty space within a material, typically expressed as a percentage of the total volume:

$$\phi = \frac{V_{void}}{V_{void} + V_{rock}} = \frac{V_{void}}{V_{total}} \quad (3.4)$$

Porosity in rocks occurs due to several factors during deposition and beyond, and can be grouped into primary and secondary porosities. The primary porosity of a rock is the porosity of the rock at the time of formation, and is largely influenced by the makeup and shape of individual grains and how they were joined together during formation. Secondary porosity arises from processes that occurred after the rock was formed, and includes fractures caused by deformation or voids caused by weathering. For instance, chemical frameworks of interlocking mineral grains can be the remnants of biochemically produced structures (Murray, 1960). Sedimentary rocks generally have the highest porosities, whereas igneous and metamorphic porosities are often much lower due to the tightly interlocking mineral grains formed under much different conditions. Porosity is generally highest in cases where grains are well-sorted, uniform in size and shape, and grains are loosely packed within the matrix. The porosity of all rock types can be enhanced by secondary processes. The study areas explored in Nunavut in Chapters 5 and 6 are characterized by near surface carbonate rocks. Dissolution and frost shattering are examples of two physico-chemical porosity types that can develop in such environments. The effective porosity of a rock evolves from the combined effects of primary and secondary porosity, and can be the result of a complex series of processes over a rock's existence.

Permeability is a property that describes how easily fluid can flow through a material. Porosity and permeability are linked; porosity provides the void spaces that permits fluid flow, however,

greater permeability requires that the pore spaces be interconnected to provide direct fluid pathways. Even a connected network of pore space can inhibit material permeability if the channels for fluid flow are meandering through the material. Tortuosity is the quality of fluid pathways that are winding or irregular, instead of straight, impeding fluid flow. Overall permeability in rocks is primarily related to the porosity, but may be enhanced or diminished by factors such as composition, particle size, and the occurrence of fracturing or dissolution.

Archie's Law (Archie, 1942) is an empirical mixing law for two-phase systems (i.e. a rock with fluid saturated pores) that relates the porosity and saturation of a rock to its bulk electrical resistivity. This relationship was originally developed in the context of hydrocarbon exploration for relating the properties of sedimentary rocks to their degree of saturation of oil, but is also applicable to rocks saturated with other fluids, such as water. Archie's Law is given by:

$$\rho_0 = \rho_w S^{-n} \phi^{-m} \quad (3.5)$$

where ρ_0 is the bulk resistivity of the rock, ρ_w is the resistivity of the saturating fluid (water), S is the degree of saturation of the rock, ϕ is the porosity of the sample, n is the saturation exponent, and m is the cementation exponent.

In this model, the resistivity of the rock grains, ρ_r , is assumed to be much greater than that of the resistivity of either the bulk material or fluid; all conduction is assumed to occur as a result of ion movement through the fluids in pore space. Archie's Law has been successfully used in relating measured resistivity values to rock porosity (Hanneson & Unsworth, 2023). The parameter m is known as the cementation factor and represents the interconnectedness of pores within the rock, modelling the permeability of the rock. The cementation factor for most sedimentary rocks lies in the range of $\sim 1 - 2$, with smaller values modelling a more interconnected pore geometry. The parameter n is the saturation exponent and models the geometry in which the fluid adheres to the pore walls. Typically, this value is close to 1 or 2. Figure 3.1 demonstrates how the interconnectedness of the pore system affects the overall resistivity.

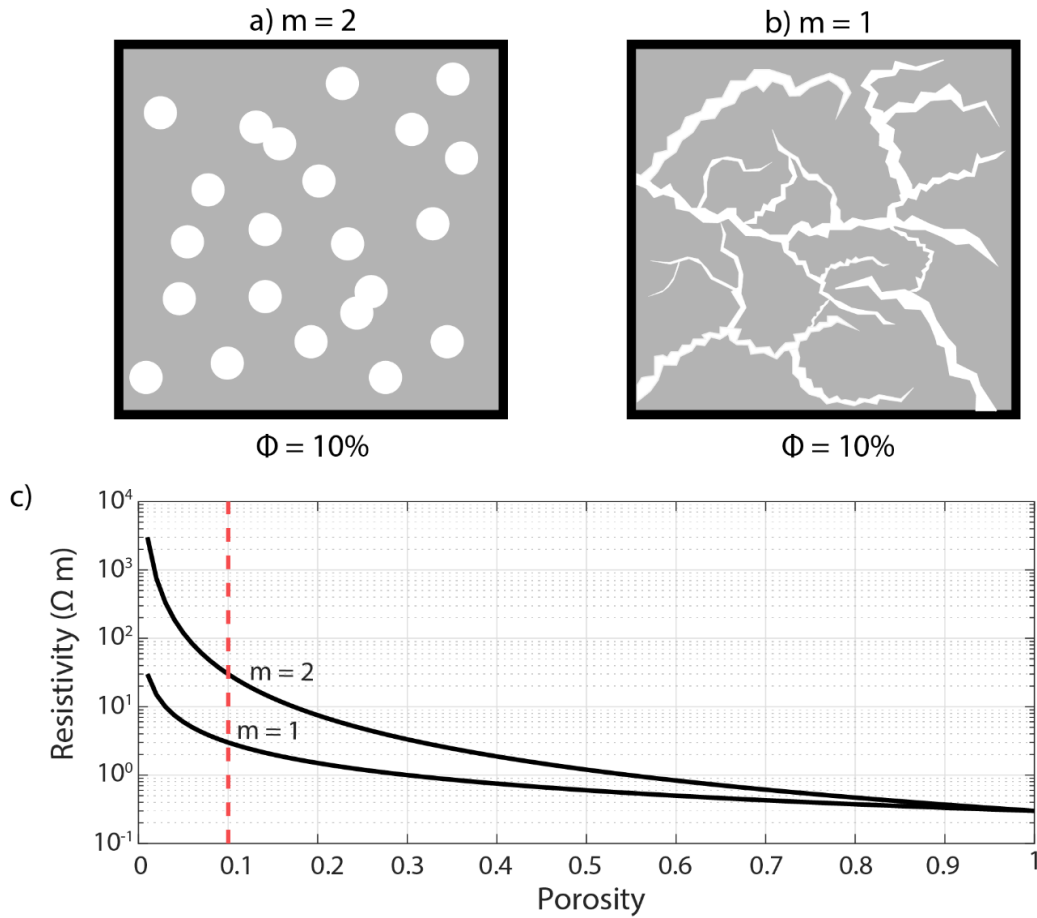


Figure 3.1: a) a physical interpretation of a rock with 10% porosity and $m = 2$, representing spherical pores; b) an interpretation of 10% porosity and $m = 1$, representing elongated pores; c) a plot showing the bulk resistivity described by Archie's Law for $m = 1$ and $m = 2$. The samples are completely saturated ($S = 1$), and $n = 1$. A dotted red line is used to show the estimated of bulk resistivity of the samples in a) and b) under the same conditions.

3.2.4. Resistivity of frozen rocks

Previously, it was shown that the electrical resistivity of most materials varies significantly with temperature. Higher temperatures contribute energy to the material that is required to mobilize charge carriers, lowering the resistivity. Conversely, lowering the temperature generally has the opposite effect, increasing the overall resistivity. In the case of water, which exists nearly ubiquitously across the Earth, near-surface temperatures can commonly reach the freezing point,

which drastically affects conductivity of both surface water and groundwater. This undoubtedly affects the bulk resistivity of water-saturated rocks under freezing conditions. The study of subsurface resistivity in environments with near surface freezing thus requires knowledge of how the resistivity of rocks changes during the freezing process.

In the case of a conductive saline solution, cooling will slowly lower the conductivity of the solution until freezing occurs. Once all the water is frozen, the ions are immobilized and conductivity will decrease dramatically. However, the freezing process does not occur at a single temperature, but instead occurs gradually as more water in the pore space freezes. As the temperature decreases, ice crystals begin to form only containing pure water. Salt is expelled from ice crystals during freezing, raising the salinity and lowering the freezing point of the remaining water. As the overall temperature decreases, the volume fraction of water in the pore space decreases, and its salinity increases. This process is depicted in Figures 3.2 and 3.3.

Experiments showed that a thin layer of unfrozen water remains along the edges of rock grains (Figure 3.2) (King et al, 1988). In sufficiently porous rock, charge carriers along this rock-water-ice interface facilitate electrical conduction in the frozen medium. Due to this process, the electrical conductivity of freezing saturated rocks tends to decrease over a range of temperatures, typically 0 to -10°C (King, et al, 1988; Daniels, et al., 1976; Anderson & Morgenstern, 1973). It is important to note that it is often assumed that frozen rocks containing high ice content will have a high-resistivity (low conductivity), potentially 1000 Ωm or more, depending on the salinity and composition (Pandit & King, 1979). However, if sufficient salt and/or clay is present the resistivity can be 100 Ωm or less (Minsley, et al, 2012; Overduin, et al., 2012; Dafflon, et al, 2013; Pandit & King, 1979). The experimental results of Pandit & King (1979) describe the effects of temperature and salinity on the bulk resistivity of various rocks. These results are used as a cornerstone for analysis and interpretation of magnetotelluric data in chapters 5 and 6.

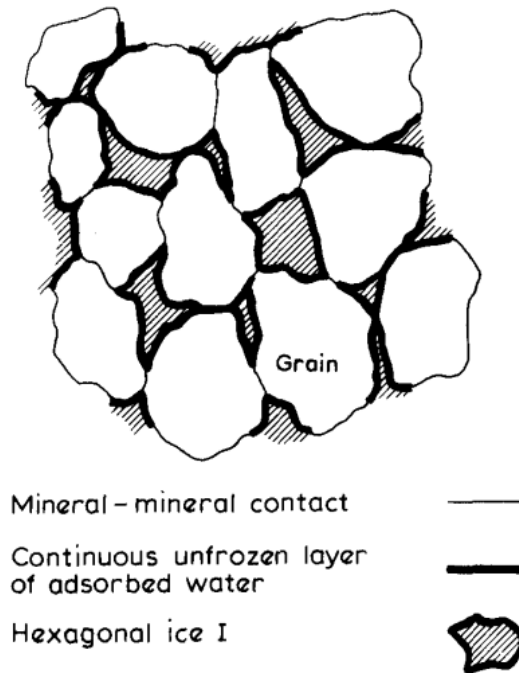


Figure 3.2: A diagram depicting the mineral-water-ice contact and thin continuous water layers in which conduction occurs in frozen rocks. From King et al., (1988).

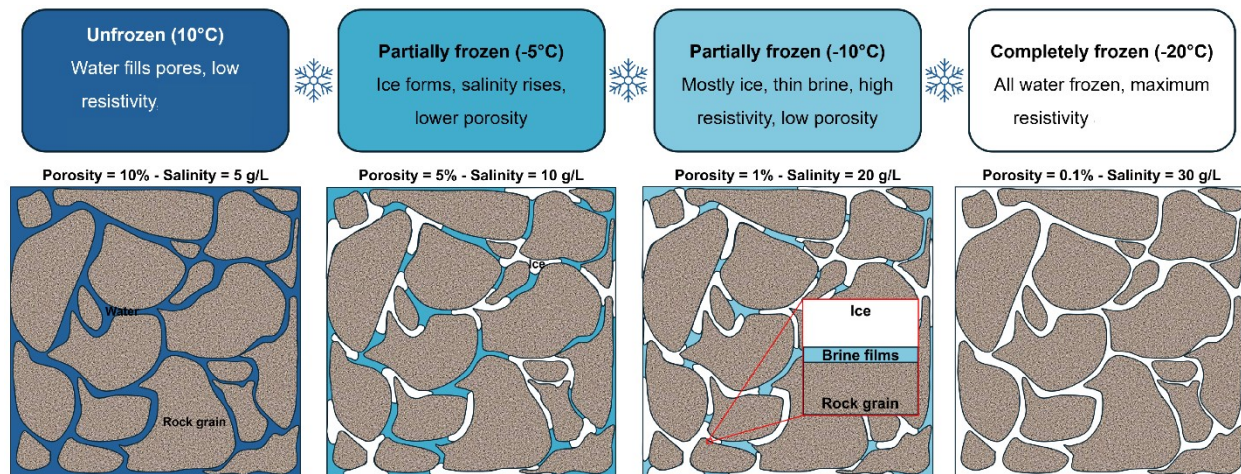


Figure 3.3: Shows the progressive decrease in fluid-filled rock pore space during the freezing process. Ice crystals expel salt into the decreasing water volume, increasing the salinity. Electrical conduction in frozen rocks occurs primarily through saline films along the rock-ice interface. (Figure courtesy of Keytash Moshtaghian).

Electromagnetic methods have been successfully used to map permafrost and other ice-rich horizons (Madsen, et al, 2022; Minsley, et al, 2012), and in many cases are a preferred method of

investigation due to the generally deep depth of investigation in frozen environments. Pandit & King (1979) presented laboratory-based insights into the temperature-salinity-resistivity relationship of porous, saturated limestones and sandstones. Figure 3.4 shows how the resistivity of a limestone sample from these lab experiments varied over several orders of magnitude, depending on the temperature and salinity. The Pandit & King (1979) experiments also found that the resistivity of frozen rocks also depends on the frequency of the electrical signals used to measure them, however, these changes are much weaker compared to those caused by temperature and salinity changes. These resistivity observations will form the basis for geological models that will be used for the interpretation of EM geophysical data in Chapters 5 and 6.

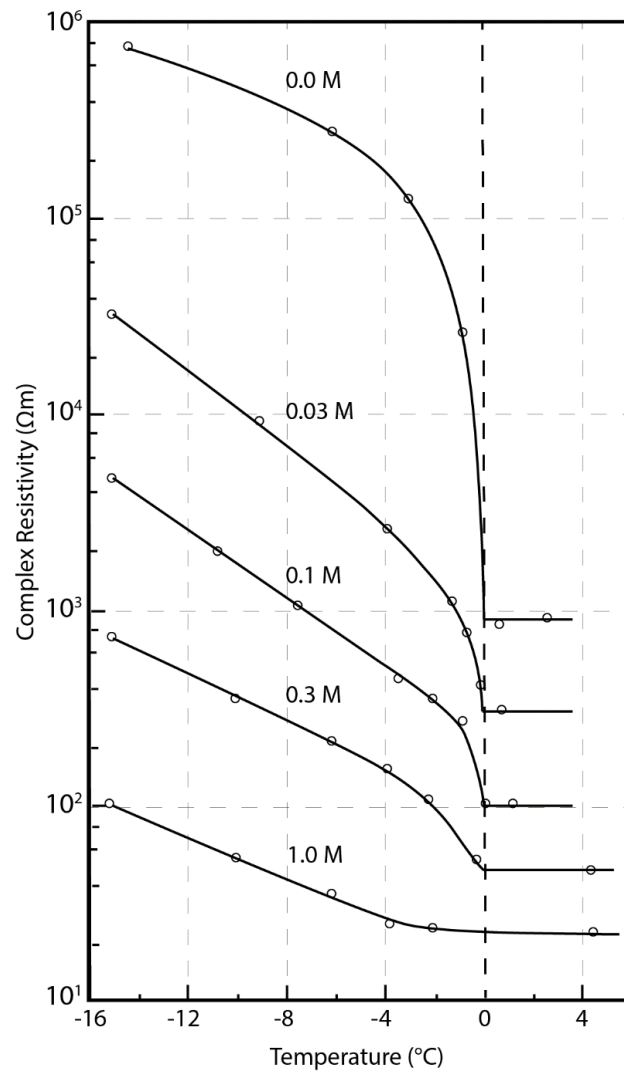


Figure 3.4: Complex resistivity as a function of temperature and salinity for a Salem limestone sample with a porosity of 13%, using a frequency of 1 kHz. Adapted from Pandit & King (1979).

3.3. Resistivity in geological settings

The conduction mechanisms described above also apply in geological settings. Rocks can display a variety of electrical behaviours that are a result of their composition, fabric, porosity, and fluid saturation. The range of resistivity that is seen in rocks spans several orders of magnitudes. Geophysical methods are able to exploit the naturally occurring contrasts in electrical resistivity.

The primary composition of rocks is perhaps the most important factor affecting their electrical resistivity. Conductive ore bodies are able to effectively conduct electricity through their rock grains and can exhibit extremely low resistivity, but are not widely distributed. In exploration settings, ore bodies act as targets with their anomalously low resistivity, compared to typical rocks. Most geologically abundant minerals (quartz, feldspar, mica) are extremely resistive (Freedman & Young, 2016), and thus require elevated porosity and saturating fluids to conduct electricity. For this reason, rock types that have generally low porosity, such as igneous and metamorphic rocks, exhibit generally high resistivity values. Conversely, sedimentary rocks are typically more porous, and thus require pore fluid to enable conduction. Post-depositional weathering can further increase or decrease the resistivity of rocks, depending on the form of weathering. For instance, fracturing tends to elevate the porosity of rocks, thereby lowering the resistivity, whereas the precipitation of material cause cementation of existing pore spaces. Due to this wide range of materials and post-depositional effects, rock resistivity can vary significantly from one type to another (Figure 3.5).

Pore fluids play an important role in the bulk resistivity of rocks, particularly in sedimentary rocks. A range of pore fluids can be present in the pore spaces of rocks, including water, hydrocarbons, gases, partially melted rocks, and brines. Many such fluids can provide pathways for current flow through otherwise resistive rocks; however, this is not always true. Rocks with significant oil content in the pore space will exhibit higher resistivity (Tiab & Donaldson, 2016), due to the high resistivity of the fluid. Analysis of the resistivity of rocks beneath the surface can provide key information about the fluids present, and potentially the size of the reservoir.

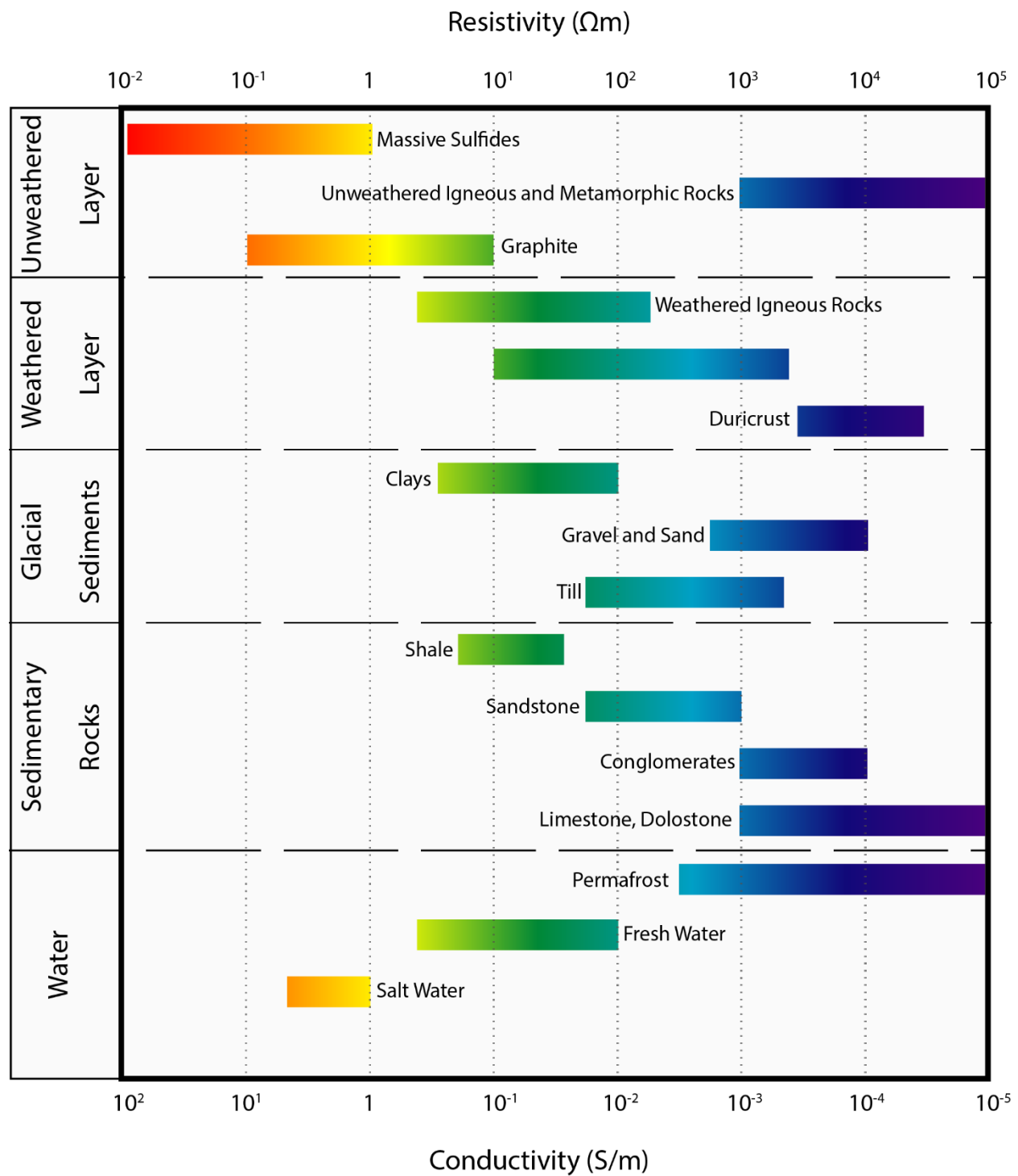


Figure 3.5: A table showing the typical ranges of earth materials. Adapted from Palacky (1987).

Clay content also has a tendency to depress the resistivity of rock. An electric double layer which forms at the water-mineral interface of clay minerals provides an additional pathway for ionic conduction. The cation exchange capacity describes a clay's ability to transfer and receive positively charged particles with adjacent materials, and is much higher in rocks and soils with elevated clay content.

3.4. Permafrost structure

Permafrost is defined as ground that has remained below 0°C for longer than two years. Permafrost is primarily located at extreme latitudes, particularly in the Northern hemisphere. In North America, permafrost is present at up to half of Canada's surface, and more than 80% of the Alaskan ground. Permafrost is an important aspect of engineering considerations and its thawing can lead to damage and failure of infrastructure and buildings. As well, unnatural warming conditions, often associated with anthropogenic climate change is leading to the quickening thaw of existing permafrost. Bush & Lemmen (2019) have estimated that there are up to 1.7 trillion tonnes of organic carbon stored in northern permafrost. Rising temperatures in permafrost worldwide have also been well-documented (Smith et al., 2022). The release of stored carbon dioxide and methane from permafrost is a hotly investigated topic and is predicted to play a substantial role in future climate change acceleration. For these reasons, permafrost monitoring is important in polar regions.

The perennially frozen permafrost layer is typically overlain by an active layer, a relatively thin layer that freezes and thaws with annual changes in temperature. The depth at which permafrost can extend is variable, with the mean annual ground temperature (MAGT) and the geothermal temperature gradient playing important roles in its thickness; however, permafrost thickness can typically range from below 1 m to greater than 1 km (Dobinski, 2011). The base depth of the permafrost layer is coincident with the depth where the geothermal temperature is equal to the freezing point of the earth, which depends on the types and composition of saturating fluids. As permafrost thickness is related to the geothermal gradient, it is highly controlled by the geology, particularly the thermal properties. Figure 3.6 depicts the typical structure of permafrost, along with the general temperature distributions expected. Seasonal changes in surface temperature propagate downwards through the active layer, leading to variation in ground temperatures that become smaller with depth. At a depth known as the depth of zero annual amplitude (DZAA), the

time of propagation for annual surface variations to depth is less than that of the variations themselves. The lower the frequency of temperature variations at the surface, the further they propagate into the Earth, thus ground temperature measurements can be used to identify historic temperature patterns. Higher temperatures at the surface than in the active layer are frequently recorded and are known as the thermal offset, resulting from differences in thermal conductivities between frozen and unfrozen materials (Smith, et al., 2022, Romanovsky & Osterkamp, 1995).

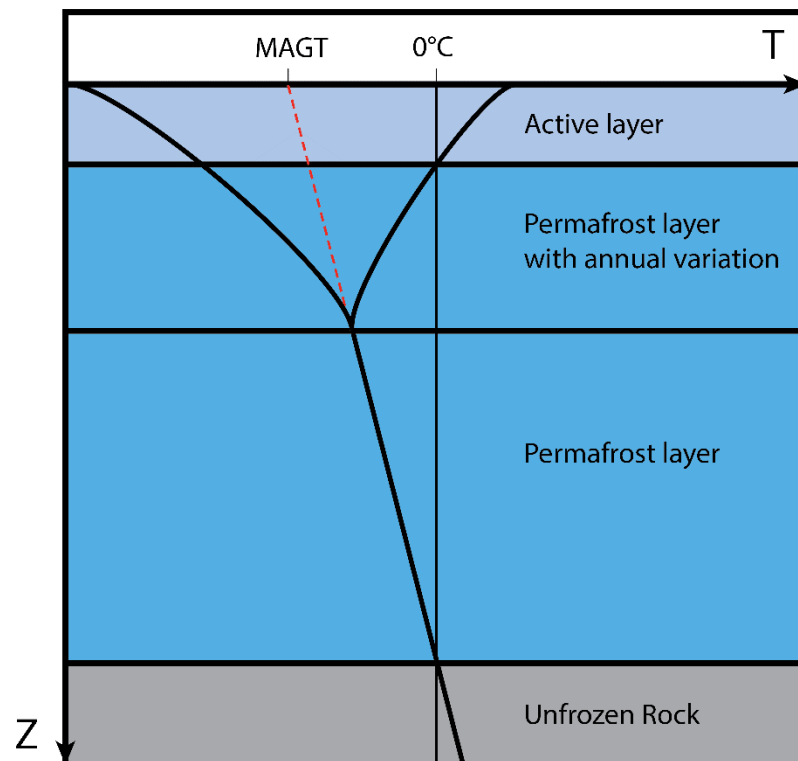


Figure 3.6: A schematic depicting the ground thermal regime of permafrost. Temperature varies with time in the active and upper-permafrost layers due to surface temperature variations. The base of the permafrost layer is determined by the mean annual ground temperature (MAGT) and the geothermal gradient (The slope of the dashed red line). Higher temperature at the surface than within the active layer is known as the thermal offset.

An understanding of the how the composition and physical state affects the electrical resistivity of rocks is essential for using electrical and electromagnetic methods that measure these properties. Both geothermal systems and regions with permafrost contain characteristic features that can be distinguished using their electrical properties. The following chapter will introduce a variety of

geophysical methods that measure subsurface electrical resistivity, and will discuss the strengths and weaknesses of each. The magnetotelluric method will be described in detail as a primary method capable of successfully imaging regions of interest for geothermal development.

Chapter 4: Electrical and Electromagnetic Methods for Geothermal and Permafrost Studies

4.1. Introduction

Electromagnetic (EM) methods is a rapidly developing field of geophysics, which provides information about the subsurface structure and which is complementary to the information obtained from seismic surveys. EM methods measure the electrical resistivity of the Earth, a parameter that is sensitive to the presence of water, ice, and a number of economically important minerals. These methods have become important in the search for geothermal resources, permafrost monitoring, and the exploration of mineral resources, amongst other applications (Ferguson et al., 2012; Oldenborger, 2024). A range of electrical and EM methods have been developed to measure subsurface resistivity, each of which have distinct advantages and disadvantages. The equipment, logistical and financial costs, and possible depths of investigation vary between these methods, and thus the choice of technique is extremely important, and dependant on the area of study.

Electrical geophysical techniques involve the measurement either of electric fields, naturally occurring or of those produced by injecting currents into the ground. Electrical signals that are used in these methods are direct currents with a frequency of 0 Hz. Electromagnetic methods involve the measurement of electromagnetic fields that interact with the Earth. The frequency of the signals being measured largely controls the depth of investigation, and thus the source of the signal is very important. At the most fundamental level, EM methods can be divided into active and passive methods. Active, or controlled source methods utilize electromagnetic sources that are produced artificially, and are more easily controlled by the operator. Controlled source methods utilize powered transmitters to transmit EM signals into the subsurface and can be further divided into two major categories: frequency-domain systems, where signals are transmitted continuously at specific frequencies, and time-domain systems that involve measuring the decay of induced fields when the transmitter is pulsed. Passive source methods use naturally occurring electromagnetic signals, and thus do not require a transmitter to be operated. Natural source magnetotellurics (MT) measures time-varying electromagnetic signals that occur naturally, originating from solar and lightning activity. The variety of available EM techniques enables them to be used for a number of survey types. Electromagnetic surveying techniques are often flexible,

and data can be collected with instruments that are airborne, placed on the surface of Earth, under the ocean, and in boreholes.

This chapter will outline some common types of electrical and electromagnetic methods. The physical mechanisms underlying these techniques and their advantages and limitations will be described. Magnetotelluric surveys performed in Cambridge Bay and Resolute Bay are a central part of this thesis and are described in Chapters 5 and 6. To give the necessary background the fundamental theory of MT will be covered in detail in this chapter. The following summary serves to explain why the MT method is a suitable method for geothermal exploration in Arctic environments.

4.2. Electrical geophysical techniques

Electrical resistivity methods were among the earliest developed techniques used for determining the subsurface resistivity (Zhdanov, 2010). They were first developed during the late 1800s, but it was not until the early 1900s that the first direct current (DC) resistivity surveys were conducted (Zhdanov, 2010). Today, DC resistivity methods are widely used for a variety of purposes: hydrocarbon and mineral exploration, hydrogeology assessment, environmental studies, geotechnical studies, archaeology, and agricultural monitoring.

4.2.1. Electrical resistivity tomography

DC current methods measure subsurface resistivity structure by injecting a direct electrical current into the ground and measuring the voltage response at different locations at or near the surface. In a DC resistivity survey, current travels between electrodes, and the path is affected by the electrical resistivity structure of the subsurface. Electrical resistivity tomography (ERT) uses a series of such measurements to produce a model of the subsurface resistivity. Practically, this involves the measurement of the current and potential difference at a range of different electrode spacings. Modern ‘smart’ acquisition systems are capable of simultaneous measurements between a large number of electrode pairs along a profile or grid, drastically increasing the efficiency of data collection. Figure 4.1 shows an example of the arrangement of electrodes in a DC resistivity survey. Key characteristics of DC methods are that the depth of exploration is largely dependant on the separation of electrodes in the measurement array, and that there are a range of electrode configurations that have distinct differences in their capabilities.

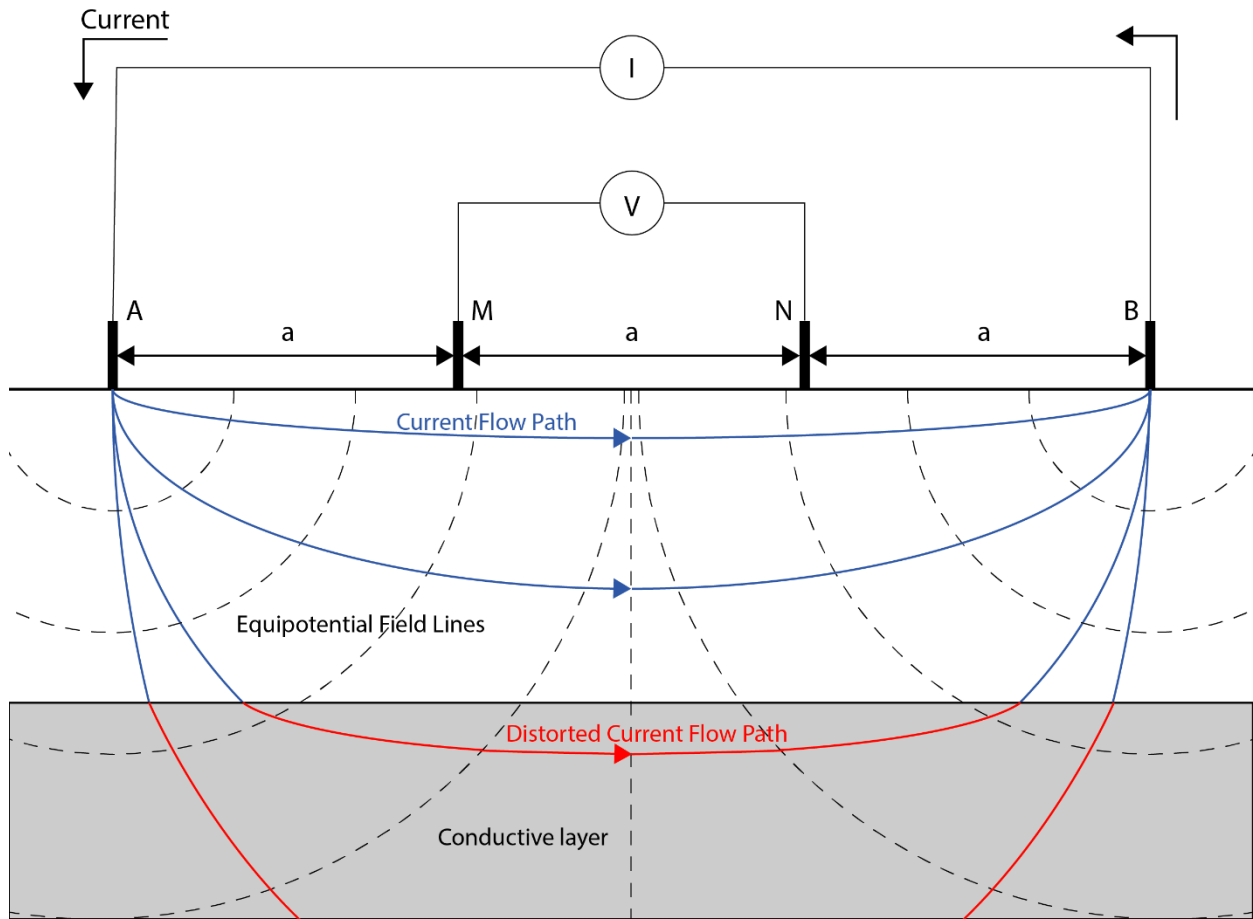


Figure 4.1: A diagram showing the geometry of a DC resistivity survey, using the Wenner array configuration over a resistive layer overlying a more conductive layer. Black dashed lines represent equipotential surfaces, and solid arrows in the subsurface show current flow paths. A current, I , is injected through the subsurface between current electrodes A and B. The potential difference, V , between M and N is measured, allowing for the calculation of the apparent resistivity below the surface. The depth of exploration depends on the electrode spacing, a . Current paths are deflected upon entering the less resistive layer below.

The geometry of the DC resistivity survey greatly influences the information that is collected, and there are a number of commonly used configurations for surveying. The number of possible electrode configurations provides flexibility in the design of DC resistivity surveys, and the ideal survey configuration can depend on the geometry or depth of the target, the type of model that is required, or other logistical constraints. Some of the most common arrays for DC resistivity surveying include the Wenner, Schlumberger, Dipole-dipole, and Pole-dipole arrays. Some survey

configurations are more challenging or time consuming to produce data, while others may be capable of deeper exploration, or provide a higher signal-to-noise ratio, improving data quality. For instance, the Dipole-dipole configuration is less sensitive to electromagnetic coupling effects from conductive materials in the subsurface, but has generally lower signal strength between dipoles with large separation (Loke, 1999). In all cases, the maximum depth of exploration of a DC resistivity survey is limited by the spacing of the electrodes, and the largest electrode spacing must generally be several times larger than the greatest sampling depth. Loke (1999) provides a detailed description of the most widely used array configurations, and suggests that the maximum depth of investigation is at most one third of the total array length, and depends on the configuration used. In most cases, the deployment of DC resistivity systems for deeper exploration is very costly, or even logistically impossible. For deeper exploration, or surveys that require high-density measurements near the surface, other geophysical methods may be far more practical.

Numerous studies exist on the wide range of applications of DC resistivity. Mapping aquifers and the extent of groundwater resources is a common use of DC resistivity (U.S. Environmental Protection Agency, 1994). Additionally, fracture systems that are sufficiently shallow can be successfully studied using DC resistivity methods to identify the orientation and fracture porosity of such systems. In recent decades, these methods have been commonly applied to the detection and characterization of permafrost (Herring et al., 2024).

The flexibility that DC methods allow in survey design, as well as their relative cost-efficiency and straightforward application make them an attractive choice for exploration in a number of different contexts. The methods are largely hindered by the requirement for a large electrode spacing to image deeper structure.

4.2.2. Induced polarization

As electrical prospecting methods were being developed, experiments by the Schlumberger brothers revealed the induced polarization (IP) effect. This occurs when current is transmitted into a polarizable material in the subsurface. Under the presence of an electric field, mobile charged particles separate within the medium, producing a secondary electric field that is opposite the inducing field (Figure 4.2). The primary cause of the IP effect is electrochemical in nature. In the presence of electrolyte-filled pores that are blocked by metallic minerals, external electric fields produce a buildup of charge at the electrolyte-metal interface (Sumner, 1976). Availability of the

electrolyte-metal interface is integral to the formation of strong polarization, and so the effect is most strongly seen concurrently with metallic mineralization. When the inducing field is removed (turned off), the induced voltage briefly persists as the particles return to their original state. Measurements of the decaying induced voltage can provide information about the subsurface chargeability, which is a material property that is commonly used in mineral exploration. As the configurations needed for IP and DC resistivity surveying are very similar, it is common for modern exploration programs to measure resistivity and chargeability contemporaneously as part of a single DCIP survey.

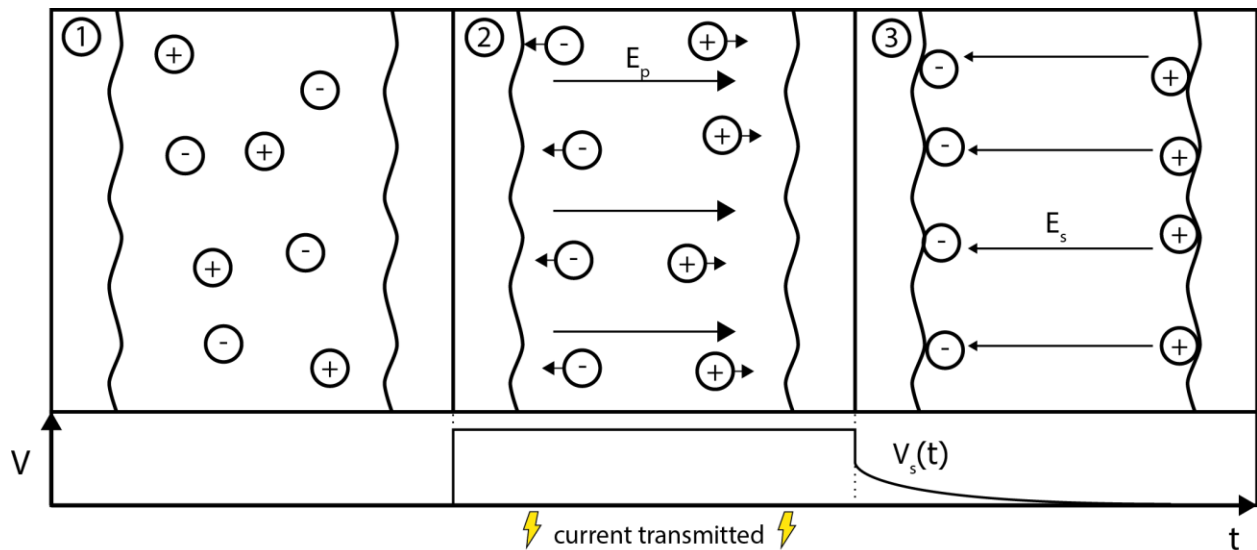


Figure 4.2: An illustration of the IP effect in polarizable material. Under an external electric field, polarizable materials will align. After the external field is removed, a decaying secondary voltage can be measured, which provides information about the chargeability of the material.

4.3. Electromagnetic methods

Electromagnetic methods involve the use of time-varying electromagnetic fields that interact with the Earth. With knowledge of how EM signals interact with material beneath the surface, information about the subsurface electrical resistivity can be obtained. The signals used in EM methods have non-zero frequencies, and can either be naturally occurring or produced by an artificial source, often called a transmitter (Tx). The depth of exploration of EM methods is dependent on the frequency of the signals being measured and the properties of the Earth. These methods comprise a family of techniques that differ in their source and receiver types and

configurations, which ultimately influence their utility in imaging the subsurface. EM methods used for measuring subsurface electrical resistivity work by using transmitter-generated signals that induce currents in the ground. The magnitude of these currents is dependent on the resistivity structure. The currents generate a secondary magnetic field that is measured at the surface. Measurements of the secondary magnetic field allow the resistivity to be determined. The range of EM methods and equipment available today allow for a wide variety of surveys, from near-surface exploration within the first few metres, to deep structure studies reaching depths of 100 km or more (Spies, 1989).

The electromagnetic skin depth is defined as the depth at which an EM signal attenuates to a value that is $1/e$ of its original amplitude at the surface. The skin depth is crucial to understand how deep EM signals are able to penetrate into the Earth, and is consequently important to determine the depths EM geophysical techniques are able to investigate. The skin depth, δ , is calculated (in meters) by:

$$\delta \approx 503 \sqrt{\frac{\rho}{f}} \quad (4.1)$$

where ρ is the resistivity in Ωm , and f is the frequency in Hz. This result shows that the propagation of EM waves depends on their frequency, and the resistivity of the Earth. Thus, lower frequency signals are generally more useful for deeper investigation. This concept will be covered in more detail in Section 4.4.2.

4.3.1. Frequency-domain controlled source systems

Frequency-domain controlled source electromagnetic (CSEM) systems generate continuous EM signals that couple inductively with subsurface materials. The transmitter (Tx) generates a primary magnetic field at either a single frequency or a set of specific frequencies. This primary magnetic field induces electric currents in the subsurface. If a low-resistivity body is present, the induced currents will be stronger than in the host rock. The secondary currents can be measured from the secondary magnetic fields that they generate. Importantly, the primary magnetic field is generated continuously. Thus, at the receiver, both primary and secondary fields are measured simultaneously. Controlled source electromagnetic transmitters can be either magnetic or electric dipoles and are produced by transmitting current in either a loop of insulated wire, or a straight

length of wire connected to the Earth at each end. A number of controlled source transmitter-receiver configurations exist, and some modern systems are capable of surveying using multiple fixed frequencies at a given time.

Modern frequency-domain CSEM systems can be adapted to a range of configurations in different types of exploration. Controlled source methods using higher frequencies are capable of resolving buried targets with greater resolution than other methods, whereas adoption of lower frequencies can achieve greater depths of exploration. The use of a transmitter helps to mitigate the effect of electromagnetic noise, whereas other EM methods may be more sensitive to noise. EM induction systems are commonly used in a Slingram configuration (Figure 4.3), where the transmitter and receiver are fixed at either side of a boom that can be transported on the ground, by air, or underwater for rapid profiling. Such fixed-configuration methods are useful for providing continuous data along a large survey region.

The depth of exploration of Slingram systems depends on the distance between the transmitter and receiver, as well as the frequencies used, thus smaller ground-based systems are generally used for shallow exploration applications, such as groundwater and contaminant mapping, or geotechnical investigation. Sophisticated airborne frequency-domain CSEM systems such as the RESOLVE and DIGHEM systems are much larger and operate at multiple frequencies (Legault, 2015). The RESOLVE system is capable of investigating to depths of ~ 100 m, and is primarily used for groundwater and engineering applications, whereas the DIGHEM system can reach depths of ~ 150 m, and is useful for shallow mineral exploration (Legault, 2015). Frequency-domain CSEM systems are mainly disadvantaged for studying structures deeper than a few hundred meters at most, where the induced response becomes challenging to detect in the presence of the transmitter field. For deeper studies, time domain or natural source systems are typically more capable.

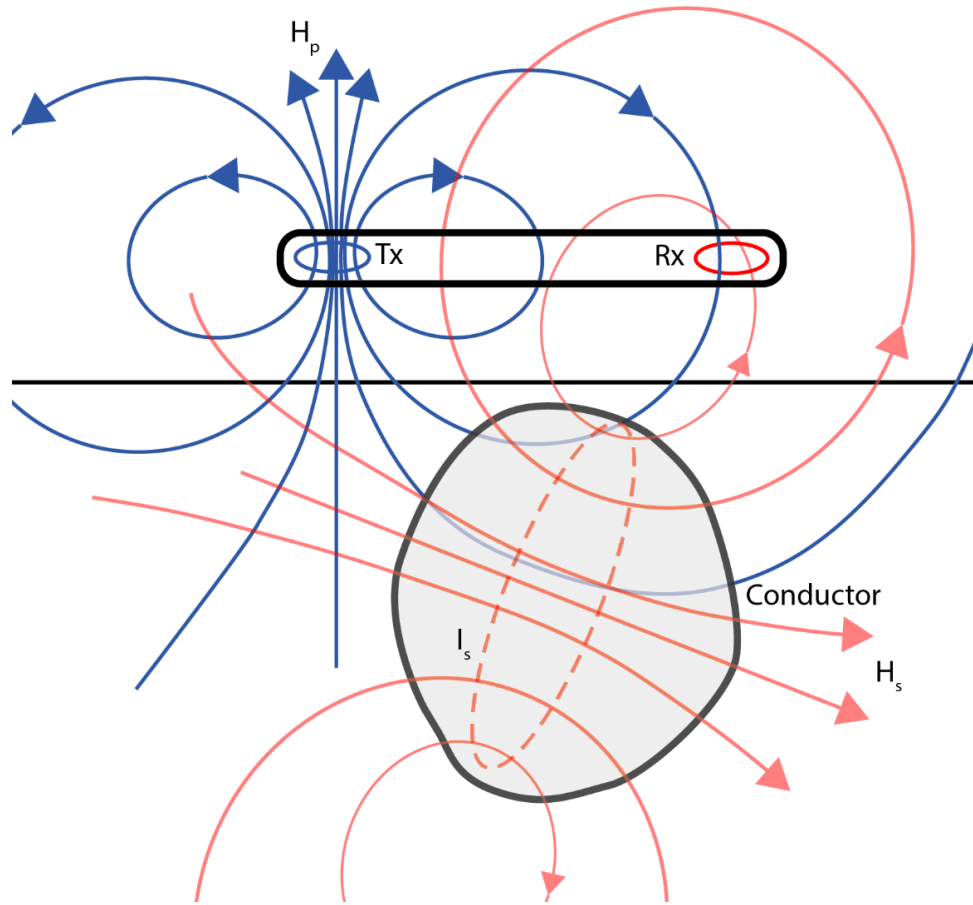


Figure 4.3: A schematic of a frequency-domain CSEM system using a Slingram setup, where the transmitter, Tx, and the receiver, Rx, are fixed to either end of a boom. Blue lines represent the primary magnetic field generated by Tx. Red lines represent induced electromagnetic fields in the buried conductive body that can be detected by Rx. Similar systems can be carried on the ground, flown underneath an airborne vehicle, or towed underwater by a ship.

4.3.2. Time-domain electromagnetic systems

As discussed in the previous section, controlled-source frequency domain systems are limited by their depth of exploration. In the presence of the stronger transmitter field, induced fields can be difficult to detect, particularly those produced by deep conductors. An alternative approach to solve this problem is the use of time-domain electromagnetic methods (TDEM), also commonly called transient electromagnetics (TEM), which uses a pulsing source field that is turned on and off to provide a window in time in which the secondary field can be measured when no primary field is present.

TEM methods typically use a loop transmitter to produce a primary magnetic field at, or above the surface of the Earth. Pulsing of the source field produces a change in the primary magnetic field, which induces a secondary electric current in subsurface conductors. Secondary magnetic fields in the Earth are generated from these ground currents, and can be detected by receivers at or near the surface. Over time, the induced ground currents diffuse downwards and away from the original source, providing information at greater depths (Figure 4.4a). The diffusive pattern of induced electric currents has been likened to the motion of smoke rings (Nabighian, 1979). In the absence of the primary source field, the secondary field can be more easily detected, allowing for greater depths of exploration when compared to frequency domain systems, where the primary and secondary fields are measured simultaneously.

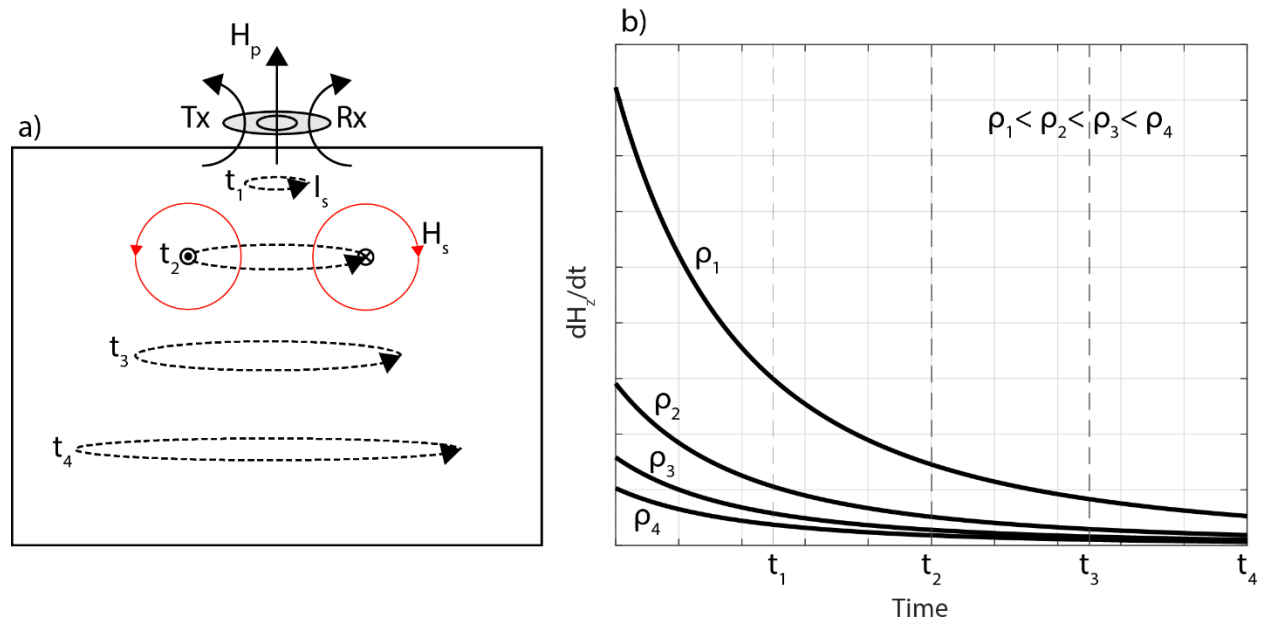


Figure 4.4: A schematic of a central loop time-domain EM (TDEM) system, where the transmitter and receiver are centered at the same location. a) a primary magnetic field, H_p , is produced at the transmitter (Tx) and induces ground currents, I_s , in the subsurface, which diffuse downwards and laterally away from Tx. Induced ground currents produce a secondary magnetic field, H_s , which can be measured at the receiver (Rx). b) The rate of decay of the secondary field measured during the off-time provides information about the subsurface. Induced ground currents dissipate more slowly in less resistive rock, allowing for greater depths of exploration.

The key measurement in TEM methods is the time decay of the secondary magnetic field. As the induced secondary eddy currents diffuse to greater depths, the secondary magnetic field measured at the receiver weakens. The diffusive energy of the ground currents is dissipated more quickly in resistive ground, leading to a faster decay of the secondary magnetic field (Figure 4.4b). Consequently, the rate of decay of the induced magnetic field provides information of the subsurface resistivity with depth. The maximum depth of exploration is limited by the length of time that the secondary field can be detected, since later times correspond to samples at greater depths. Fast transient decay in resistive media may limit the effectiveness of the TEM method where subsurface resistivity is high. Increasing the initial amplitude of the induced magnetic field can extend the time at which it can be detected, increasing the depth of exploration. This is done in practice by using transmitters with the largest possible dipole moments, by increasing the electric current, area of the transmitter loop, or increasing the number of turns of wire in the loop.

Currently, a wide range of TEM systems exist for exploration of the subsurface. Ground based TEM systems are relatively simple and have the benefit of allowing for flexible survey design. Such systems often allow for larger transmitter loops to be achieved, and thus do not always require extremely powerful transmitters to produce a large dipole moment at the source. The stationary setup of ground-based systems also allows for prolonged data collection over multiple transmitter cycles at a single location, where data stacking can significantly improve the signal-to-noise ratio, improving data quality and late time measurements. Unfortunately, the setup of ground-based TEM systems is often far more labour intensive than vehicle-based acquisition systems, and data collection can be challenging and time consuming. A number of airborne TEM systems have been developed that allow for data collection over a much larger region. Airborne systems involve transmitter loops that are either flown underneath the flight vehicle, or wrapped around the wings, in the case of some fixed-wing surveys (e.g. VTEM, MEGATEM). Airborne EM systems are often disadvantaged by the high costs of aircraft operation and generally lower horizontal resolution. Transmitter systems used in airborne TEM systems must be more powerful to compensate for the smaller transmitter loops possible in airborne operations.

Time domain systems are capable of measuring subsurface conditions to depths of several hundred meters or even several kilometers, depending on the strength of the TEM system and subsurface conditions (EPA, 1994). Additionally, the range of available TEM systems enables exploration at

both large and small scales. Because of these reasons, these systems can be well-suited for a variety of exploration purposes, including mineral exploration, groundwater study, and permafrost monitoring (Auken et al., 2003; Zhdanov et al., 2018; Oldenborger, 2024). The challenging requirements for TEM soundings to sample greater depths can make it unsuitable for the exploration of very deep structure, such as some geothermal or deep mineral systems.

4.3.3. Natural source electromagnetic methods

Natural source EM methods utilize naturally occurring electromagnetic fields to investigate the subsurface. These methods rely on similar interactions of a primary magnetic field that induces secondary electric currents in subsurface materials with non-zero conductivity. However, unlike controlled-source EM methods, they do not require the generation of primary electromagnetic fields, which can have distinct advantages. Magnetotellurics, audio-magnetotellurics, and telluric current methods are examples of methods that use these natural signals to explore beneath the surface.

The ability of natural source EM methods to be implemented without using a transmitter significantly reduces the costs and complexity of using such systems. Additionally, the frequencies of natural signals can penetrate deep into the earth (Eq. 4.1), allowing for the exploration of deeper structure (Jones & Ferguson, 2001). Such systems do have their disadvantages: natural EM signal strength is inconsistent and varies significantly over time, and natural source methods are most effective in electromagnetically quiet areas. Following solar activity, signal strength can undergo annual and longer-term variations that can affect data quality of natural source methods (Viljanen, 2012). Additionally, natural source methods are more sensitive to external noise, and utilization of natural sources can be more complicated, as the source field is not precisely known. Magnetotelluric studies have proven very useful for tectonic studies, geothermal exploration, groundwater studies, as well as many other applications (Ferguson et al., 2012). Next, the magnetotelluric method will be covered in detail, including the physical background of the method, and several techniques that are commonly used in the analysis of MT data.

4.4. The magnetotelluric method

Magnetotellurics (MT) is a passive electromagnetic geophysical method that is used to determine the resistivity structure of the Earth. Cagniard (1953) provided the first theoretical overview of the

method, which has since seen widespread adoption and development. Signals used in MT exploration are typically in the range of $10^3 - 10^5$ Hz, with high frequency signals produced by global lightning activity, and lower frequencies (< 1 Hz) originating from interactions between solar winds and the Earth's magnetosphere. Processed MT data provide estimates of the electromagnetic impedance tensor as a function of frequency, which also provides information on the apparent resistivity and phase. The impedance is used in MT because it only depends on properties of the Earth, and therefore, the primary field does not need to be known. These quantities are typically used as input information for inversion, a mathematical process that converts data as a function of frequency into a model of the subsurface resistivity within space. MT inversions use apparent resistivity and phase information to produce a best-fitting geophysical model of the subsurface resistivity in one, two, or three dimensions. Geophysical inversion models show resistivity distributions in the subsurface, which are commonly interpreted with additional information to make inferences about the geology of the investigated region. The resistivity of Earth materials have been shown to vary by many orders of magnitude (Simpson & Bahr, 2005), and thus resistivity contrasts can provide meaningful information about the geology.

4.4.1. Electromagnetic signal propagation in the Earth

The MT method is based upon the behaviour of electromagnetic waves in conductive, magnetisable media, as described by Maxwell's Equations (Maxwell, 1873):

$$\nabla \cdot \mathbf{E} = \frac{Q}{\epsilon_0} \quad (4.2)$$

$$\nabla \cdot \mathbf{B} = 0 \quad (4.3)$$

$$\nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t} \quad (4.4)$$

$$\nabla \times \mathbf{B} = \mu_0 \sigma \mathbf{E} + \mu_0 \epsilon_0 \frac{\partial \mathbf{E}}{\partial t} \quad (4.5)$$

where $\mathbf{E} = \mathbf{E}(x,y,z,t)$ is the electric field, $\mathbf{B} = \mathbf{B}(x,y,z,t)$ is the magnetic flux density, Q is the electric volumetric charge density, ϵ_0 is the dielectric permittivity of free space, μ_0 is the magnetic permeability of free space, σ is the electrical conductivity, and t is time. Here the Cartesian x - y - z coordinate system is being used with z in the downward direction. For materials typically found in

the Earth the assumption that $\mu_0 \approx \mu$, and $\varepsilon_0 \approx \varepsilon$ will be used. It will be shown that the second term of Eq 4.5 can be ignored in the context of MT.

Gauss' Law (Eq. 4.2) states that the electric flux through a surface is related to the charge contained within the surface. Gauss' Law for magnetism (Eq. 4.3) states that the magnetic flux through a surface is always zero, implying that magnetic monopoles do not exist. These two equations describe the behaviour of stationary distributions of electric and magnetic charge. Equations 4.4 (Faraday's Law) and 4.5 (Ampère's Law) describe the inductive coupling of EM waves, where a time-varying electric field will produce a magnetic field, and a time-varying magnetic field will produce an electric field. Faraday's Law and Ampère's Law describe the mechanism in which electromagnetic signals propagate.

Maxwell's Equations can be used to analyze the behaviour of electromagnetic waves as they propagate through the Earth, forming the basis for how magnetotellurics can be used to investigate Earth's electrical and magnetic properties. In the case of an Earth with time invariant properties and no free electric charge, the divergence of the electric field equals zero, according to Equation 4.2. By taking the curl of Equation 4.4 and substituting it into Equation 4.5 one can derive a diffusion equation for the electric field:

$$\nabla^2 \mathbf{E} = \mu\sigma \frac{\partial \mathbf{E}}{\partial t} + \mu\varepsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad (4.6)$$

A time-dependent harmonic solution to Equation 4.6 is sought and thus a solution for this differential equation is given by:

$$\mathbf{E}(x, y, z, t) = \mathbf{E}_0 e^{-i\omega t} \quad (4.7)$$

where $i = \sqrt{-1}$, and $\omega = 2\pi f$ is the angular frequency.

Substituting our solution into the diffusion equation we can write Eq. 4.6 as:

$$\nabla^2 \mathbf{E}_0 = -i\omega\mu\sigma \mathbf{E}_0 + \omega^2\mu\varepsilon \mathbf{E}_0 \quad (4.8)$$

The right-hand side of Eq. 4.8 is separated into a conduction current component, and a displacement current component. The conduction current is a result of the motion of charged particles in the medium, whereas the displacement current originates from changes in the electric displacement field. In the case of frequency and conductivity values common in MT the

conduction current will dominate, resulting in the diffusion of EM signals in the Earth. Assuming that the conduction current is much greater than the displacement current, Eq. 4.8 can be approximated as:

$$\nabla^2 \mathbf{E}_0 = -i\omega\mu\sigma\mathbf{E}_0 \quad (4.9)$$

which is the differential equation that governs the behaviour of EM waves that is useful for the MT method.

4.4.2. Electromagnetic skin depth

A unique aspect of electromagnetic methods is that the depth of investigation is directly related to the frequencies of the signals being observed. To investigate this relationship one can find a solution to Eq. 4.9 for the case of a plane wave that is polarized in the x -direction travelling towards the Earth. The plane wave assumption is generally valid in MT because of the large distance from the source. At the free-air boundary (the interface between the ground and air) an incident EM wave will have a small fraction of its energy transmitted across the boundary, while the remainder is reflected away from the boundary. The equation

$$E_x(z) = E_x^0 e^{-\sqrt{\frac{\omega\mu\sigma}{2}}z} e^{i\sqrt{\frac{\omega\mu\sigma}{2}}z} \quad (4.10)$$

is a solution to the ordinary differential equation of Eq. 4.8. and describes an oscillating wave that decays in amplitude with depth. If we consider the magnitude of this solution as it diffuses downwards through the Earth, we see that:

$$|E_x(z)| = |E_x^0| e^{-\sqrt{\frac{\omega\mu\sigma}{2}}z} \quad (4.11)$$

The sensitivity of these signals is related to their skin depth, that is, the depth at which the signal amplitude decays to $\frac{1}{e}$ of its value at the surface. The skin depth, δ , can be derived from Eq. 4.11:

$$|E_x(\delta)| = \frac{|E_x^0|}{e} = |E_x^0| e^{-\sqrt{\frac{\omega\mu\sigma}{2}}\delta} \quad (4.12)$$

it follows that

$$\delta = \frac{1}{\sqrt{\pi\mu}} \sqrt{\frac{\rho}{f}} \quad (4.13)$$

Using the assumption that $\mu = \mu_0 = 4\pi \times 10^{-7}$ H/m, one can obtain the same result for the skin depth as Eq. 4.1. This result shows that the penetration depth of an EM signal diffusing through the earth is proportional to the square root of the resistivity of the ground material divided by the frequency of the signal. Recognizing that the period relates to the frequency as $T = \frac{1}{f}$, one can see that the longer the periods of the signals measured at an MT station, the greater the depth of investigation.

4.4.3. Apparent resistivity and phase

One can further study the propagation of a plane wave, polarized in the x -direction travelling through a 1-D Earth from the surface. From the derivation above, an incident wave upon the surface will result in a reflected wave returning upwards, and a transmitted wave diffusing downwards through the Earth.

Eq. 4.9 can be solved by:

$$\mathbf{E}_x = \mathbf{E}_1 e^{i\omega t} e^{-\sqrt{i\mu\sigma\omega}z} + \mathbf{E}_2 e^{i\omega t} e^{\sqrt{i\mu\sigma\omega}z} \quad (4.14)$$

It is apparent that the second term must be zero, as the electric field strength will not grow with depth, but will instead be dissipated. Therefore,

$$\mathbf{E}_x(t, z) = \mathbf{E}_0 e^{i\omega t} e^{-\sqrt{i\mu\sigma\omega}z} \quad (4.15)$$

One can also take advantage of Faraday's Law (Eq. 4.4);

$$\frac{\partial E_x}{\partial z} = -\frac{\partial B_y}{\partial t} = -i\omega B_y = -\sqrt{i\mu\sigma\omega} E_x \quad (4.16)$$

The relationship shown by Eq. 4.16 allows us to relate the conductivity (or resistivity) to the ratio of the electric and magnetic fields:

$$\frac{\partial E_x}{\partial z} = -\frac{\partial B_y}{\partial t} = -i\omega\mu H_y = -\sqrt{i\mu\sigma\omega} E_x \quad (4.17)$$

$$\rho = \rho(\omega) = \frac{1}{\sigma} = \frac{1}{\omega\mu} \left| \frac{E_x}{H_y} \right|^2 = \frac{1}{\mu\omega} |Z_{xy}|^2 \quad (4.18)$$

relates the resistivity of the Earth to the ratio of the orthogonal electric and magnetic field components. This ratio is also known as the impedance, Z . It is important to note that the resistivity, as well as the impedance are frequency dependent. In the case of a homogeneous Earth with constant resistivity, ρ , is equivalent to the resistivity of the Earth. In the case of a heterogeneous Earth, $\rho = \rho_a$ is the apparent resistivity, an averaged value over a hemisphere of radius equal to the skin depth of the signals being measured. Critically, since the source field is not well known, it is important that the impedance is being measured. Since the impedance depends solely on the physical properties of the Earth, the absolute magnitude of the source field does not need to be known.

Additional information can also be obtained from the phase difference between measured fields:

$$\Phi(\omega) = \tan^{-1} \left[\frac{E_x}{H_y} \right] = \tan^{-1} [Z_{xy}] \quad (4.19)$$

For a homogeneous half-space, it can be shown that the phase will be constant, $\Phi = \frac{\pi}{4}$. For a system with changing resistivity with depth, the phase is sensitive to the changes with depth. $0 \leq \Phi < \frac{\pi}{4}$, when resistivity is increasing with depth, and $\frac{\pi}{4} < \Phi \leq \frac{\pi}{2}$ when resistivity is decreasing. That is, the phase difference between the coupled electric and magnetic fields is sensitive to contrasts in resistivity as the waves propagate through the Earth. The relationship between apparent resistivity and phase is important, with consistency of the two being a helpful indicator of data quality. While phase is generally confined to the range of $0 \leq \Phi \leq \frac{\pi}{2}$, near surface conductors can produce outlying values due to galvanic effects.

4.4.4. Dimensionality of MT data

It is important to understand the characteristics of MT data before inversion. Therefore, one should aim to identify whether their data are consistent with a 1-D, 2-D, or 3-D resistivity structure (Figure 4.5). The impedance tensor is represented by the relationship:

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} Z_{xx}(\omega) & Z_{xy}(\omega) \\ Z_{yx}(\omega) & Z_{yy}(\omega) \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix} \quad (4.20)$$

where the off-diagonal impedance components, Z_{xy} and Z_{yx} , differ due to the 2-D resistivity structure. The geoelectric strike is defined as the angle that will minimize the diagonal components

of the impedance tensor when rotating the tensor by that angle. The strike direction is important in characterizing a 2-D electrical structure and may be consistent with the strike of geologic structures being investigated. The presence of different impedances in the xy and yx quadrants of the impedance tensor implies that measurements of the apparent resistivity and phase differ in a 2-D setting, depending on the components of the electric and magnetic fields used:

$$\rho_{xy}(\omega) = \frac{1}{\omega\mu} |Z_{xy}(\omega)|^2; \quad \Phi_{xy}(\omega) = \arg(Z_{xy}(\omega)) \quad (4.21)$$

$$\rho_{yx}(\omega) = \frac{1}{\omega\mu} |Z_{yx}(\omega)|^2; \quad \Phi_{yx}(\omega) = \arg(Z_{yx}(\omega)) \quad (4.22)$$

Thus, apparent resistivity and phase sounding curves can provide a preliminary indication on the dimensionality of the resistivity structure.

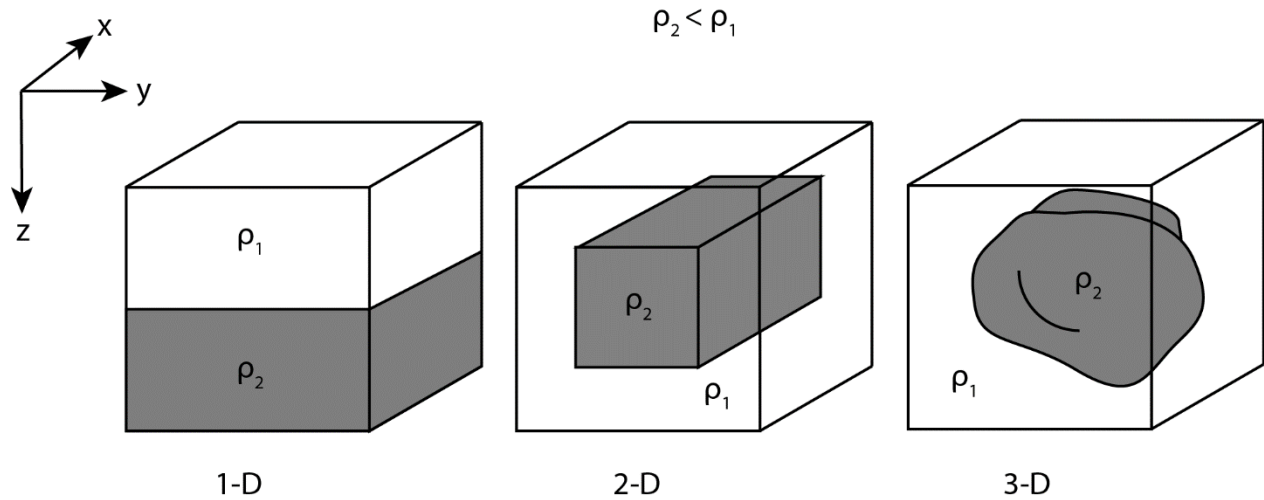


Figure 4.5: A diagram demonstrating the differences between 1-D, 2-D, and 3-D resistivity structures. 1-D structure only varies in a single direction, 2-D structure varies in two directions, and 3-D structure varies in all three directions. In this case, the geoelectric strike of the 2-D structure is in the x -direction.

4.4.5. EM field polarizations

Within a 2-dimensional resistivity structure, two electromagnetic polarizations are defined:

1. Transverse Electric (TE) mode: The electric field is parallel and continuous along the geoelectric strike direction. Currents flow along this direction and are unimpeded by the boundaries formed by the 2D structure.
2. Transverse Magnetic (TM) mode: The electric field is perpendicular to the geoelectric strike direction and is discontinuous across the boundaries of the 2D structure. Current flow is impeded by the boundaries of the 2D structure and discontinuous across these boundaries.

The physics of electromagnetic waves travelling through a conductive material lead to additional behaviours as they diffuse through the ground. Inductive effects arise from electromagnetic signals that attenuate as they pass through the ground. An attenuating electric field will induce a secondary magnetic field in the Earth, which can be measured at the surface. As the attenuation of these signals is described by the skin depth, this behaviour is frequency dependent and tends to disappear at lower frequencies. Galvanic effects comprise a second set of behaviour and are caused by the accumulation of charge at boundaries of electrically dissimilar materials. Free electric charges at boundaries will produce a secondary electric field that may also be recorded. Galvanic effects are frequency independent and so can be seen at all frequencies.

In the TE mode, charges do not develop along the edge of the conductor, because the electric field is continuous along the length of the conductor. The TE mode is therefore unaffected by galvanic effects – it is more sensitive to conductors, and is typically better at resolving deeper structure. The TM mode may be affected by both inductive and galvanic effects, because the electric field is discontinuous across the conductor. The TM mode is more sensitive to near-surface structures than the TE mode, and can sense both conductors and resistors. Due to the strengths and weaknesses of both modes, a bimodal analysis that includes both components is most often used. Analysis of both polarizations is important as both modes are affected differently by subsurface structures. Differences in the TE and TM modes may indicate a departure from one-dimensional electrical structure.

4.4.6. Galvanic effects and static shifts

As discussed in the previous section, discontinuities in material resistivity can result in charge buildup along boundaries. As free charges collect along these boundaries, they will produce a secondary electric field that can be registered at the surface, affecting the accuracy of surface measurements. The phenomenon of static shift is caused by multi-dimensional contrasts in conductivity that are of a much smaller scale than the penetration depth (skin depth) of the EM signals measured (Simpson & Bahr, 2005). Depending on the nature of the conductivity discontinuity, the net electric field that is recorded may be greater than or less than it would be otherwise. In practice, static shifts are presented in data as over or under-determinations of the electrical impedance and the apparent resistivity at all frequencies, with no changes to the phase. Thus, without additional knowledge of the environment, the vertical shift effect of static shifts can be difficult to identify.

With the presence of a local heterogeneity of small scale, the measured apparent resistivity will be scaled by a factor, s :

$$\rho_s = s\rho \quad (4.23)$$

Recall that the skin depth, δ , depends on both the resistivity and period (Eq. 4.1), and so scaling of the apparent resistivity by a factor s , will also scale estimates of the depth of features. Additionally, measurements of apparent resistivity can be used to infer other properties, such as the porosity of sediments. Incorrect estimates of the apparent resistivity from static shifts will also affect these values. For these reasons, it is important that static shifts are identified and corrected to avoid inaccurate analysis of results.

Static shifts may be corrected using several techniques (Simpson & Bahr, 2005), although these will not be discussed in detail here:

1. Comparison of MT data with TEM data, which are unaffected by static shift.
2. Statistical techniques.
3. Long-period measurements of magnetic field data.

4.4.7. Vertical magnetic field (tipper) and induction vectors

When recorded, the vertical magnetic field can be a useful piece of additional information in MT data analysis. In the case of a two-dimensional conductive feature (Figure 4.5) buried in a more resistive space electric currents that travel along the strike of the structure (TE mode) will generate a secondary magnetic field, analogous to the magnetic field generated by the flow of current through a wire. The tipper is defined as the ratio of the vertical and horizontal magnetic field components:

$$I_x = \frac{H_z}{H_x}; I_y = \frac{H_z}{H_y} \quad (4.24)$$

where I_x/I_y and H_x/H_y are the tipper and magnetic field in the x and y directions, and H_z is the vertical magnetic field component. The x and y tipper components can be combined as added vector components:

$$I = \sqrt{I_x^2 + I_y^2} \quad (4.25)$$

The tipper is most commonly visualized on a map view as induction vectors (IVs), visual representations of measured tipper data. Figure 4.6 demonstrates how induction vectors can be used to identify the location of subsurface conductors.

Two conventions are used for displaying induction vectors: the Parkinson convention where IVs point towards conductive bodies, and the Weise convention where IVs point away from conductors. Tipper data are useful in resolving the horizontal location of buried conductors; however, they are poorly suited for resolving their depth.

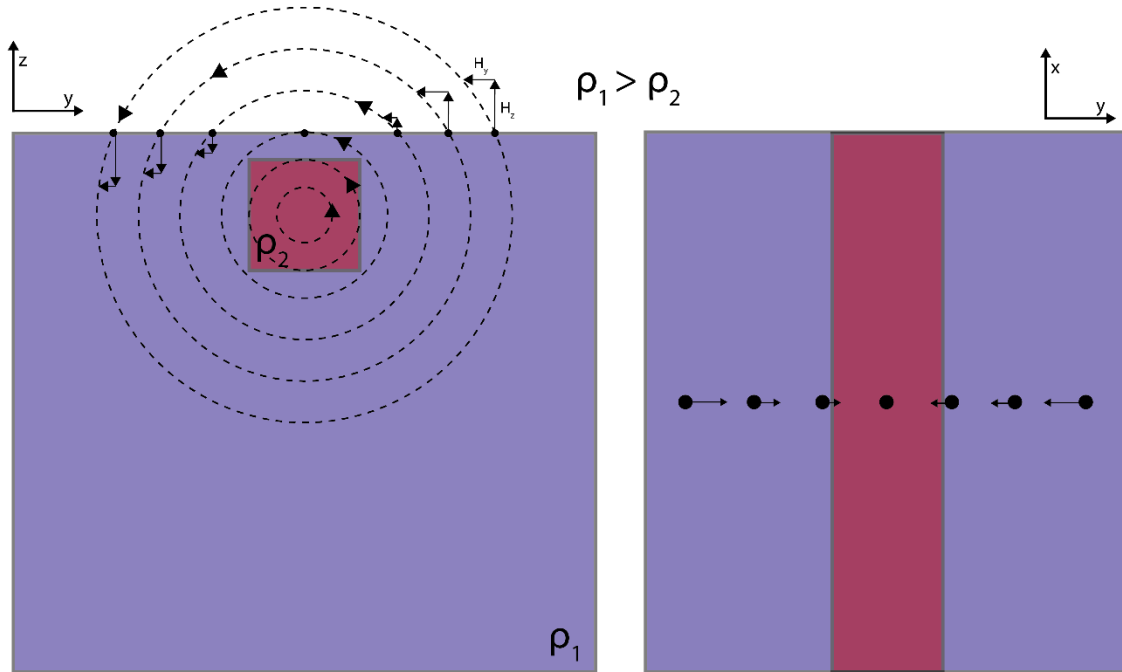


Figure 4.6: Shows how the components of the magnetic field vary across a buried, elongate conductive body. Here, red is more conductive, and blue is more resistive. It is important to note the sign of the tipper, as the directions of the induction vectors change across the buried structure. A top-down view of resulting induction vectors is also shown, using the Parkinson convention.

4.4.8. The magnetotelluric phase tensor

Phase tensors are a visual method of showing changes in the impedance and phase with azimuth. As the phase is being used in this method, phase tensors are unaffected by galvanic effects. The phase tensor was originally developed by Caldwell et al. (2004) and has proven to be a highly effective tool in evaluating the dimensionality of the subsurface electrical structure.

The electrical impedance is a complex valued quantity:

$$\mathbf{Z} = \mathbf{X} + i\mathbf{Y} \quad (4.26)$$

where $\mathbf{X}$ is the real component of the impedance tensor, and $\mathbf{Y}$ is the imaginary component. Using the components of the impedance one can define the phase tensor:

$$\boldsymbol{\Phi} = \mathbf{X}^{-1}\mathbf{Y} \quad (4.27)$$

where $\mathbf{X}^{-1}$ is the inverse of the real component, and Φ is real. Additionally, Φ can be expressed in terms of the components of $\mathbf{X}$ and $\mathbf{Y}$:

$$\begin{bmatrix} \Phi_{11} & \Phi_{12} \\ \Phi_{21} & \Phi_{22} \end{bmatrix} = \frac{1}{\det(\mathbf{X})} \begin{bmatrix} X_{22}Y_{11} - X_{12}Y_{21} & X_{22}Y_{12} - X_{12}Y_{22} \\ X_{11}Y_{21} - X_{21}Y_{11} & X_{11}Y_{22} - X_{21}Y_{12} \end{bmatrix} \quad (4.28)$$

and $\det(\mathbf{X}) = X_{11}X_{22} - X_{21}X_{12}$ is the determinant of $\mathbf{X}$.

The shape of the phase tensor shows the azimuthal variation of the phase, with the semi-major and semi-minor axes representing the maximum and minimum phase values, respectively. The skew angle, β , is an additional rotationally invariant quantity that provides a measure of the asymmetry of the phase tensor. The skew angle is defined as:

$$\beta = \frac{1}{2} \tan^{-1} \left(\frac{\Phi_{12} - \Phi_{21}}{\Phi_{11} + \Phi_{22}} \right) \quad (4.29)$$

The skew angle can be positive or negative. Phase tensors are typically shown as colored ellipses, with the shape and skew together providing information about the dimensionality of the electrical structure (Figure 4.6):

1. For an isotropic, 1-D resistivity structure: A one-dimensional resistivity structure will not produce an azimuthal variation in the measured data. The phase tensor will be circular, and $\beta = 0$.
2. For a 2-D resistivity structure: A two-dimensional resistivity structure will show variation as the phase tensor is rotated. The phase tensor will be an ellipse with the major or minor axis parallel to the strike direction, and the skew angle, $\beta = 0$.
3. For a 3-D resistivity structure: A three-dimensional resistivity structure will also show variation as the phase tensor is rotated. The phase tensor will be ellipsoidal, and the skew angle, $|\beta| > 0$.

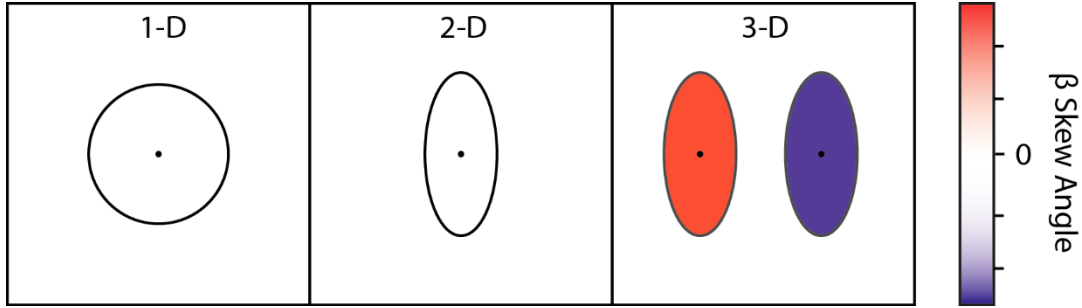


Figure 4.7: Examples of phase tensors that are expected for 1-D, 2-D, and 3-D subsurface dimensionality. High β skew values are indicative of 3-D behaviour. The β skew threshold between 2-D and 3-D behaviour varies, but is typically taken to be between 2° - 3° .

It is important to note that the skew value is not a robust measure of the dimensionality. While a large skew value is prescriptive of a 3-D structure, a low skew value does not necessarily indicate a 2-D resistivity structure, as noise in the data can produce a non-zero skew. A threshold of 2° - 3° is commonly used as an indicator for 3-D behaviour.

4.4.9. Magnetotelluric data collection

The MT method is based upon the collection of orthogonal components of the electromagnetic field. Typically, this is accomplished for the electric field by measuring the voltage between two buried electrodes. The magnetic field is measured with a ground-based magnetometer; an induction coil type magnetometer is usually more sensitive to higher frequency bands (audio magnetotellurics, AMT, and broadband magnetotellurics, BBMT), and a fluxgate type magnetometer is more appropriate for lower frequencies in the case of long-period MT. Data are recorded at a central data logger, with a complete MT station recording on five channels: two electric field components and three magnetic field components. Magnetic sensors and electric field electrode positions are most commonly aligned to the cardinal directions in geomagnetic or geographical coordinates using a compass or GPS system. When surveying at extreme polar latitudes compasses may not be reliable, and alignment to a geographical coordinate system is required.

4.5. Conclusion

A number of powerful electrical and electromagnetic techniques are available for making measurements of subsurface resistivity, each capable of investigating beneath the surface to varying depths, and resolutions. From DC resistivity methods and induced polarization, to electromagnetic induction and magnetotellurics, each technique brings unique advantages that are suited to specific applications. Understanding how these methods work, as well as their strengths and limitations is essential for selecting the most appropriate approach for exploration.

Method	Advantages	Disadvantages
DC Resistivity	• Capable of high resolution • Cost-effective • Versatile	• Labour intensive • Generally large footprint • Restricted depth of exploration • Power source is required
Controlled Source Frequency-Domain EM	• Capable of high resolution • Capable of large coverage • Fast data acquisition • Source signal is known	• Restricted depth of exploration • Sometimes costly • Transmitter is required
Controlled Source Time-Domain EM	• Deeper depth of exploration • Capable of large coverage • Fast data acquisition • Source signal is known	• Limited in resistive settings • Sometimes costly • Transmitter is required
Magnetotellurics	• Flexible depth of exploration • No transmitter required • Cost-effective	• More susceptible to noise • Signal strength is unpredictable • Can be time consuming

Table 4.1: A summary of the advantages and disadvantages of some electrical and electromagnetic methods discussed in this chapter.

The magnetotelluric (MT) method uses recordings of natural electromagnetic signals at the surface to infer the resistivity structure at depth. There are several techniques that can be used to assess the dimensionality of an MT dataset, which determines which modelling technique is most appropriate. Measured resistivity values can then be converted to a resistivity model of the subsurface through the process of inversion, which can be useful in identifying structure as the resistivity of Earth materials can vary by many orders of magnitude. MT is a useful electromagnetic method due to the small, non-destructive footprint, and because there is no requirement to produce an artificial signal. Additionally, the range of frequencies in natural EM

signals allow for flexibility in the depth of investigation, depending on those required by the investigator.

The following chapters will focus on the collection and analysis of magnetotelluric data at two communities in Nunavut: Cambridge Bay, and Resolute Bay. Data in both locations were collected, processed, and used for 1-D and 3-D modelling to assess the subsurface conditions beneath each community. The results will be discussed within the context of potential geothermal development in the communities, with additional insights into the permafrost structure present.

Chapter 5: Magnetotelluric Study at Cambridge Bay, Nunavut

5.1. Introduction

Having introduced the use of electromagnetic methods in geothermal and permafrost studies, the survey activities in both Cambridge Bay and Resolute Bay, Nunavut can be considered. This chapter will focus on the extent and results of the Cambridge Bay survey, with an overview of the magnetotelluric (MT) data collected, as well as an analysis and interpretation of these results. Due to a limited amount of prior work in this region, the local geology of this settlement is poorly mapped. The work done at Cambridge Bay was primarily motivated as part of an investigation into the geothermal potential of this community. As such, the goals of this survey were to:

1. Characterize the local subsurface electrical resistivity structure.
2. Identify potential features of interest in the subsurface, including geological boundaries, evidence of aquifers, and the permafrost structure near the surface.
3. Evaluate the geothermal potential of the community.

Cambridge Bay is positioned along the southeast edge of Victoria Island, a part of the Arctic Archipelago. The community is mainly clustered, with most major infrastructure within approximately 2 km of the main settlement area or airport. A high-voltage transmission line carries power from the power plant in town to various important services including: the Cambridge Bay Airport, the Cambridge Bay Radar Station, and the local water facilities. A Distant Early Warning (DEW) station and seismic line lie along the major roadway trending northwest from the town site. The name DEW line road will be used to describe this roadway.

Geological information of Cambridge Bay, and Victoria Island as a whole is scarce (Okulitch, 1991; Sharpe, 1993; Harrison, et al., 2013). The immediate environment and surficial geology are dominated by a layer of glacial till of variable thickness. In regions defined as till veneer it is less than 1 m thick. Elsewhere, the surface is characterized as a till blanket and ranges from 1 – 15 m thick. A thin, saturated clay layer was encountered while installing MT instruments beneath the surface at many sites, which was frozen below a depth of 1 m or less. Outcrops of bedrock are present at the surface in some locations, such as the gravel pit west of the community and along the shores of the bay and are largely composed of Precambrian or Cambrian limestones and dolostones, with some minor sandstone, siltstone, and shale content.

Knowledge of the thermal regime is essential to interpret the near-surface resistivity in the Arctic. The active layer thickness (ALT) at Cambridge Bay is approximately 1 m (Ran, et al., 2022), and was confirmed during installation of MT instrumentation in July 2023. The MAGT at Cambridge Bay is -10°C (Heginbottom et al., 1995; Ran et al., 2022). There is limited knowledge about subsurface temperature in Nunavut (Minnick et al., 2018). Assuming a relatively low geothermal gradient of $20\text{-}30^{\circ}\text{C}$ per kilometer that would be typical for the Arctic platform, this predicts a permafrost thickness of 300-500 m.

Evidence of post-glacial rebound is also widespread near the community. The surface of Victoria Island was significantly depressed by the weight of the Laurentide Ice Sheet. Melting of the ice sheet triggered a rapid rise in sea level, inundating the surface with seawater that was 100 – 200 m deep (Simon et al., 2016). Johnson (1962) estimated historical rates of uplift at 10 mm per year, which have been more recently confirmed by GPS measurements (Simon, et al., 2016). As a result, rocks at Cambridge Bay may have been submerged in seawater for a significant portion of the Holocene and likely have significant salt-water content in their pore space.

5.2. Magnetotelluric data collection at Cambridge Bay

In July 2023, broadband MT (BBMT) data were recorded at 26 sites in an approximately 17 km x 8 km area around the community of Cambridge Bay, Nunavut (Figure 5.1). Phoenix MTU-5C instruments were used to record the data. As the primary objective of this study relates to the installation of geothermal to service the community, data collection was focused on areas very close to the hamlet. At 20 stations the instruments recorded both electric and magnetic field data. At the remaining 6 stations, only electric field data were recorded, with synchronous magnetic fields from nearby stations being used to process data from these stations. Deployment of magnetic sensors is time consuming and labour intensive; using additional stations with only electric field data can allow for greater data coverage in a time-limited field campaign. Vertical magnetic field data were also recorded at several stations. Care was taken to install equipment an appropriate distance away from obvious sources of cultural noise; however, proximity to the settlement introduced significant interference at some sites. Additionally, the nearby coastal waters of the Northwest Passage introduced noise from the coupled motion of strong winds and the ground surface.

Typical MT field procedures use compasses to orient electric and magnetic sensors in geomagnetic coordinates. However, at high magnetic latitudes such as Cambridge Bay, the geomagnetic field is nearly vertical and the horizontal field components are comparatively weak. Thus, a magnetic compass cannot be used reliably to determine alignment directions. Instead, sensors were aligned using a handheld GPS device. Estimates on the maximum orientation error are between 1-2°, based on the accuracy of GPS positioning.

Electric field measurements used non-polarizable electrodes buried at a depth of 20-30 cm in the ground. Permafrost is continuous around Cambridge Bay and the active layer was found to be around 1 m deep when the measurements were made in July 2023. There were no problems placing the electrodes within the thawed, active layer. Typical contact resistances were in the range 0.1-3.0 k Ω . This is lower than expected for a region with extensive ground ice, and may reflect the presence of saline ground water and clay (Minsley et al., 2012).

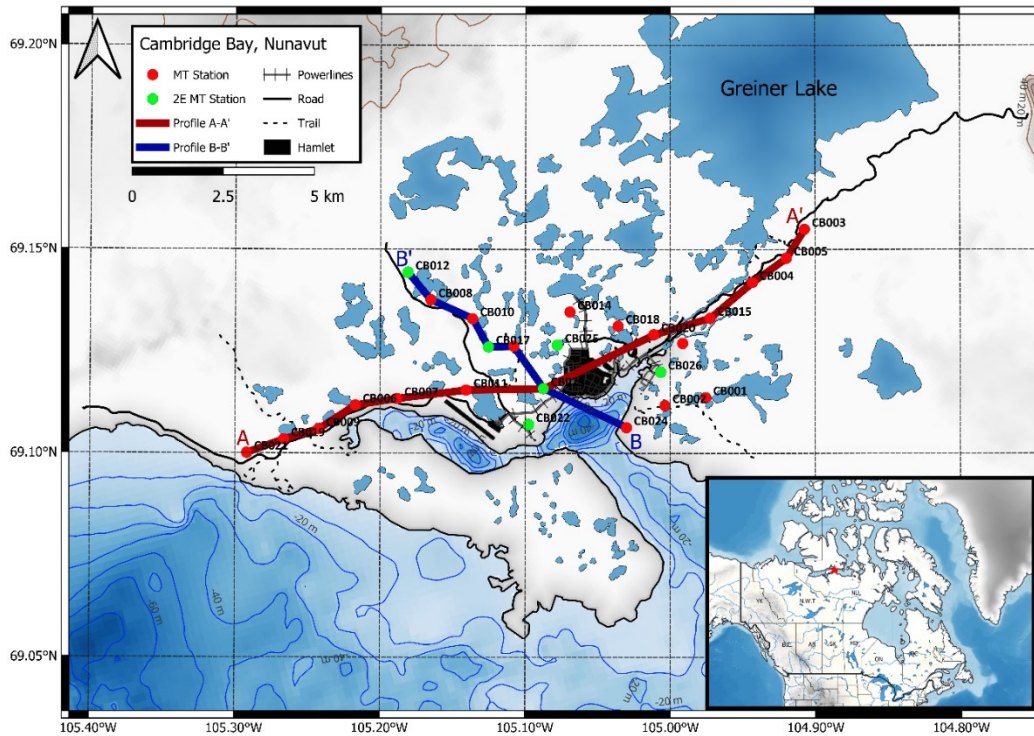


Figure 5.1: Overview map of the MT survey data collected at Cambridge Bay, Nunavut in July 2023. Red dots denote sites where electric and magnetic field data were collected, and green dots where only the electric field was recorded. The red and blue lines show the profiles A-A' and B-B', respectively. These profiles are used for analysis of the inversion models.

5.3. Cambridge Bay electrical dimensionality

As discussed in Chapter 4, there are multiple methods to assess the dimensionality of MT data. For the Cambridge Bay data, the apparent resistivity and phase curves, induction vectors, and phase tensors are used. Cultural noise present near the community increased the difficulty in assessing the dimensionality of these data. Sites that produced higher MT data quality were more likely to be further from the settlement.

5.3.1. Cambridge Bay apparent resistivity and phase

Apparent resistivity and phase are typically among the first aspects of data inspected, after raw time series. It is important to ensure that apparent resistivity and phase are consistent at each station.

Apparent resistivity and phase soundings are highly consistent across the entire survey area. A typical station exhibits a resistivity profile that is in the range of 30 - 100 Ωm at the highest frequencies, and increases to $10^3 - 10^4 \Omega\text{m}$ at frequencies near 1 Hz. Finally, curves tend to show slight decreases towards $10^3 \Omega\text{m}$ in apparent resistivity around 0.1 Hz, the lowest frequencies examined. For the phase, low phase angles are observed at similarly high frequencies, which increase past 45° and then decrease at the lowest frequencies examined and are consistent with the observed apparent resistivity curves. Figure 5.2 shows example apparent resistivity and phase curves for stations CB005 and CB025.

Many stations, particularly those less affected by noise, exhibit apparent resistivity and phase components that are somewhat consistent between xy and yx components. This consistency is predominantly true for high frequency signals above approximately 100 Hz. This indicates one-dimensional electrical structure in the shallow subsurface and this dimensionality may or may not extend to greater depths. For rock with a lower end resistivity of $\sim 30 \Omega\text{m}$ and frequency of 100 Hz, we can estimate a depth of exploration of at least 270 m using the skin depth equation (Eq 4.1).

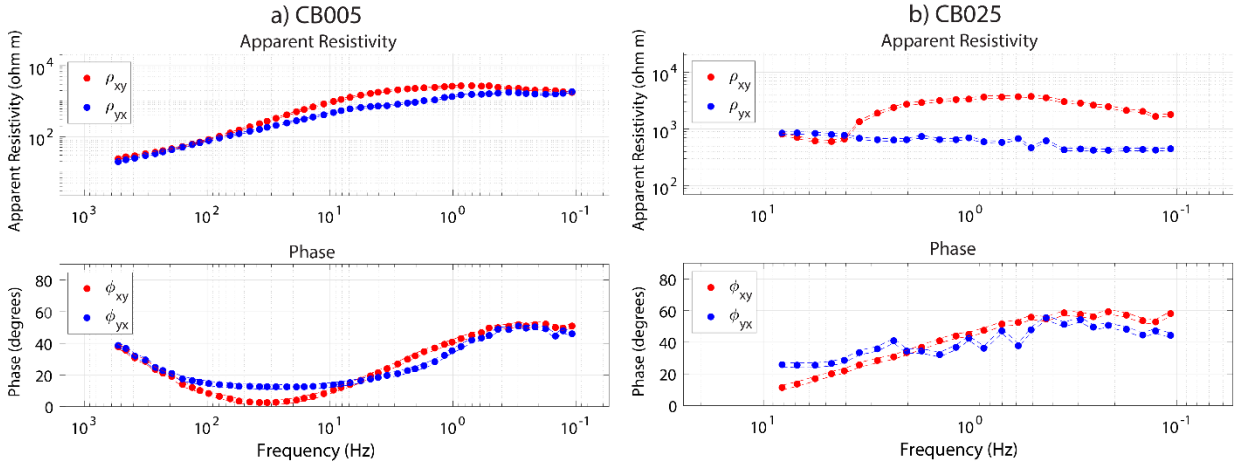


Figure 5.2: Examples of processed apparent resistivity and phase curves: a) station CB005, away from the settlement; b) station CB025, located near to settlement infrastructure.

At frequencies lower than ~ 100 Hz, separation of the xy and yx components may indicate a departure from 1-D structure, but this observation is variable between stations and is highly affected by the noise present. At the lower limit of frequencies observed, $\sim 1 - 0.1$ Hz, sounding curves appear to once again show evidence of 1-D structure. Simple observation of the apparent resistivity and sounding curves is not enough to strongly support a characterization, and is incapable of distinguishing 2-D and 3-D structure. Additional information should be used to assess the dimensionality of these data.

It can be seen in the example sounding curves of Figure 5.2 that cultural noise can severely impact the data quality collected at a station. This noise can produce less smooth results, and apparent resistivity and phase estimates that are not consistent with one another. Post-processing is done to address this, whereby data points that cannot be fit using a D+ model are eliminated (Parker, 1980; Parker & Whaler, 1981; Parker, 1994; Parker and Booker, 1996). The D+ algorithm works by fitting the data using a 1-D model that consists of a finite number of infinitely-thin, infinitely conductive sheets with finite conductance. The D+ conductivity model gives the best possible fit for a given set of apparent resistivity and phase measurements, and is a useful way to determine if the measured apparent resistivity and phase are consistent. Most datasets can be fit by D+, and inability to fit data with a D+ model may indicate errors in data collection or processing, or a unique geology.

Table 5.1 shows how D+ editing can be used to eliminate data points that are inconsistent in apparent resistivity and phase. The root mean square (RMS) misfit is a measure of how well a model fits a given dataset. In this case an RMS greater than 1 indicates that the model does not completely fit the data. An RMS below 1 indicates that the data are being overfit, given the level of noise. An improvement in the RMS misfit for D+ modelling at several stations confirms that removal of certain data points improves the internal consistency of the data.

Station	XY RMS	YX RMS	XY D+ RMS	YX D+ RMS
CB001	6.12	6.3	0.49	0.55
CB002	0.68	1.48	0.22	0.82
CB003	0.65	0.25	0.65	0.25
CB004	0.33	0.2	0.33	0.2
CB005	6.27	6.24	0.29	0.4
CB006	0.28	0.4	0.32	0.37
CB007	0.41	0.56	0.18	0.32
CB008	0.32	0.37	0.38	0.47
CB009	0.18	0.32	0.31	0.31
CB010	0.24	0.49	0.67	0.16
CB011	6.18	5.86	0.63	0.38
CB012	0.31	0.31	0.88	0.23
CB013	6.39	0.38	0.39	0.51
CB014	5.12	1.59	0.72	0.96
CB015	0.88	0.23	0.25	0.65
CB016	4.65	3.2	0.32	1.13
CB017	0.72	0.96	0.69	0.6
CB018	5.8	6.08	0.3	1.7
CB019	0.36	1.39	0.14	0.26
CB020	5.74	5.76	0.86	2.48
CB021	5.62	4.22	0.82	0.67
CB022	0.14	0.26	0.85	0.99
CB023	5.01	3.69	0.4	0.89
CB024	0.99	2.5	0.79	0.61
CB025	0.94	3.2	1.19	0.84
CB026	4.68	4.02	0.73	1.2

Table 5.1: Station RMS misfit to D+ inversion models before and after D+ editing.

5.3.2. Cambridge Bay induction vectors

The vertical component of the magnetic field was measured at several stations around Cambridge Bay to examine whether the tipper is sensitive to subsurface features in this region. These measurements are sensitive to lateral changes in subsurface resistivity and is unaffected by galvanic distortion of the electric field. Tipper measurements and induction vectors were discussed in more detail in Chapter 4.

Figure 5.3 shows induction vector (IV) maps at six frequencies, adopting the Weise convention where the induction vectors point away from conductors. At high frequencies (>10 Hz), IVs are scattered and show little agreement in direction, failing to indicate any lateral changes in subsurface structure. At frequencies less than about 10 Hz, the IVs begin to align, with all vectors having some northward component. This pattern evolves and more strongly shows consistent northward alignment at lower frequencies, and tipper magnitude is also largest at these frequencies. Extremely large tipper at station CB010 is probably caused by high noise content at lower frequencies. It is also likely that the conductive salt water of the Northwest Passage causes this alignment at lower frequencies, as this constitutes a massive conductor adjacent to the survey area that will begin to present itself in data at lower frequencies.

Overall, the induction vectors derived from the Cambridge Bay MT dataset are dominated by coast effects caused by the Arctic Ocean. There is insufficient vertical magnetic data away from the coastline to identify the location of non-ocean structure and data dimensionality.

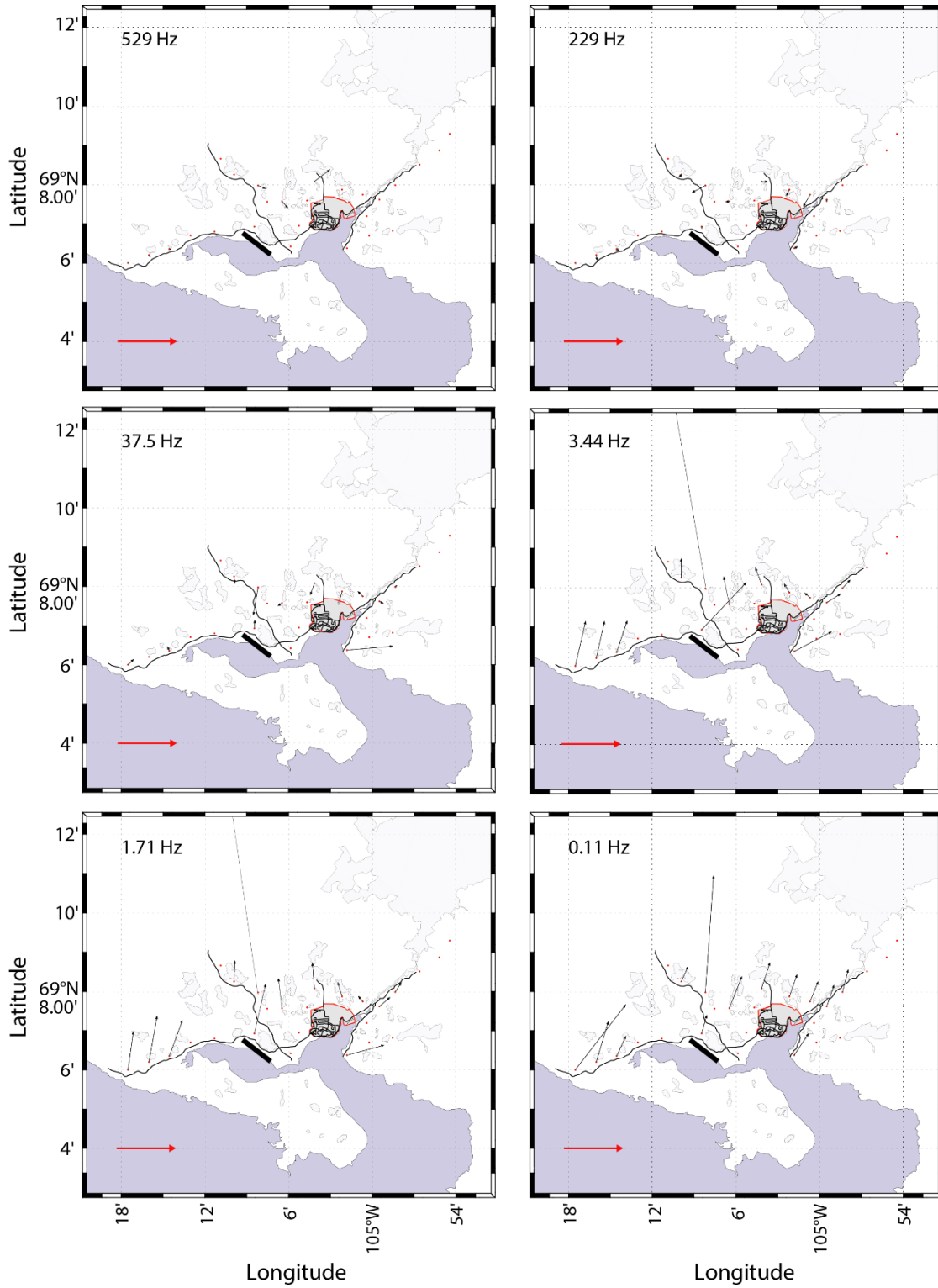


Figure 5.3: Induction vector maps are shown for the Cambridge Bay dataset at six frequencies: 529 Hz, 229 Hz, 37.5 Hz, 3.44 Hz, 1.71 Hz, and 0.11 Hz. The Weise convention is used, where induction vectors point away from conductors. Unit vectors are shown in red

5.3.2. Cambridge Bay phase tensors

Phase tensor analysis (Caldwell, et al., 2004; Booker, 2013) is another common technique for examining the electrical dimensionality of a dataset. This method is again insensitive to galvanic distortion, and, unlike tensor decomposition methods, does not require a 2-D assumption of the impedance tensor (Chave & Smith, 1994). Figure 5.4 shows a phase tensor map of Cambridge Bay at six frequencies. It is important to note that ellipses that are circular and have low skew angle represent 1-D structure; ellipses with low skew represent 2-D structure; ellipses with higher skew angle represent 3-D structure (Figure 4.6).

At high frequencies (529 Hz and 229 Hz) the MT data were relatively 1-D, with phase tensors that were somewhat circular with low skew angles. At intermediate frequencies (37.5 Hz – 1.71 Hz) the phase tensors indicate 2-D or 3-D dimensionality, with bulk alignment of phase tensor ellipses along a SW-NE axis and generally high skew angles. At low frequencies (< 0.1 Hz) the phase tensors become somewhat less elliptical with smaller skew angles, potentially indicating weaker 2-D or 3-D behavior or signaling a return to quasi-1-D structure. These results are consistent with the conclusions obtained from the apparent resistivity and phase curves, and suggest that 1-D modelling of this dataset is a reasonable approach in the frequency band of 529 Hz – 37.5 Hz.

At lower frequencies (< 37.5 Hz) the phase tensors at Cambridge Bay exhibit significantly 3-D behavior and the phase tensor ellipses show strong alignment in the SW-NE direction. This could be attributed to regional structures, including the seawater of the Arctic Ocean. As the MT signals decrease in frequency, they are expected to sense features that are not only deeper, but also at greater lateral distances.

An anisotropic electrical structure could provide an additional explanation for the more complex resistivity structure at greater depths. Directional fracturing or faulting, or similar features with strong orientation could influence estimates of the subsurface resistivity. It is not possible to determine the effects of anisotropy on the MT data, particularly at these lower frequencies where the resolution is relatively low.

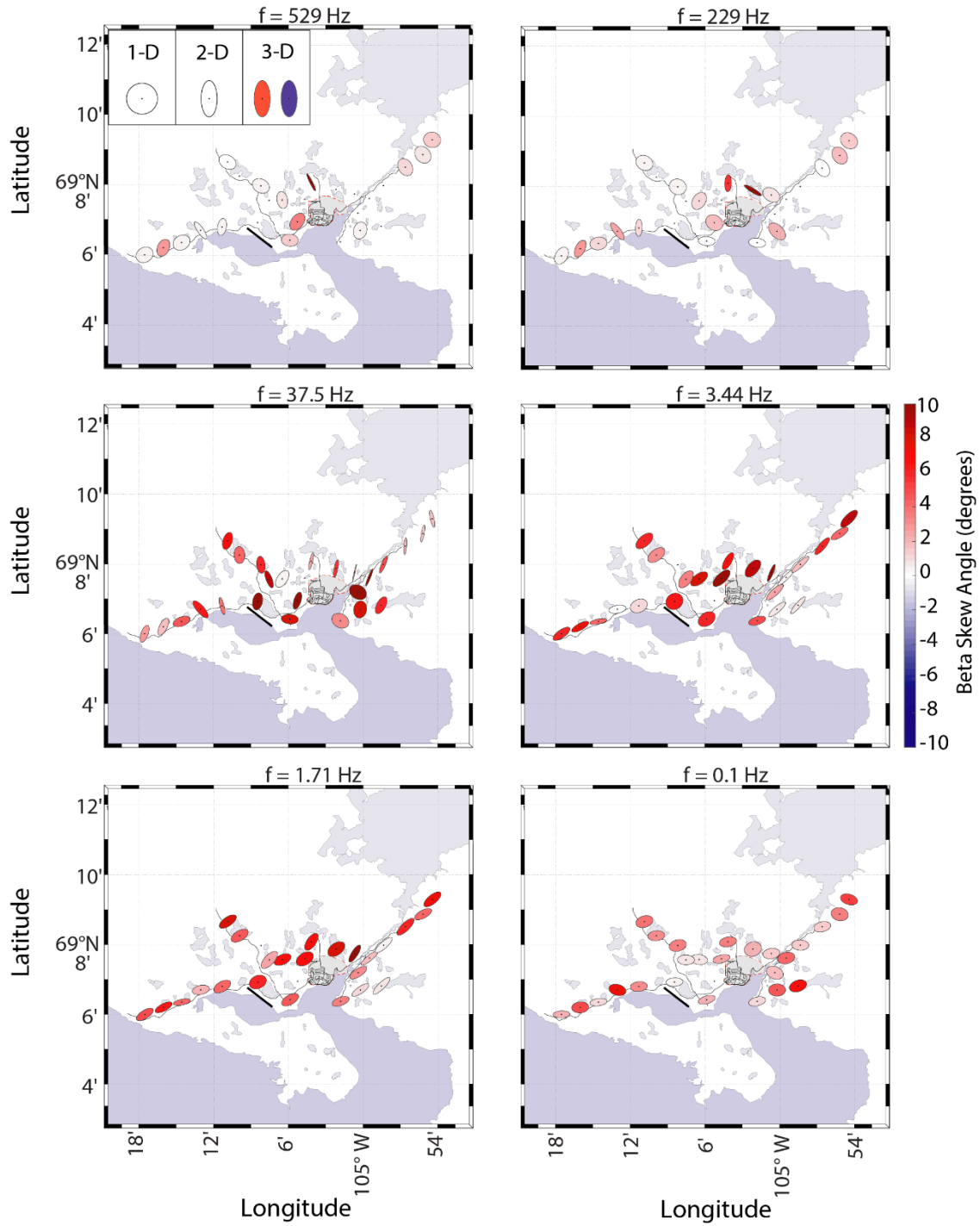


Figure 5.4: Phase tensor map at frequencies of 529 Hz, 229 Hz, 37.5 Hz, 3.44 Hz, 1.71 Hz, and 0.1 Hz. The ellipses show the azimuthal variation in the impedance phase and provide a graphical representation of whether the MT data can be considered 1-D, 2-D or 3-D. It is important to note that ellipses that are circular and have low skew angle represent 1-D structure; ellipses with low skew represent 2-D structure; ellipses with higher skew angle represent 3-D structure.

5.4. Cambridge Bay resistivity modelling

Processed apparent resistivity and phase data are functions of frequency. To interpret these data, we must first convert these to models of true resistivity as a function of depth through the process of inversion. Inversion is a mathematical process whereby a physical model is found that could produce the measured dataset to a certain degree of error. The inversion process is non-unique – there are an infinite number of models that can fit a given dataset. The goal of inversion is to produce a physical model that can fit the measured data, while being geologically reasonable based on prior information about the setting. MT inversion can be performed in 1-D, 2-D, and 3-D.

Based on the dimensionality analysis of the previous sections, three methods were chosen to produce resistivity models from the MT dataset:

- (1) First, station data were converted individually into simple layered models using a manual forward modelling trial-and-error approach.
- (2) 1-D inversions were then performed using the Occam inversion approach of (Constable, et al., 1987).
- (3) Finally, as most MT data are 3-D to some degree, more complex 3-D models were produced using the ModEM inversion algorithm of (Kelbert et al., 2014).

For analysis, the MT data were grouped into two approximately perpendicular profiles, both of which are shown in Figure 5.1. Profile A-A' extends approximately west-east from the nearby gravel pit in the west, through the town site, and towards Mt. Pelly to the East. Profile B-B' has a southeast – northwest trend and roughly follows the DEW line road. Profile B-B' crosses the water of Cambridge Bay between CB023 and CB024. Examination of 1-D inversion models in 2-D model pseudosections was done to show how the resistivity models vary across the survey area.

The RMS misfit quantifies the fit of a model to the measured data. This is a numerical value that compares the observed data to the expected data that would be produced by the model and is given by:

$$RMS\ misfit = \sqrt{\frac{1}{2N} \sum_1^N \left[\left(\frac{\rho_{obs} - \rho_{exp}}{\rho_{error}} \right)^2 + \left(\frac{\varphi_{obs} - \varphi_{exp}}{\varphi_{error}} \right)^2 \right]} \quad (5.1)$$

ρ_{obs} and φ_{obs} are the measured apparent resistivity and phase, ρ_{exp} and φ_{exp} are the expected values apparent resistivity and phase that would be measured from the model. ρ_{error} and φ_{error} are the errors in measured apparent resistivity and phase, and N is the number of observations for both apparent resistivity and phase. A target value for the RMS misfit is 1, representing a model that perfectly fits the data. Larger values reflect models that do not completely fit the data, and smaller values represent models that overfit the data for the level of error present.

5.4.1. Cambridge Bay 1-D magnetotelluric forward modelling

To first investigate how the local resistivity structure responds to a 1-D treatment, simple 3-5 layer resistivity models were produced using a trial-and-error approach. The average of the xy and yx components of measured apparent resistivity and phase values were compared to the actual response of input resistivity models. This average is rotationally invariant, which is expected in a 1-D model. The resistivity model at each station was manually altered until an RMS misfit close to one was obtained with a simple layered model. Due to the poor quality data collected at two stations, simple models could not be found that fit these stations adequately. This manual approach has the benefit of revealing the sensitivity of the MT response to small changes in various model parameters. For example, small changes in shallow layer thickness could result in a large increase of RMS misfit. Conversely, large changes in deeper layer parameters could result in negligible changes to the misfit.

The forward modelling across Profiles A-A' and B-B' show the presence of three distinct vertically separate regions. At the surface, a low-resistivity layer (30 – 100 Ωm) extends to a depth of approximately 100 m. Below the surface layer, the resistivity increases sharply to values in the range of 1000 – 10,000 Ωm to depths on the order of 10 – 20 km. Finally, a third region of intermediate resistivity ($\sim 100 - 1000 \Omega\text{m}$) is suggested by the modelling, however, this feature is poorly constrained by the MT data.

The forward model for station CB023, which lies on both profiles, shows a thin, low-resistivity layer between 1000 and 10,000 m depth. The presence of this layer is not validated by other stations and is likely due to significant noise at this station, and is substantiated by the notably high RMS misfit of this model.

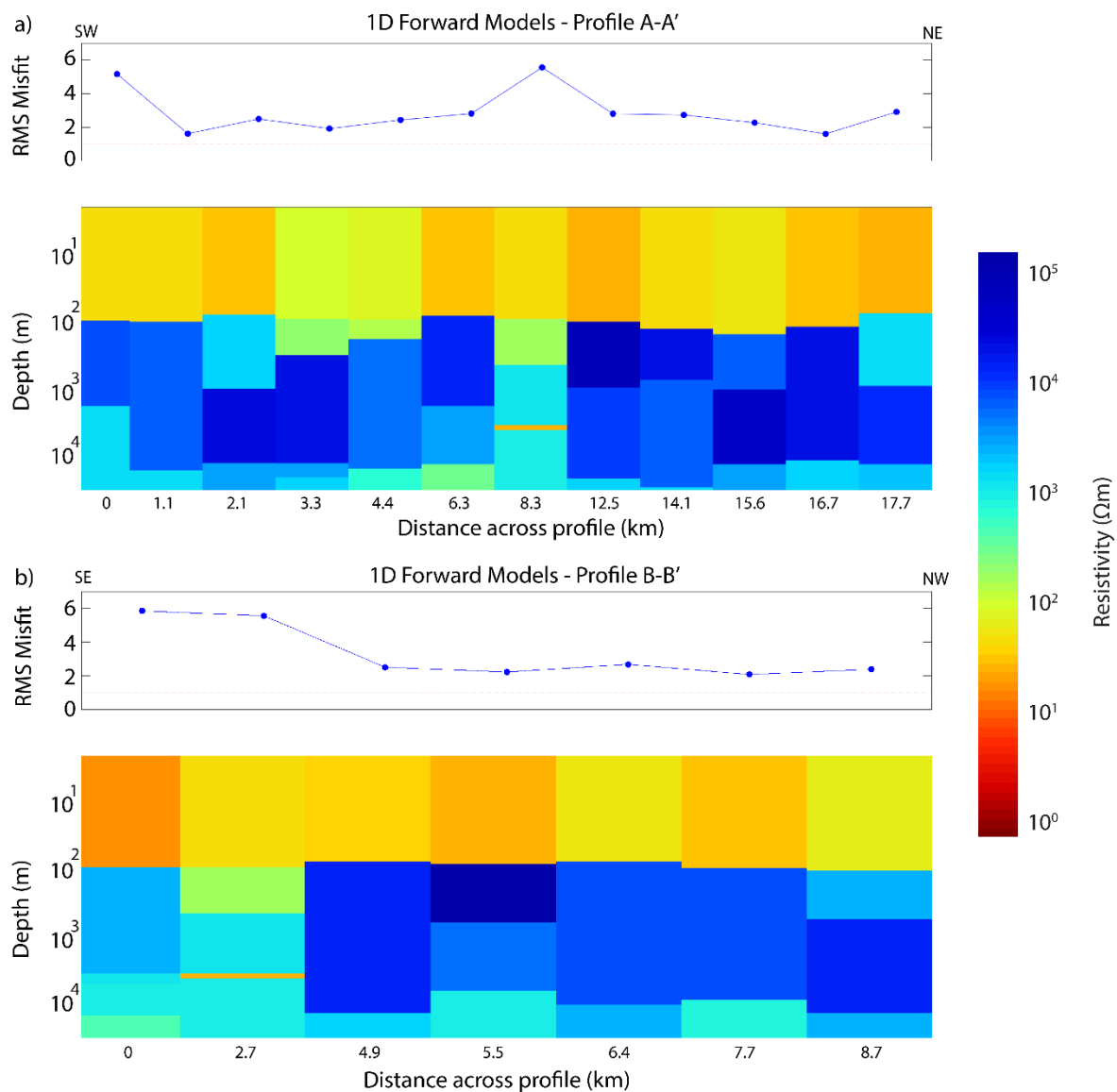


Figure 5.5: 1-D resistivity forward models for profiles A-A' and B-B' obtained by trial-and-error forward modelling. A logarithmic depth scale is used to effectively display layers with greatly different thicknesses. a) 1-D forward models across Profile A-A'; b) 1-D forward models across Profile B-B'.

5.4.2. Cambridge Bay 1-D magnetotelluric inversion

A second modelling approach used the Occam 1-D inversion algorithm of Constable et al. (1987). 1-D inversion is a more sophisticated approach to modelling this dataset than the forward modelling of section 5.4.1. The Occam inversion algorithm produces models that vary smoothly in space, which eliminates arbitrary features in the models caused by data error. Essentially, the Occam algorithm automatically finds the simplest model that fits the data, following the principle of Occam's razor. In this study, 100 layers were used with a 5% error floor for all stations except for CB003, CB021, CB023, and CB024, which required a 10% error floor to adequately fit. Sections depicting the resulting inversion models along Profiles A and B are shown in Figure 5.6.

The results of the Occam 1-D inversions are similar in character to those produced through forward modelling. The model again exhibits three distinct layers: a thin, low-resistivity surficial layer, a higher resistivity intermediate layer, an intermediate-resistivity layer at depth. It is important to note that the Occam algorithm is able to consistently achieve RMS misfit values of ~ 1 , because of the greater number of parameters in the modelling.

Several stations across Profiles A-A' and B-B' exhibit 1-D inversion models that show a significantly thicker surface layer than most other stations. This is likely caused by elevated noise levels in the data, and is evidenced by the generally higher RMS misfit values of these inversion models.

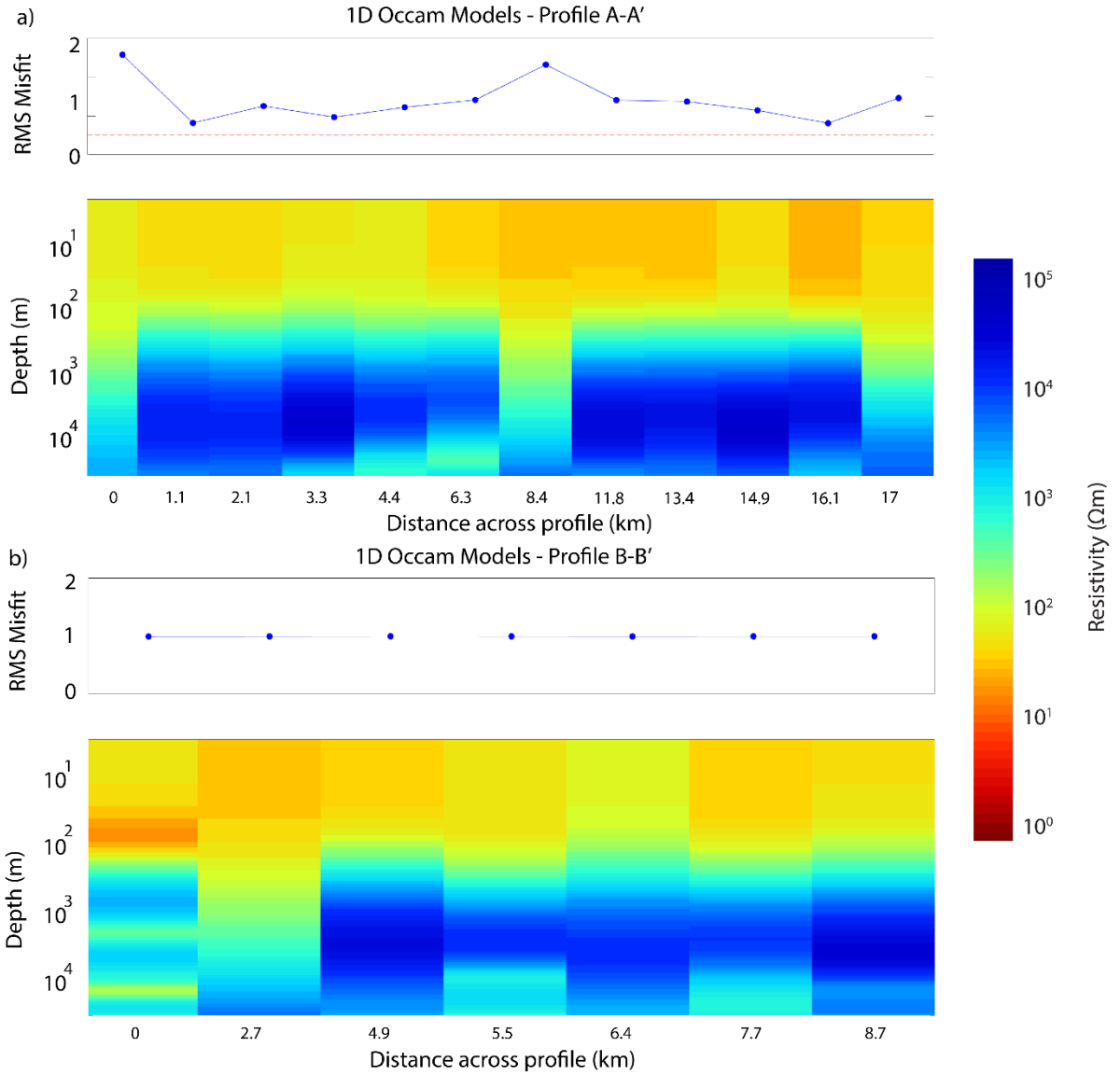


Figure 5.6: 1-D resistivity models for profiles A-A' and B-B' obtained by inversion using the 1-D Occam inversion algorithm. A logarithmic depth scale is used to effectively display layers with greatly different thicknesses. a) 1-D inversion models across Profile A-A'; b) 1-D inversion models across Profile B-B'.

5.4.3. Cambridge Bay 3-D magnetotelluric inversion

While dimensionality analysis shows that the data collected at Cambridge Bay in 2023 was relatively 1-D, most data is 3-D to some degree. 3-D inversion is a more complex task that involves significantly more computational power, but is capable of inverting an entire dataset altogether to produce models that are more accurate to the true subsurface. Additionally, using all stations simultaneously offers more constraints for the inversion process. Care must be taken when interpreting 3-D inversion models, as features may be present in locations that are not well covered by existing data.

3-D modelling was performed for the Cambridge Bay dataset using the ModEM inversion algorithm of Kelbert, et al., (2014). Due to the non-unique nature of the inversion process, a series of inversions were performed on this dataset, each with different model parameters. Parameters that were varied between inversions include: model grid size, starting model, model covariance (smoothing), regularization parameters, and the inclusion of bathymetry and topography. The model covariance parameters control the smoothness of the inversion model, and can be controlled for all three spatial directions. The inversion regularization parameter controls the trade-off between fitting the data and model simplicity. A total of 15 final 3-D inversion models were produced, with the preferred model being *CB_inv13*. Topography and bathymetry data from the IBCAO database were included in several inversions (Jakobsson, et al., 2022).

The model space used for MT inversion includes a model grid composed of two regions, a model core surrounded by padding cells. The station locations are within the core of the grid and are the smallest grid cells at the surface. Padding cells are large cells that increase in size away the centre of the grid and surround the model core. Padding cells are required in modelling low frequency signals that have large skin depth (penetration depth). Since a large grid extent is required, padding cells that increase geometrically are used to fulfill this requirement without drastically increasing the total number of model cells. A depth extent is typically chosen that is approximately three times the estimated skin depth at the lowest frequency. Additionally, an initial layer thickness was chosen that was small enough to model the highest frequency signals with the smallest skin depth. The grid spaces of the first few models were chosen such that stations were separated by one or more cells. Subsequent models used slightly different cell sizes to observe the effect on the final inversion models. For certain models, topography and bathymetry data were included in the

modelling. Topography and bathymetry data from the IBCAO database (Jakobsson, et al., 2022) were included in four inversions with reduced data content, where cells that correspond to seawater were given resistivity values of 0.2 Ωm .

The data collected at all 26 stations at Cambridge Bay were of generally good quality, and so 3-D inversions included every station. All inversions used data at 50 frequencies from 529 – 0.1 Hz. A 10 % error floor was applied to the impedance to ensure that certain frequencies were not given small weights due to unreasonably high estimates of the error.

The ModEM 3-D inversion algorithm (Kelbert et al., 2014) works by iteratively adjusting the model and then calculating the RMS misfit between the observed data and the data predicted by the new model at each step. The choice of an initial starting model, which is a parameter of the inversion algorithm, can play an important role in the final inversion model. If prior geological information is known, a starting model that incorporates this information can be used to influence the inversion process. Two different starting models were used. The first was a simple 100 Ωm half-space. The second used a 10 Ωm surface layer that was 100 m thick, followed by 9.9 km layer with a resistivity of 1000 Ωm , all overlying a 100 Ωm half-space. The non-half-space starting model was chosen based on the features that were identified in 1-D modelling and preliminary 3-D inversion models. The choice of starting model did not drastically change the main features of the inversion model, however, in some instances aspects of the starting model could be seen in the final model.

The preferred inversion model, *CB_inv13*, used 54 model cells in the x -direction, 100 cells in the y -direction, and 90 cells in the z -direction. Each interior model cell was 200 m wide in both the x and y -directions, and 8 padding cells were used along each boundary of the model with a padding factor of 1.3, corresponding to a lateral model extent of 20 km x 29.2 km. In the z -direction, the initial model layer had a thickness of 1 m at the surface, which increased by a geometric factor of 1.15 at each layer beneath. Model covariance values of 0.5 were used for the x and y -directions, and 0.3 for the z -direction to prevent unnecessary vertical smoothing as features at Cambridge Bay were assumed to be vertically stratified. An inversion regularization parameter of $\lambda = 1$ was used for inversion *CB_inv13*. Topography and bathymetry information was also included in this inversion. The inversion for *CB_inv13* attained a final RMS misfit value of 0.92 after 122 iterations. This inversion used a 100 Ωm half-space as the starting model.

The inversion model of the preferred model *CBinv_13* was analyzed in a series of horizontal depth slices (Figure 5.7), as well as along Profile A-A' and Profile B-B' (Figure 5.8).

It can be seen in Figures 5.7 and 5.8 that the 3-D MT inversion model consists of a thin, layer of low-resistivity ($\sim 1\text{-}100\ \Omega\text{m}$) that is continuous across the survey area. This feature is nearly uniform across both profiles A-A' and B-B', with a thickness of approximately 100 m. There is minor variation in the extent of the surface layer in the inversion model – in particular, the model appears to thicken towards the east side of Profile A-A'. Below the surface layer, the resistivity of the model increases drastically by several orders of magnitude to values between $1000\text{--}10,000\ \Omega\text{m}$. This high-resistivity zone transitions back to $100\ \Omega\text{m}$ (the starting model resistivity) at depths of 10 - 20 km. Along Profile B-B', the transition from low to high resistivity extends much deeper between 1-2 km along the profile. This feature is largely consistent with where the profile crosses the seawater of Cambridge Bay. Overall, the resistivity structure of CB013 can be summarized using a 2-layer model: a 100 m thick surficial layer of relatively low resistivity (CB1), and an underlying high-resistivity layer that extends to greater depths (CB2).

Figures 5.9 – 5.11 illustrate how well the inversion model fits the data. Figure 5.9 displays the RMS misfit by station. Figure 5.10 shows the RMS at each station and frequency along a W-E profile. Figure 5.11 shows the difference between the collected data and those predicted by inversion model *CB_inv13*.

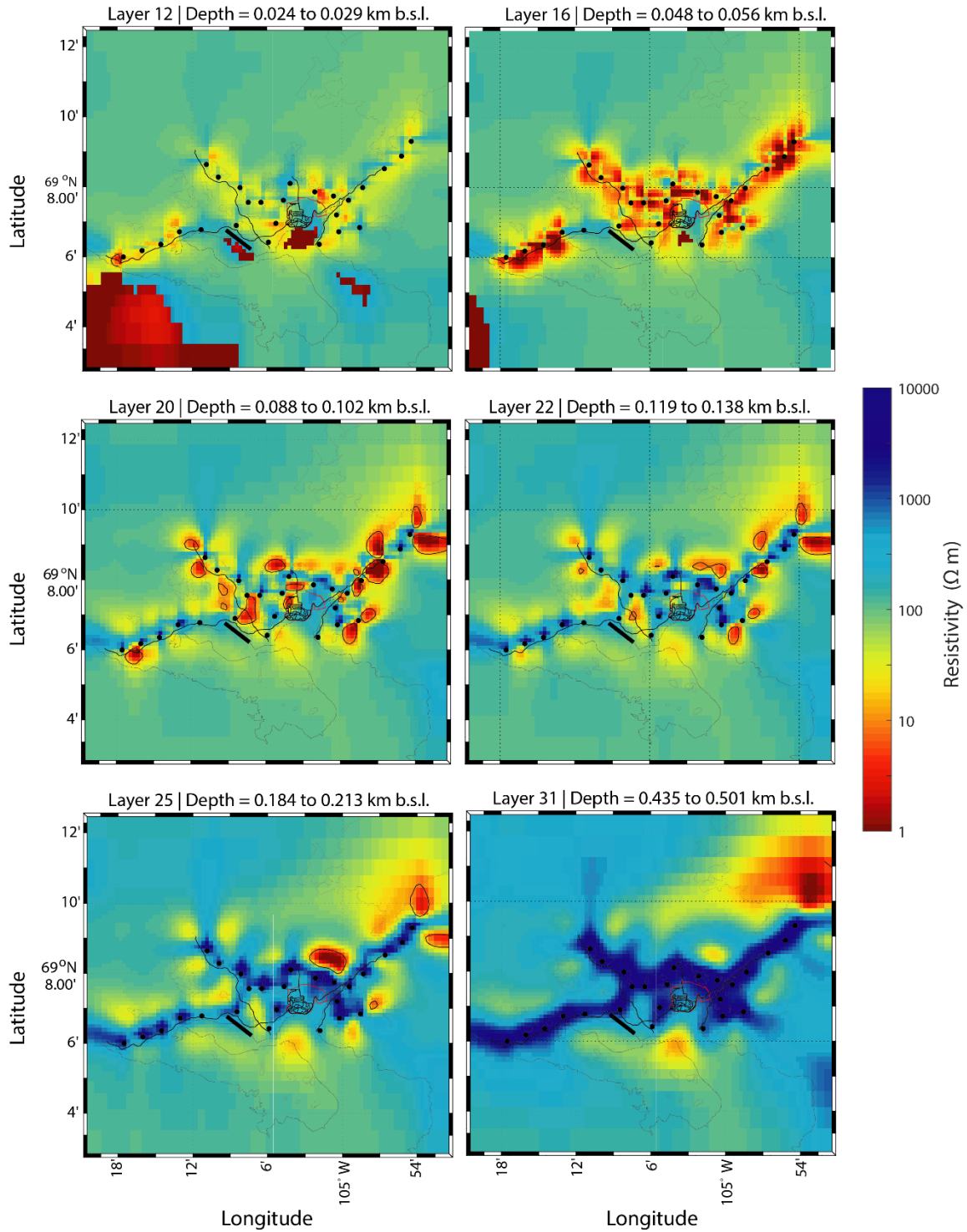


Figure 5.7: Horizontal sections of the 3-D resistivity model *CB_inv13* between 24 m and 501 m. Model layer thickness increases geometrically with depth by a factor of 1.15 per layer. Seawater with resistivity of 0.3 $\Omega \text{ m}$ is included in the model based on Arctic bathymetric models.

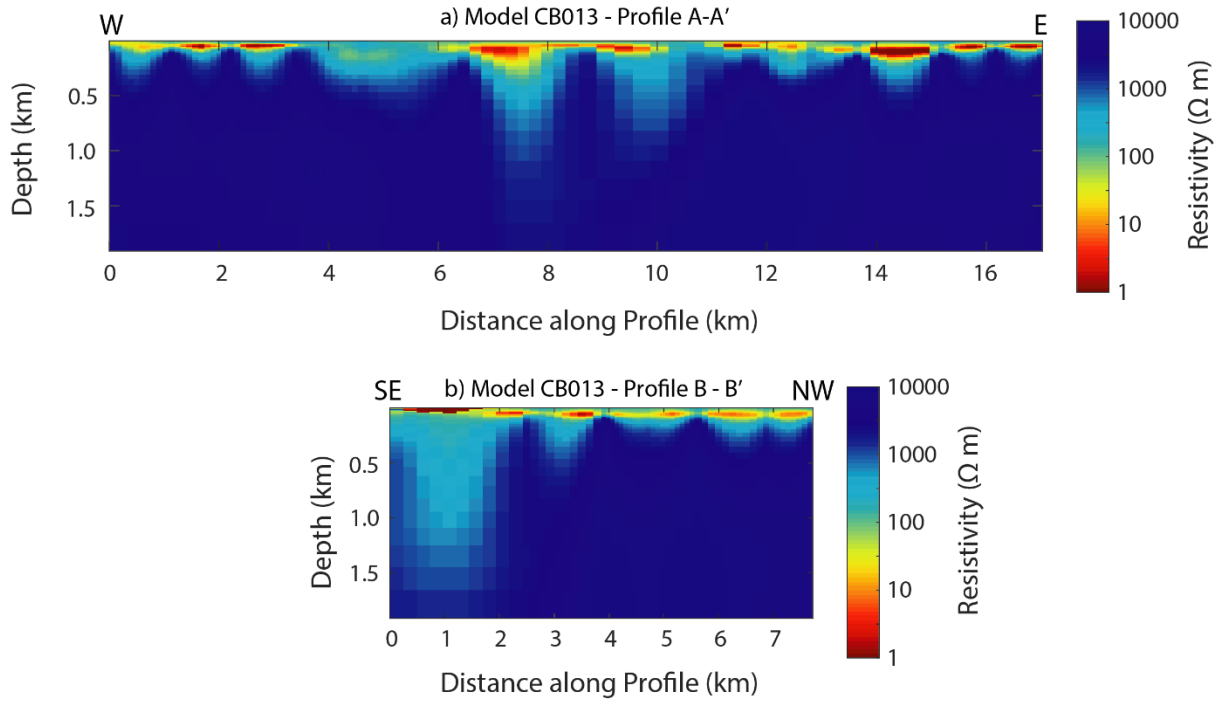


Figure 5.8: Vertical sections of the 3-D resistivity model *CB_inv13*: a) along Profile A-A' to a depth of 2 km; b) along Profile B-B' to a depth of 2 km. A vertical exaggeration factor of 2 is applied to the sections.

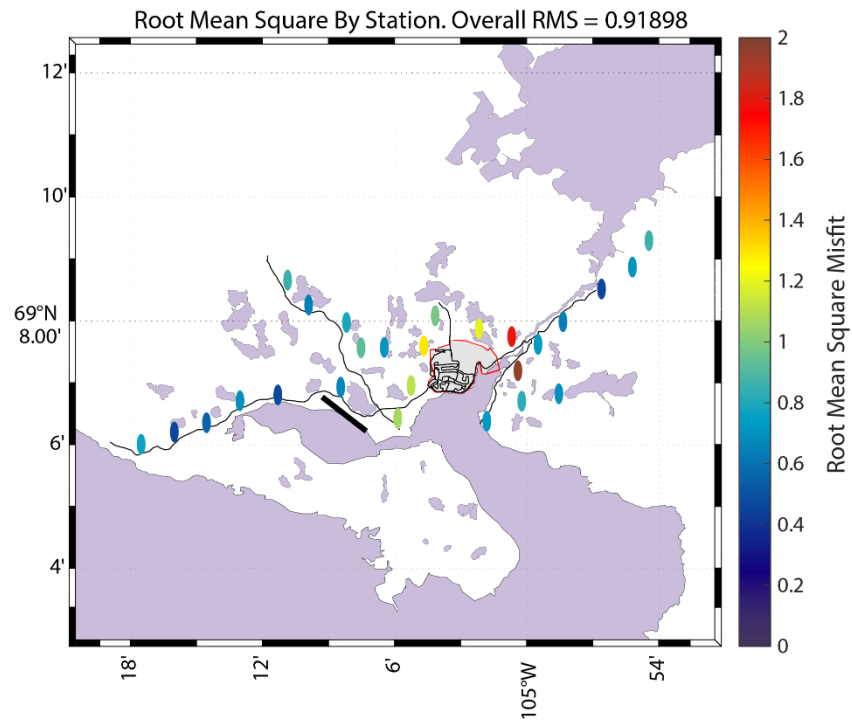


Figure 5.9: A map displaying the RMS misfit by station for inversion model *CB_inv13*.

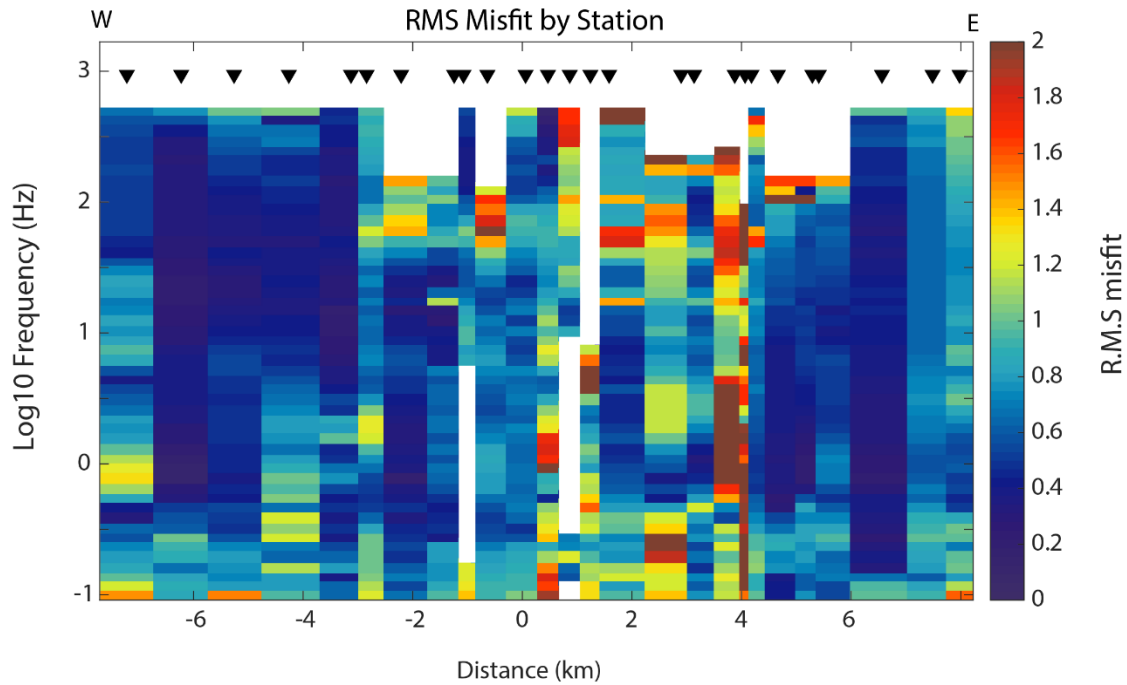


Figure 5.10: A pseudosection displaying the RMS misfit by station from W-E for inversion model CB_inv13. An RMS value of 1 indicates an ideal fit, with values greater than 1 indicating data that is being underfit by the model, and values less than 1 indicating data that is being fit too closely, given the estimated error. White blocks on the profile represent data that were removed during D+ editing of processed sounding curves.

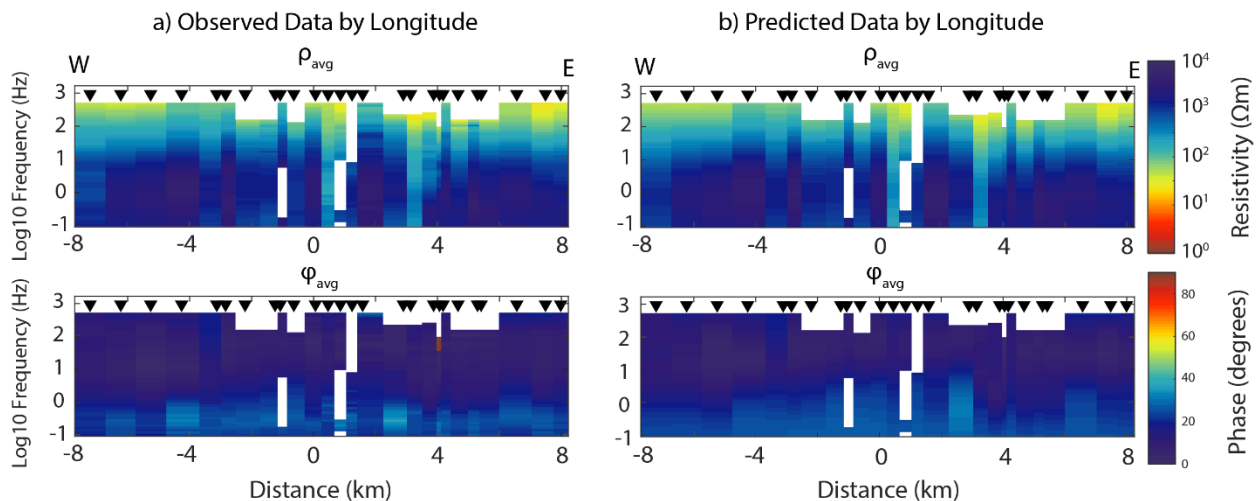


Figure 5.11: a) Pseudosections of apparent resistivity and phase projected along profile. a) observed data from W-E; b) the predicted response of inversion model CB_inv13 from W-E.

The resistivity model *CB_inv13* is chosen as the preferred model for several reasons. For one, the model is strongly consistent with previous modelling; the simple 2-layer structure is evident in both previous 1-D modelling, as well as virtually all other 3-D inversion models. Additionally, the use of a 100 Ωm half-space for *CB_inv13* led to features in the final resistivity model that are realistic. Other inversions that used artificial starting model constraints often contained features that appeared unnatural. A final inversion model was sought which did not show features that could be clearly attributed to the choice in starting model, but instead features that were strongly supported by the data. Finally, *CB_inv13* did not show or did not strongly present features thought to be inversion artifacts or caused by noisy data. For instance, previous inversions presented deeper, low-resistivity features that are inconsistent, or not present in all inversions. These features are not as prominent in resistivity model *CB_inv13*.

It can be seen in Figures 5.9 and 5.10 that most stations that exhibit the highest RMS misfit with the predicted response are those situated near to the high voltage power line along the main highway or other infrastructure in the area. Stations that were less affected by cultural noise were generally of better quality and were easier to fit with the 3-D inversion algorithm. Additionally, stations farther from the center of the inversion model typically had larger RMS misfit values at lower frequencies, whereas stations closer to the center had larger RMS misfit values that were more evenly distributed across all frequencies. This is likely to be caused by the increased noise levels recorded at stations near the settlement, which is more likely to have high frequency content.

Figure 5.11 shows a comparison of the observed data and the data predicted by *CB_inv13*. This illustrates the similarity between the predicted data from *CB_inv13* and the data that were actually observed at Cambridge Bay. The model produces a response that is very similar to the original data, although small differences exist.

5.5. Analysis of 3-D model sensitivity

The 3-D model produced by inversion was analyzed to determine the sensitivity of the main model features. Multiple approaches exist for sensitivity analysis, however the approach used here involves editing the model at various depths and examining how the RMS misfit changes. Here, the resistivity model *CB_inv13* was edited to add a half-space to the model at various depths in the range 0.5 – 50 km. A summary of the sensitivity analysis is displayed in Figure 5.12.

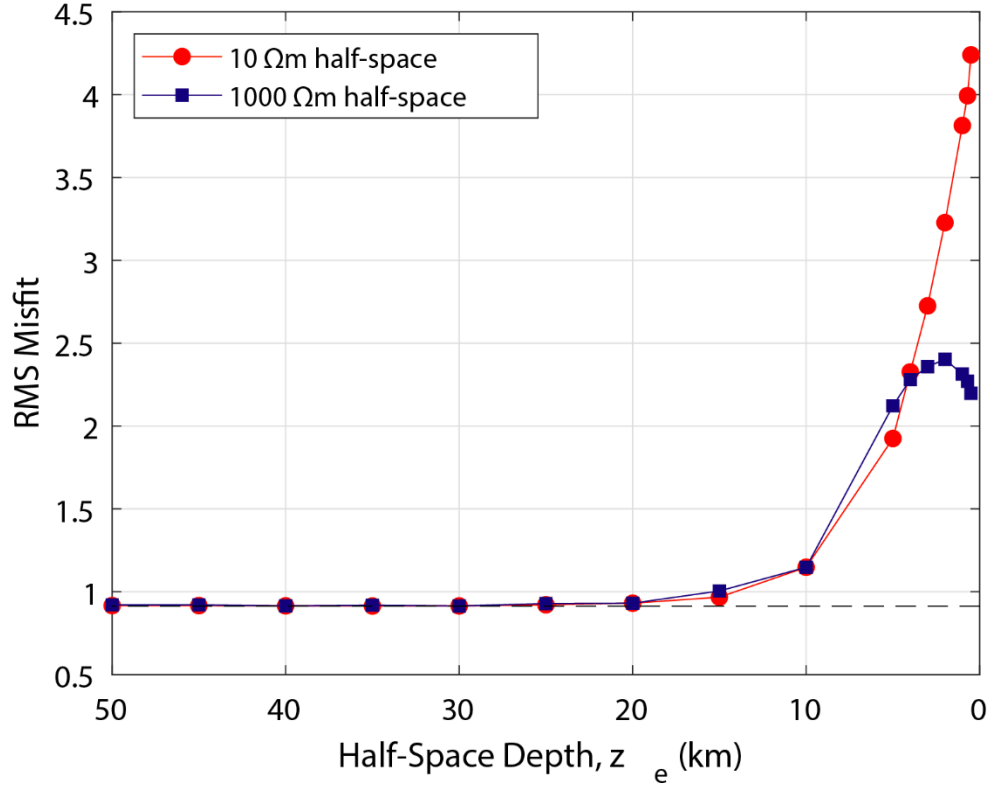


Figure 5.12: The RMS misfit when the resistivity of the inversion model, *CB_inv13*, was changed to a half-space below a depth, z_e . Analysis was done using a half-spaces with $\rho = 1000 \Omega\text{m}$ and $\rho = 10 \Omega\text{m}$. The dotted line shows the RMS misfit of the unedited *CB_inv13*.

The results of sensitivity analysis show that changes to the original model below approximately 20 km had essentially no effect on the fit of the model to the original data. For shallower depths above 20 km, the RMS misfit increased for both the 1000 Ωm and 10 Ωm half-spaces, with the most pronounced effects at depths below about 10 km. The key features of inversion model *CB_inv13* lie predominantly within the first few hundred meters, with the deepest features around depths up to 1.5 km (Figure 5.8). These results suggest that the MT data in the frequency band of 529 – 0.1 Hz were sensitive to depths far greater than those identified in the previous section. Ultimately, CB1 and CB2 were well constrained by the measured data.

Interestingly, the RMS misfit of the edited models began to decrease when the half-space was added below very shallow depths (< 1 km). This suggests that the feature of the original model that is seen extending from the bay to depths of ~ 1.5 km is likely not real, and may be an artifact

produced by the inversion modelling algorithm and the low-resistivity seawater overlying this feature.

5.6. Interpretation of MT inversion models

It was shown in Chapter 3 that the resistivity of rocks near the surface is highly sensitive to temperature, and whether they are frozen or unfrozen. As rocks freeze, the resistivity becomes more sensitive to changes in temperature. Additionally, the salinity of fluids in rock pore space has a substantial effect on the bulk resistivity. In effect, frozen rocks do not necessarily have high electrical resistivity. Accurate interpretations of resistivity models that include frozen and unfrozen rock requires information about the temperature. For this analysis, the temperature gradient is taken to be linear, with a surface temperature of -10°C , based on the MAGT at Cambridge Bay. A geothermal gradient of $20^{\circ}\text{C}/\text{km}$ is adopted based on values expected of the Arctic sedimentary basin (Minnick, et al., 2018).

5.6.1. Low-resistivity near-surface layer (CB1)

It was apparent during MT data collection in July 2023 that a low resistivity layer was present at all measurement locations in the survey. Initially, it was assumed that this feature was a layer of glacial till (Sharpe, 1993; Harrison et al., 2013) overlying the resistive carbonate rocks, however, this explanation has two problems. Firstly, the resistivity models show the low resistivity layer extending to depths of approximately 100 m, which is an order of magnitude deeper than the depth extent of the mapped layer. Secondly, the low-resistivity layer is present at all regions sampled, including those without till, till veneer and till blanket (Sharpe, 1993; Harrison et al., 2013).

An alternate explanation for the existence of a low-resistivity layer around Cambridge Bay is the presence of saline groundwater in the near surface permafrost. The surface of Cambridge Bay was submerged in seawater for a significant part of the Holocene, allowing saline water to infiltrate the surface layer of rock. There is potential that seawater has diffused into the pore space of the surface rock, including the glacial till and underlying carbonate rock. It is common to use Archie's Law to assess the effects of pore fluid on resistivity, however, this approach cannot be used in the case of frozen rock. Instead, the experimental results of Pandit and King (1979) are used to relate the fluid content of sedimentary rock to bulk resistivity. Using data presented for a sample of Salem limestone, estimates of the porosity of the frozen surface layer CB1 were made. It was assumed

that the temperature in the upper 100 m is approximately -10°C , and that the porosity is $\sim 13\%$, the same as the experimental sample of Pandit and King (1979). Using a bulk resistivity value of $\sim 1 - 100\ \Omega\text{m}$, the Pandit and King (1979) results can be used to estimate a salinity in the range of 30 - 40g/L (Figure 5.13).

To verify whether it is reasonable for seawater to diffuse to depths of $\sim 100\text{ m}$ in the permafrost layer over the interglacial period, it is important to use additional numerical tests. These tests are outlined in Chapter 7.

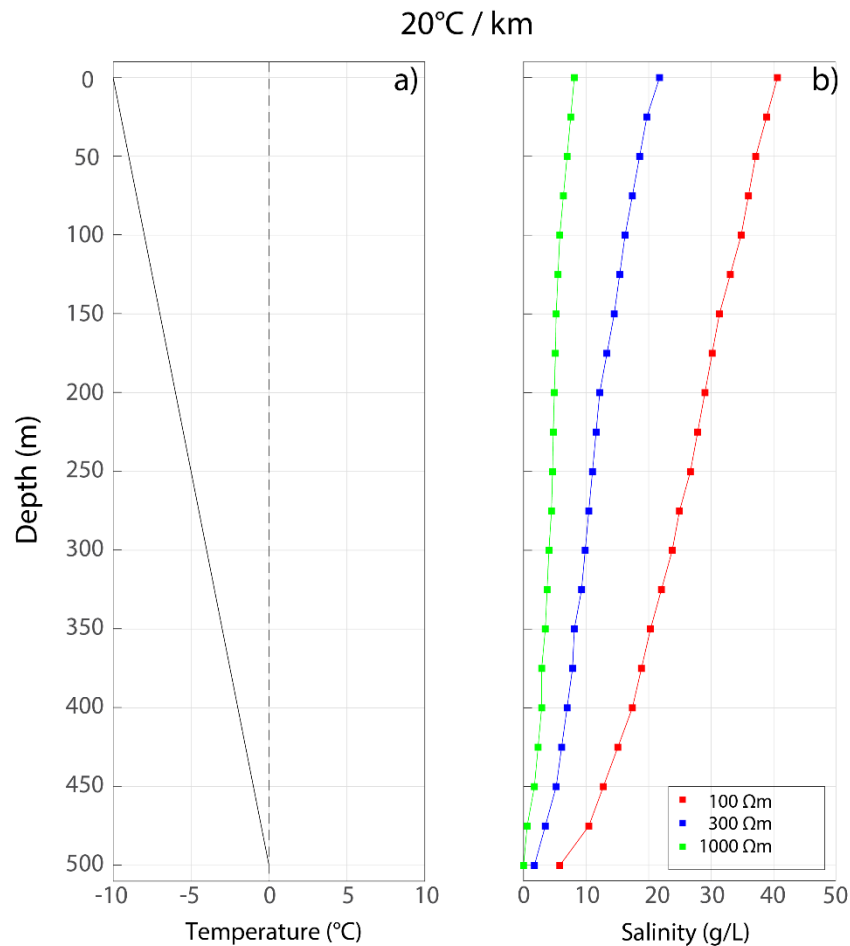


Figure 5.13: a) A simple linear geotherm derived using the mean annual ground temperature (MAGT) and estimated geothermal gradient at Cambridge Bay; b) Estimates of subsurface salinity obtained from the MT data. A porosity of 13% was used based on the Salem limestone sample of Pandit and King (1979).

5.6.2. High-resistivity layer (CB2)

The second layer is highly resistive and it is likely that the several distinct lithologies are responsible for the high resistivity. A reduction in salinity and porosity with depth can be attributed to the increase in resistivity within this region.

(1) The upper part of this layer is composed of frozen Paleozoic carbonate rocks of the Arctic platform (Harrison et al., 2013). The temperature will be below 0°C to a depth of about 500 m based on the geotherm plotted in Figure 5.13. The resistivity studies of Pandit and King (1979) on limestone samples suggests that the salinity of this layer is very low. For example, a bulk resistivity of 1000 Ωm at a depth of 200m will require a salinity of less than 5 g/L. At a depth of 500 m the required salinity is less than 1 g/L.

(2) The middle part of this layer is composed of unfrozen carbonate rocks. The depth at which 0°C is reached will depend on the geothermal gradient. With a low estimate of 20°C/km this depth will be 500 m. Higher geothermal gradients will correspond to shallower depths of the permafrost base. Archie's Law can be used to estimate the salinity and porosity of this layer. To produce a bulk resistivity of 1000 Ωm the required porosity is less than 10% for all salinities in the range 1 – 30 g/L.

(3) The lower part of this layer is composed of unfrozen basement rocks. Based on regional geological mapping across Nunavut, the carbonate sequence at Cambridge Bay is believed to be several kilometres thick (Dewing et al., 2019). The underlying crystalline basement rocks are of Archean or Paleoproterozoic age and expected to be generally resistive.

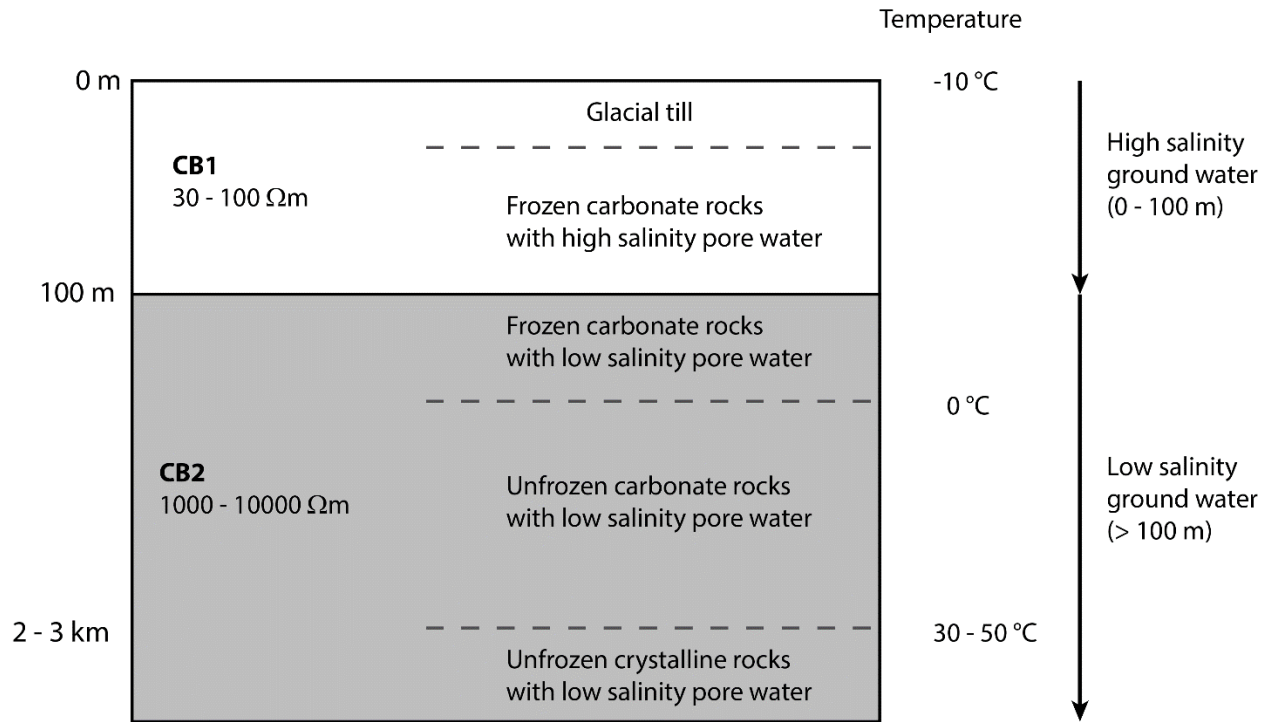


Figure 5.14: Summary of near surface resistivity structure at Cambridge Bay, based on resistivity models derived from magnetotelluric data.

5.7. Discussion and Implications for Geothermal Energy Development

The modelling of MT data collected at Cambridge Bay results in a resistivity model composed of a 100 m thick low-resistivity layer (CB1) at the surface that overlies a relatively homogeneous region of significantly higher resistivity (CB2). Results from 1-D forward modelling, 1-D inverse modelling, and 3-D inverse modelling are consistent with one another and reinforce the notion that the low-resistivity surface layer (CB1) is present throughout the survey region, and is of remarkable consistency in thickness. These results are somewhat unexpected – typical resistivity models of regions with permafrost will indicate a region of high-resistivity overlying one of much lower resistivity. Here, the model is the opposite structure, where a portion of the permafrost structure is of significantly lower resistivity than expected. It is thought that the low-resistivity permafrost of the surficial layer is caused by water of relatively high salt content that has infiltrated the subsurface rock. This is likely to have occurred during a period of inundation at the onset of the Holocene. Diffusion of salt-water from the surface would likely be a consistent process across the surface, and so would produce a layer that is of relatively uniform thickness across the region. Further investigation of this structure could be used to check the veracity of this interpretation of

the low-resistivity surface layer. The uniform topography of the area around Cambridge Bay may have affected how this structure formed. It is unknown to what extent this structure extends, but may differ at further regions where the surface elevation differs more significantly. Additional study of the broader region surrounding Cambridge Bay, and Victoria Island as a whole, could provide important insight into this effect. A number of additional shallower geophysical studies could complement this study. DC resistivity and TDEM methods could be used to effectively study this near surface layer at a greater resolution. Induced polarization could also be used if the polarizability of the frozen and unfrozen rocks are sufficiently different. Sampling of the salinity of near-surface rocks may be the most cost-effective way to validate the results of this study.

MT modelling does not conclusively indicate the change from frozen to unfrozen material, nor is there any evidence of other notable features such as subsurface aquifers. Local permafrost likely extends to several hundred meters in depth. Some 3-D inversion models show conductive features extending to depths of one or more kilometers in depth, however, these are unlikely to be real features, or may be extensions of surface features that are maintained by the inversion algorithm. The data collected are unlikely to be very sensitive to structure below several kilometers depth. Additionally, the MT models are incapable of resolving the glacial sediments that exist at the surface. Resolution of this shallow layer would likely require higher frequency signals than those measured in this study.

The target depth for geothermal development is within the upper 1 – 2 km. In this depth range the rocks have mostly high electrical resistivity and are likely carbonate rocks of the Arctic platform. The lack of conductive zones beneath the surface in this region indicates the absence of zones with elevated rock porosity or water content. It is also significant, because the high-resistivity in the upper 1-2 km implies the absence of zones of saltwater intrusion, as might be expected in a coastal environment. Overall, based on the evidence supplied by this MT study, the prospects for conventional geothermal systems are low. There is little evidence of elevated temperatures, fracturing and porosity, or water content at depths ideal for drilling. Alternatively, engineered geothermal systems could be developed to artificially elevate porosity and to make use of low-grade geothermal resources for direct-use applications in the community. Such infrastructure could sizeably offset local fossil fuel consumption and provide a more sustainable energy source.

Chapter 6: Magnetotelluric Study at Resolute Bay, Nunavut

6.1. Introduction

The second part of the geothermal investigation took place in Resolute Bay, Nunavut, in August 2023. The goals of this survey were identical to the Cambridge Bay survey, i.e. to explore the subsurface resistivity structure beneath the community, identify features of interest in the subsurface, and assess the suitability of the region for geothermal development. The dataset collected at Resolute Bay was analysed in a similar manner as Cambridge Bay. Key differences are described below.

Resolute Bay is situated in the high Arctic, on the southern coast of Cornwallis Island at a latitude of approximately 75°N. Local infrastructure consists of sparsely distributed buildings and structures within an area of approximately 30 km². The settlement is roughly separated into three distinct regions connected by roads and high-voltage transmission lines:

- The area surrounding the Resolute Bay airport and power plant, including the Polar Continental Research Project (PCSP), lies to the northwest.
- The settlement of Resolute houses most permanent residents and municipal services and lies along the northern coast of Resolute Bay.
- Fuel services and storage are located along the western side of the bay.

High-voltage power lines connect these three regions and generates a significant amount of electromagnetic noise. This makes the collection of high-quality MT data challenging. Interpretation of resistivity models requires knowledge of the local geological environment and thermal regime. Cornwallis Island is located within the Arctic Platform where a sequence of Paleozoic rocks some 2 – 3 km thick overly Pre-Cambrian crystalline basement rocks (Thorsteinsson, 1988, Dewing et al., 2019). The surface geology of southern Cornwallis Island is characterized by sedimentary rocks of the Read Bay and Allen Bay formations (Thorsteinsson & Kerr, 1965; Thorsteinsson, 1988; Harrison, et al., 2015). These formations are primarily composed of Paleozoic limestones and porous dolomites, with minor shale, siltstone, and sandstone components. In many areas around the community, a thin layer of gravel no more than 1 - 2 m thick overlies these rocks.

The active layer thickness (ALT) at Resolute Bay is approximately 1 m (Ran et al., 2022) which was confirmed during installation of MT instruments in August 2023. The MAGT at Resolute Bay is -10°C (Natural Resources Canada, 1995; Ran et al., 2022). There is limited knowledge about subsurface temperature in Nunavut (Minnick et al., 2018). Temperature data obtained as part of a diamond-drilling program at Resolute Bay suggested anomalously high temperature gradients near the surface of nearly $40^{\circ}\text{C}/\text{km}$, and a permafrost thickness of ~ 400 m (Misener, 1955). Technical challenges with the drilling program, and thermal insulation effects caused by the coastline provide possible explanations for the locally elevated thermal gradient, and are consistent with the regional gradient being more typical of the Arctic platform (Bremner, 1953; Lachenbruch, 1957). A geothermal gradient of $20\text{-}30^{\circ}\text{C}$ per kilometer is similar to other measurements on the Arctic platform, and predicts a permafrost thickness of about 300-500 m.

6.2. Magnetotelluric data collection at Resolute Bay

In 2022, MT data were collected at 11 sites by a field group from the University of Alberta from June 5-11. In 2023, MT data were collected at an additional 33 sites by field personnel from the University of Alberta and Simon Fraser University from August 4-14. In both deployments, Phoenix MTU-5C instruments were used. An overview of the Resolute Bay survey is presented in Figure 6.1. At 30 stations the instruments recorded both electric and magnetic field data. Data collection was focused on areas close to the settlement, where the development of geothermal infrastructure would be ideal. At 14 stations only electric fields were recorded. Synchronous magnetic fields from nearby MT stations were used to process these stations and compute electromagnetic impedances. Installation of magnetic sensors is a laborious task; some stations were installed without these components so that more sites could be installed, improving data coverage. Processing resulted in estimates of impedance and tipper data at 35 logarithmically spaced frequencies between 909 Hz and 0.0016Hz. Three types of induction coil magnetic sensors were used in this survey: Phoenix MTC-50, MTC-150, and MTC-80H coils. MTC-80H sensors are shorter, and therefore used to measure the vertical magnetic field. MTC-50 and MTC-150 coils were used to measure the horizontal magnetic field components, of which the MTC-150 coils are more sensitive to higher frequencies than MTC-50 coils, leading to small differences in the frequency content of the collected data. To achieve parity with all stations, 11 of the highest frequencies were removed.

In June 2022, the ground was snow covered and the electric field measurements were made using titanium sheet electrodes and custom high-impedance amplifiers manufactured by Phoenix Geophysics. Snow cover was around 20-30 cm at the start of the survey, but melted rapidly during the measurement period. In August 2023, no snow was present at the surface. During this time, electric field measurements used non-polarizable electrodes buried at a depth of 20-30 cm in the ground. Overall, the data quality was significantly higher in 2023, which is largely attributed to the change in field conditions and equipment used for electric field measurements.

The local power plant and residential areas at Resolute are separated by a highway and connected by power lines that generate significant electromagnetic noise that is a challenge to MT data collection. Finding ideal locations for MT stations was challenging because of the extensive high voltage power infrastructure. Heavy vehicle travel on the main highway also may have caused significant vibrational noise. Moreover, wind can cause movement of the carefully oriented magnetic sensors, introducing noise in the data. The survey area was subject to particularly high winds, with maximum wind speeds of approximately 90 km/h in the final days of data collection. As was the case with instrument installation in Cambridge Bay, sensors were oriented using a handheld GPS device. Again, this was estimated to result in maximum uncertainty in orientation of between 1 - 2°.

Data were processed using both Phoenix EMpower and the statistically robust processing algorithm of Egbert (1997). Robust processing was found to provide MT sounding curves of higher quality, and data processed using this algorithm were used for further analysis.

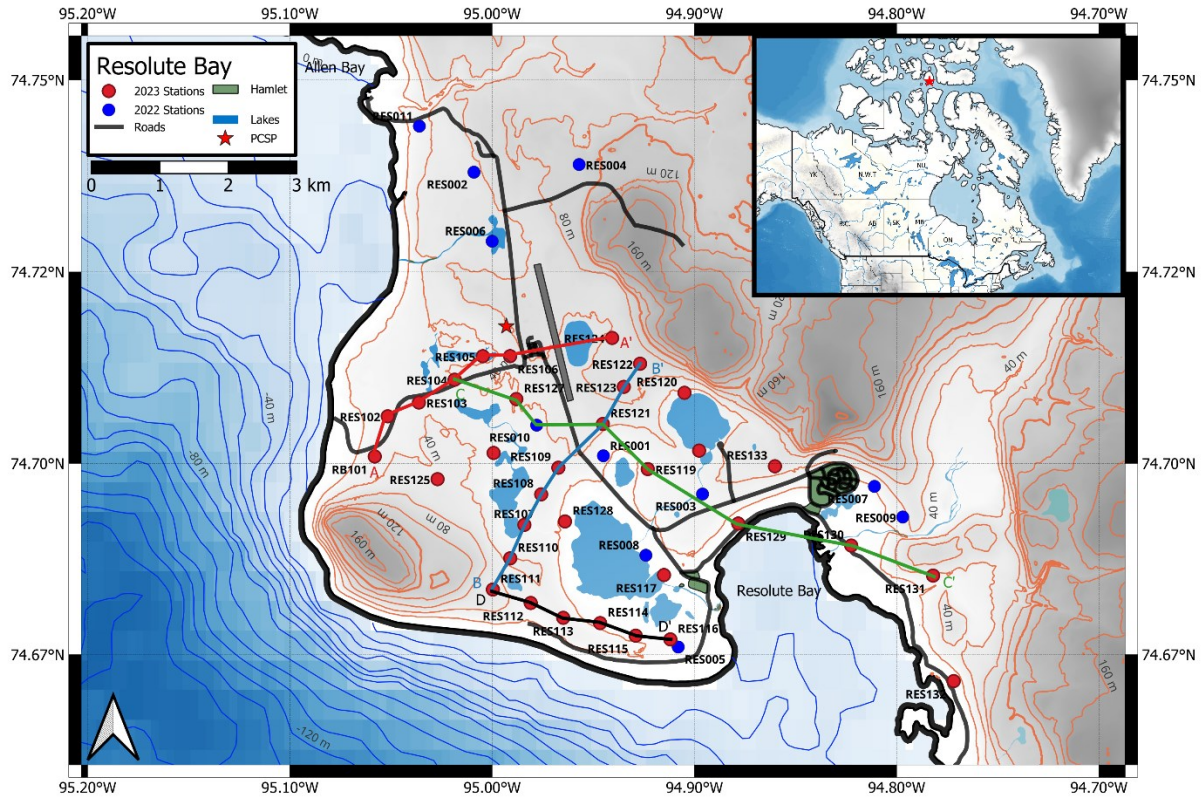


Figure 6.1: Overview map of the MT survey at Resolute Bay, Nunavut in 2022 and 2023. Red dots denote stations installed in 2023; blue dots denote stations installed in 2022. The red, blue, green, and black lines represent profiles used for analysis of the inversion models.

6.3. Resolute Bay electrical dimensionality

The same techniques used for the Cambridge Bay dataset were used for determining the dimensionality of the Resolute Bay electrical structure. Apparent resistivity and phase curves, induction vectors, and phase tensors were analyzed. These methods were discussed in more detail in Chapter 4. Low MT data quality, particularly in the 2022 stations and sites nearer to infrastructure, makes analysis of this dataset more challenging.

6.3.1. Resolute Bay apparent resistivity and phase

Apparent resistivity and phase are typically among the first aspects of MT data to be inspected, once the raw time series have been processed. It is again important to ensure that resistivity and phase are consistent at each MT station. Examples of processed data at stations of varying quality are shown in Figure 6.2.

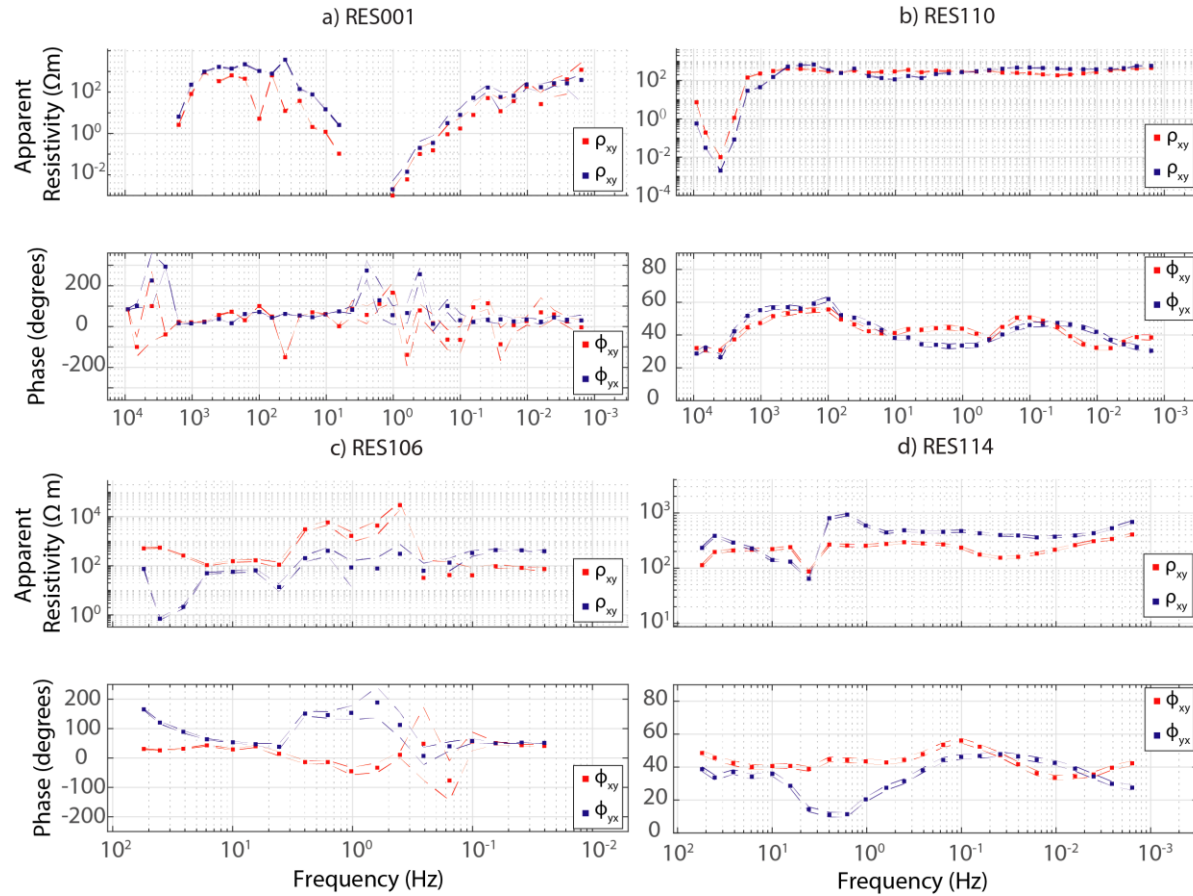


Figure 6.2: Examples of apparent resistivity and phase curves: a) RES001, collected in 2022 near the main highway; b) RES110, collected in 2023 away from infrastructure; c) RES106, collected in 2023 near PCSP; d) RES114, collected in 2023 away from infrastructure, but less than ~ 1 km from the southern coastline.

Quality of data collected at Resolute Bay is highly variable across the survey area, however, some consistent behaviour is observed. In the frequency band of about 100 Hz – 1 Hz numerous stations exhibit a drop in resistivity from values in the range 100 – 1000 Ωm to values less than 100 Ωm . At lower frequencies the apparent resistivity returns to values around 1000 Ωm at frequencies less

than ~ 1 Hz. This trend is observed at several stations, but is inconsistent in the level of apparent resistivity, and the frequencies at which these changes occur. Significant noise at high frequencies could be a major factor affecting the clarity of this feature in the processed data, as high apparent resistivity values can be observed at frequencies of 100 Hz or greater.

Another key observation in analyzing the apparent resistivity and phase curves of the processed MT data is the extreme differences observed between the xy and yx components. In most stations the components exhibit very dissimilar curves, or even curves of that appear anti-correlated. This behaviour could be indicative of a complex subsurface resistivity structure, likely 3-D.

It can be seen from the example sounding curves that the MT data quality is highly variable both spatially, and between the 2022 and 2023 campaigns. The highest quality MT data were typically recorded away from obvious sources of cultural noise, and using the non-polarizable porous pot electrodes in the 2023 campaign. A combination of unfavorable conditions was likely the cause of the poor MT data quality observed, including cultural noise, wind, and instrumentation. Several stations were removed from inversion modelling due to poor quality, which will be discussed later in detail.

As was done with the Cambridge Bay data, sounding curves were subject to additional post-processing to eliminate erroneous data points that produced rough models with poor consistency. This was again done using the D+ method (Parker, 1980; Parker & Whaler, 1981; Parker, 1994; Parker and Booker, 1996). Because of the significant noise levels present in the data, certain stations had a significant number of data points eliminated. This was particularly true at high frequencies where the effects were more pronounced.

Station	XY RMS	YX RMS	XY D+ RMS	YX D+ RMS	Station	XY RMS	YX RMS	XY D+ RMS	YX D+ RMS
RES001	3.66	4.22	0.083	0.33	RES112	1.81	1.88	0.54	1.02
RES002	1.9	2.19	1.02	1.35	RES113	1.34	1.94	0.33	0.96
RES003	3.01	3.14	1.24	0.76	RES114	0.81	1.6	0.12	0.42
RES004	5.68	4.91	3.62	4.88	RES115	1.74	0.65	1.18	0.51
RES005	1.76	1.85	0.92	0.59	RES116	0.89	1.49	0.43	0.45
RES006	6.12	5.98	2.47	4.1	RES117	6.59	6.54	0.53	0.68
RES007	1.71	3.45	0.41	1.39	RES118	1.12	3.51	0.44	1.33
RES008	2.01	1.8	0.68	1.12	RES119	0.8	3.47	0.37	0.37
RES009	0.52	1.8	0.36	0.67	RES120	1.17	4.48	0.73	0.96
RES010	2.06	2.06	0.9	0.94	RES121	2.12	2.34	1.47	1.41
RES011	0.43	0.33	0.42	0.21	RES122	1.42	2.56	0.54	0.35
RES101	0.31	0.1	0.24	0.1	RES123	0.53	2.17	0.23	1.21
RES102	0.7	0.19	0.47	0.19	RES124	0.15	1.84	0.14	1.01
RES103	1.36	0.38	1.05	0.26	RES125	0.6	0.25	0.57	0.25
RES104	1.6	0.7	1.7	0.62	RES126	1.2	0.55	0.98	0.38
RES105	3.98	2.44	0.22	0.08	RES127	1.29	0.31	1.03	0.25
RES106	2.71	4.45	0.61	1.05	RES128	0.72	0.56	0.7	0.55
RES107	1.71	1.59	0.59	0.81	RES129	3.56	3.45	0.42	0.7
RES108	0.76	0.87	0.76	0.87	RES130	4.74	1.77	1.1	0.63
RES109	1.56	3.23	0.83	1.98	RES131	1.67	1.06	0.58	1.05
RES110	0.38	0.57	0.38	0.57	RES132	6.83	0.95	6.83	0.95
RES111	2.39	1.7	0.92	0.97	RES133	6.02	1.49	0.23	0.63

Table 6.1: Station RMS misfit to D⁺ inversion models before and after D⁺ editing. Highlighted stations were removed after the first three inversions due to poor data quality.

6.3.2. Resolute Bay induction vectors

Induction vectors (IVs) provide an additional means of assessing the dimensionality of MT data. Over a 2-D resistivity structure, IVs are oriented perpendicular to the geoelectric strike, and point away from conductive features when plotted in the Weise convention. In the case of 3-D resistivity structure, IVs will also point away from subsurface conductors in the Weise convention. Induction vectors can also provide an effective method for identifying the lateral positions of sizeable subsurface conductors. Induction vectors at four frequencies are shown in Figure 6.3, adopting the Weise convention.

Induction vectors at high frequencies (40 Hz – 4 Hz) tend to point away from the nearest coastline. This is because the conductive seawater of the Barrow Strait consists of an enormous conductor leading to coast effects on nearby MT stations. A few stations that are located further inland do not point away from the nearest coastline, instead pointing to the west at 40 Hz, away from the main settlement of Resolute. This behaviour changes slightly at 4 Hz, with some of these induction vectors showing very small magnitudes. Unfortunately, vertical magnetic field data were only recorded at a small number of stations away from the coastline.

At lower frequencies (0.4 Hz – 0.04 Hz) nearly all induction vectors begin to align in directions away from the nearest coastline. As the skin depth of the signal increases, the region sensed by MT grows hemi-spherically away from the station and the data becomes more sensitive to the presence of seawater. At low frequencies, the vertical magnetic field data is dominated by the effects of the seawater.

Overall, the induction vectors derived from the Resolute Bay MT dataset are dominated by coast effects caused by the Arctic Ocean. There is insufficient vertical magnetic data away from the coastline to identify the location of non-ocean structure and data dimensionality.

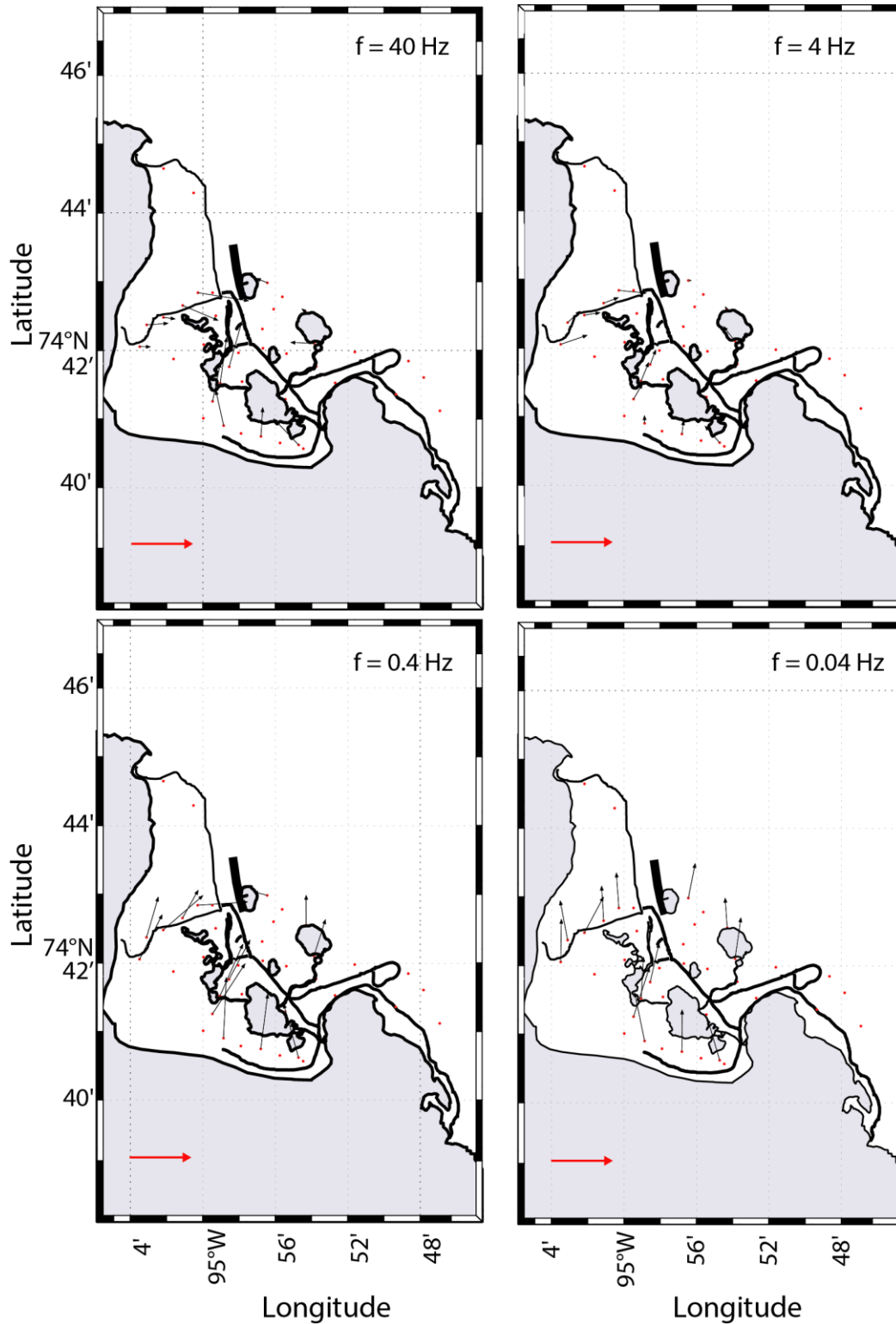


Figure 6.3: Induction vector maps are shown for the Resolute Bay dataset at four frequencies: 40 Hz, 4 Hz, 0.4 Hz, and 0.04 Hz. The Weise convention is used, where induction vectors point away from conductors. Unit vectors are shown in red.

6.3.3. Resolute Bay phase tensors

As was done with the Cambridge Bay dataset in Chapter 5, phase tensors were used to assess the dimensionality of the Resolute Bay MT dataset. The phase tensor method of Caldwell et al., (2004) provides information about the dimensionality of an MT dataset. This method is unaffected by galvanic effects, and does not require previous assumptions about the dimensionality. Figure 6.4 shows a phase tensor map of Resolute Bay at four frequencies. It is important to note that ellipses that are circular and have low skew angle represent 1-D resistivity structure; ellipses with low skew represent 2-D resistivity structure; ellipses with higher skew angle represent 3-D resistivity structure.

At a relatively high frequency of 40 Hz the MT data requires a 2-D or 3-D resistivity structure. Nearly all phase tensors show an elliptical shape with variable skew values occurring within the survey area. High values of skew angle that could be attributed to 3-D resistivity structure are found in the northern region of the survey area. It is important to note that noise in the data can significantly affect phase tensors, and as these data contained variable levels of noise a certain degree of variability can be expected in these results. At intermediate frequencies (4 Hz – 0.4 Hz) the MT data can be categorized as 3-D. This is evidenced in the ellipticity of most phase tensors, as well as the high skew values ($|\beta| > 2-3$) observed at many stations. At low frequencies (0.04 Hz) the phase tensors indicate 2-D or 3-D resistivity structure because of the observed elliptical phase tensors and high skew values at most stations. Notably, the skew angle at low frequencies is entirely positive, potentially indicating a change in the electrical resistivity structure at depth. Additionally, the phase tensors at many stations share an orientation that is parallel or perpendicular to the nearest coastline, likely indicating coast effects on the data becoming stronger at low frequencies.

It is readily apparent from the phase tensors that the electrical structure at Resolute Bay is relatively complex. At all frequencies the majority of the phase tensors are elliptical and have high skew angles ($|\beta| > 2-3$). This suggests a 3-D electrical resistivity structure should be modelled using 3-D inversion techniques. It is possible that these phase tensors are affected by the high level of noise in the data; however, the trends are fairly consistent across nearly the entire collection of stations at most frequencies.

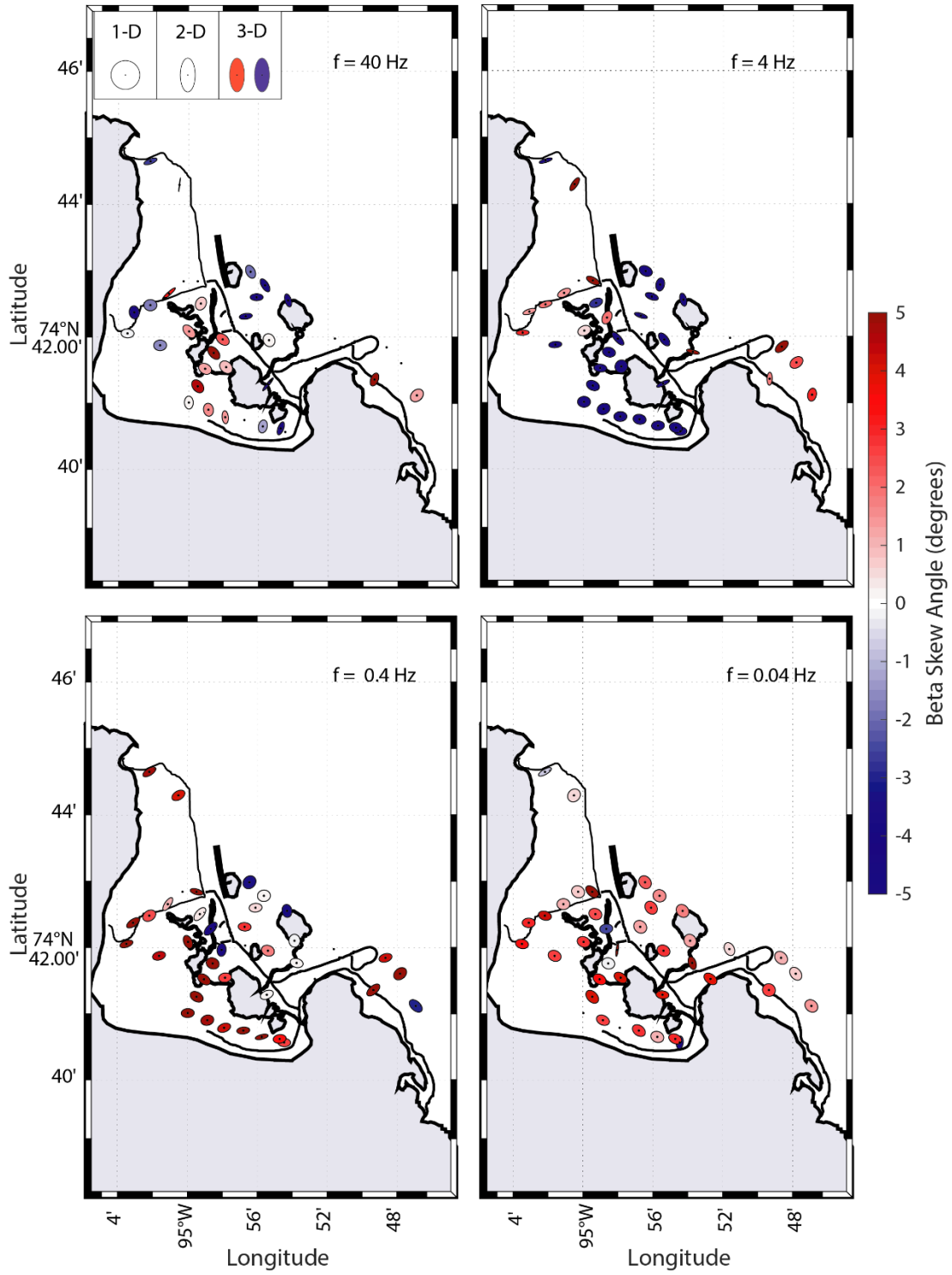


Figure 6.4: Phase tensor map at frequencies of 40 Hz, 4 Hz, 0.4 Hz, and 0.04 Hz. The ellipses show the azimuthal variation in the impedance phase and provide a graphical representation of whether the MT data can be considered 1-D, 2-D or 3-D.

6.4. Resolute Bay resistivity modelling

To interpret the MT data measured at Resolute Bay, they must be used to obtain resistivity models that are a function of depth, instead of frequency. Prior analysis of this dataset (Section 6.3) suggests that the dimensionality is largely 3-D. This MT dataset was not modelled using the manual 1-D trial-and-error method as was done with the Cambridge Bay data. This was because the apparent resistivity and phase data at Resolute Bay were more complex, and of lower quality than the Cambridge Bay data. The analysis of the Resolute Bay data used the Occam 1-D inversion algorithm (Constable, et al., 1987), and the 3-D inversion algorithm of (Kelbert, et al., 2014). The dimensionality analysis suggested that the data were consistent with a 3-D resistivity structure; however, applying 1-D inversion is a simple technique that could be compared to 3-D results to verify results.

Inversion models were displayed on four profiles, which were shown in Figure 6.1.

- Profile A-A' extends roughly 4 km southwest – northeast along the western road in the area. The profile extends across the power station and terminates on the eastern flank of the airstrip.
- Profile B-B' lies to the east and is parallel to Profile A-A'. This profile starts near the southern coastline of the island and extends ~ 4 km inland, crossing the main roadway.
- Profile C-C' is oriented approximately northwest – southeast, beginning at the midpoint of Profile A-A' and extending to the east side of the hamlet. The profile is approximately 8 km long, and crosses the seawater of the bay at ~ 4 km along its length. Profile D-D' is parallel to Profile C-C', begins at the same station as Profile B-B', extends ~ 2.5 km, and ends on the western coast of the bay.

6.4.1. 1-D Occam inversion of the Resolute Bay MT data

The Occam inversion approach of Constable, et al., (1987) was again used to produce 1-D resistivity models for this dataset. 1-D inversion provides a simple method for producing a preliminary model of the subsurface. While dimensionality analysis suggests that 3-D inversion is most appropriate for this dataset, 1-D models can be compared to inversion models produced via 3-D inversion, and can provide an indication of the MT data quality at each station. The Occam

inversion approach produces spatially smooth resistivity models that are less likely to show highly erratic resistivity structure than other algorithms that fit the data without imposing smoothness.

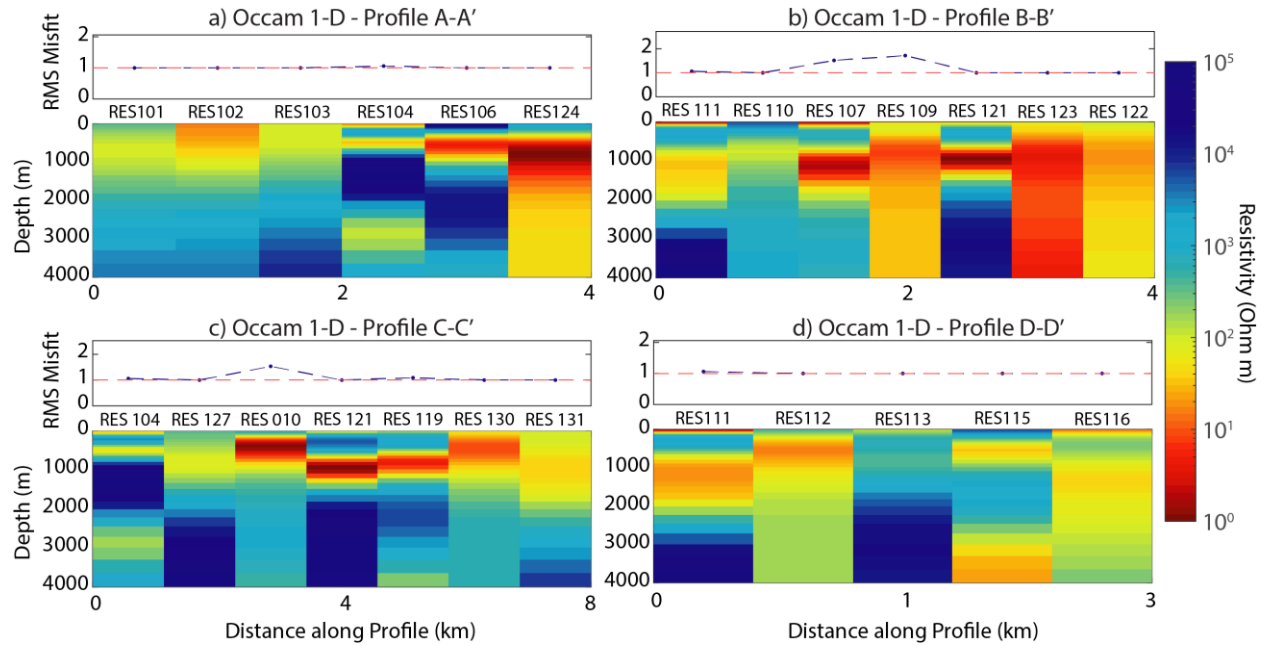


Figure 6.5: 1-D resistivity models obtained by inversion using the 1-D Occam inversion algorithm, shown along profiles for interpretation. a) Profile A-A'; b) Profile B-B'; c) Profile C-C'; d) Profile D-D'.

The one-dimensional inversion models produced by the Occam algorithm are quite variable across the survey area. In most cases an inversion model could be found that fits the MT data sounding with an RMS misfit of 2 or less. The resistivity values of different regions are variable, even when compared to neighboring stations, nevertheless, some patterns are clear. Most 1-D station models show evidence of a resistivity pattern consisting of a decreasing resistivity profile between depths of approximately 500 m – 1000 m. In many cases, there is evidence of a less resistive layer here, with resistivity values in the range 1-100 Ω m. Below a depth of 1000 m, most MT stations exhibit a rise in resistivity to values anywhere between 100 Ω m and 10,000 Ω m. The inconsistency in resistivity values between stations is likely caused by the high levels of noise in the data.

1-D inversion is likely not the best approach for the analysis of this dataset. The dimensionality analysis in section 6.3 showed that the resistivity structure is more complex and so the assumptions used in 1-D modelling are not necessarily valid. For instance, the Occam inversion 1-D resistivity models were inverted using the determinant (average) resistivity and phase data at each station. In

the case of a 2-D or 3-D resistivity structure, the xy and yx components will be quite different and so inverting the average likely does not produce accurate estimates of the true resistivity. However, 1-D inversion is a simple process that does show some key patterns in this dataset that can be explored further using 3-D inversion.

6.4.2. 3-D inversion of the Resolute Bay MT data

The dimensionality analysis of the Resolute Bay dataset showed evidence of a 3-D resistivity structure in the survey region. For this reason, 3-D modelling became the focus of the MT analysis of this dataset, and the ModEM algorithm was again employed for inversion using a variety of input parameters. A range of model grid size, starting model, model covariance, inclusion of bathymetry and topography, and the data selection were investigated in a range of inversions. A total of 10 3-D inversion models were produced, with most sharing the same primary resistivity structure.

The model space used for 3-D MT inversion includes a model grid composed of two regions, a model core surrounded by padding cells. The station locations are within the core of the grid and are the smallest grid cells at the surface. Padding cells are large cells that increase in size away the centre of the grid and surround the model core. Padding cells are required in modelling low frequency signals that have large skin depth (penetration depth). Since a large grid extent is required, padding cells that increase geometrically are used to fulfill this requirement without drastically increasing the total number of model cells. An initial layer thickness was used that was small enough to model the highest frequency signals with the smallest skin depth. For certain models, topography and bathymetry data were included in the modelling. Topography and bathymetry data from the IBCAO database (Jakobsson, et al., 2022) were included in four inversions with reduced data content, where cells that correspond to seawater were given resistivity values of $0.2 \Omega\text{m}$. The grid spaces of the first few models were chosen such that stations were separated by one or more cells. Some subsequent models used slightly different cell sizes to observe the effect on the final inversion models.

The first three inversions included all processed MT data from 2022 and 2023, and the remaining inversions excluded data from low quality stations. Stations that were removed from the inversion modelling are indicated in Table 6.1. The removal of poor quality MT data from the inversion had little effect on the overall structure of the 3-D resistivity models, indicating that the remaining MT

stations provided more coherent information about the subsurface than stations that were removed. All inversions used data at 24 frequencies from 55 – 0.0016 Hz. A 10 % error floor was applied to the impedance to ensure that certain frequencies were not given small weights due to very high estimates of the error.

The ModEM 3-D inversion algorithm (Kelbert et al., 2014) works by iteratively adjusting the model and then calculating the RMS misfit between the observed data and the data predicted by the new model at each step. The choice of an initial starting model, which is a parameter of the inversion algorithm, can play an important role in the final inversion model. If prior geological information is known, a starting model that incorporates this information can be used to influence the inversion process. Three types of starting models were used for inversion of the Resolute Bay data. The first was a simple 100 Ωm half-space which included no prior information about the geological setting. The second was a 100 Ωm half-space, but with the inclusion of low resistivity seawater based on the bathymetry data discussed above. The final starting model tested used a 1000 Ωm layer at the surface that was 500 m thick, followed by a 1000 m thick layer with a resistivity of 100 Ωm , and a half-space with a resistivity of 1000 Ωm below. The more complex starting model was used to encourage a final model with the general characteristics that were seen in the 1-D inversion modelling. In most cases the choice of starting model did not have a dramatic effect on the main features of the final inversion model, but minor differences were visible. The choice of a layered starting model produced distinct layering in the final model that was consistent with the starting model.

Of the ten inversion models produced, the preferred resistivity model was *RES_inv08*. *RES_inv08* used 66 model cells in the x -direction, 104 cells in the y -direction, and 75 cells in the z -direction. Each interior model cell was 100 m wide in both the x and y -directions, and eight padding cells were used along each boundary of the model with a padding factor of 1.3, corresponding to a total lateral model extent of 15 km x 11 km. In the z -direction, the initial model layer had a thickness of 1 m at the surface, which increased by a geometric factor of 1.15 at each layer beneath. The total model depth was 587 km. Model covariance values of 0.6 were used for the x and y -directions, and 0.3 for the z -direction. The covariance parameter controls the importance of smoothness between model cells. Since the permafrost layer is generally expected to be laterally continuous at Resolute Bay, horizontal smoothing was encouraged in the inversion. An inversion regularization

parameter of $\lambda = 1$ was used for inversion *RES_inv08*. Topography and bathymetry information was also included in this inversion. The inversion for *RES_inv08* had a starting RMS misfit of 3.52, which decreased to a final RMS misfit value of 1.48 after 72 iterations. This inversion used a 100 Ωm half-space as the starting model, incorporating 0.3 Ωm seawater at the surface based on bathymetric data. Inversion model *RES_inv08* was analyzed along both horizontal depth slices in Figure 6.6, and along the four profiles shown in Figure 6.7. Figure 6.8 and 6.9 show the RMS misfit between the observed data and the data predicted by *RES_inv08* by station and along a west-east transect, respectively.

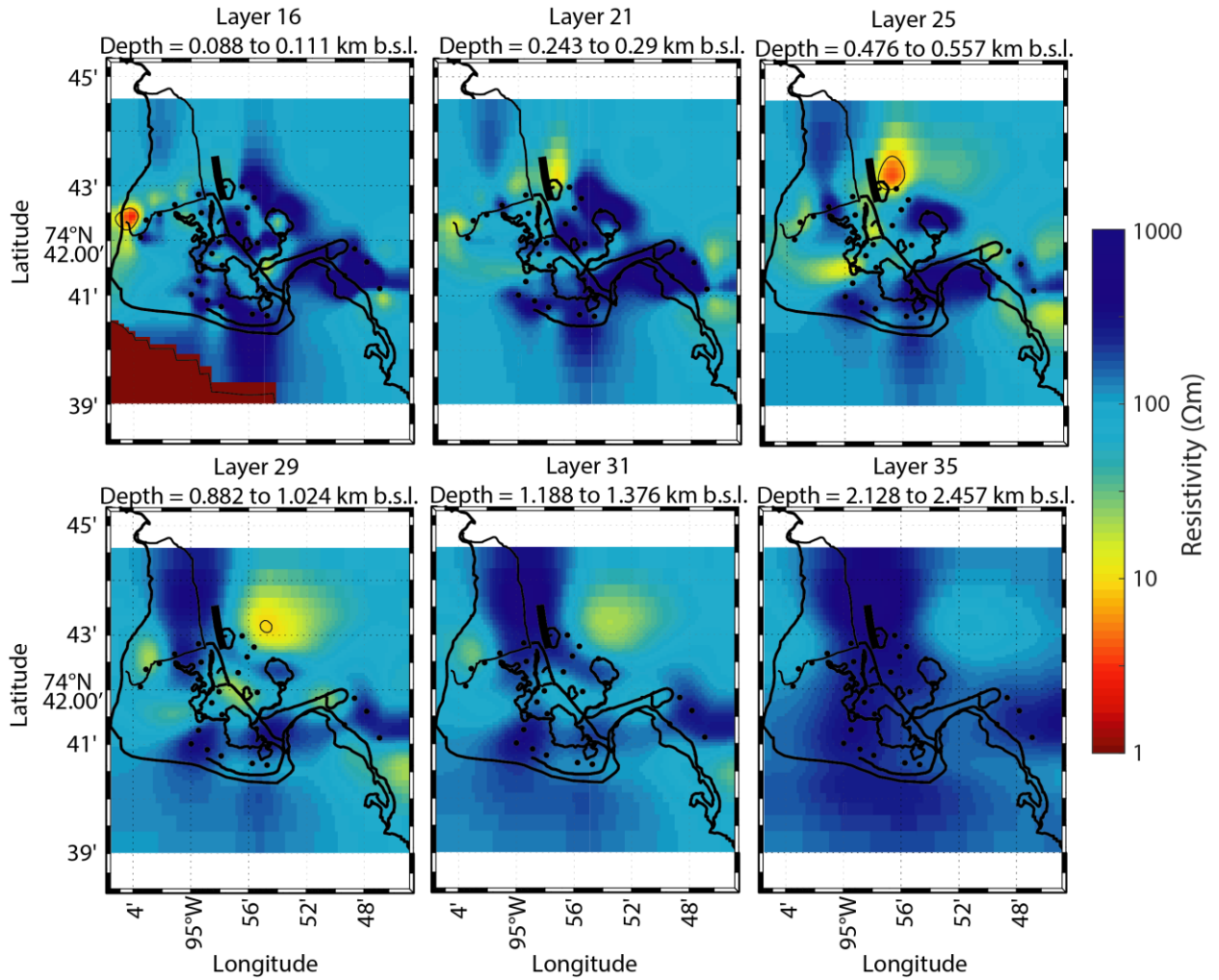


Figure 6.6: Horizontal sections of the 3-D resistivity model *RES_inv013* between 88 m and 2500 m. Model layer thickness increases geometrically with depth by a factor of 1.15 per layer. 0.3 Ωm seawater is included in the model based on Arctic bathymetric models.

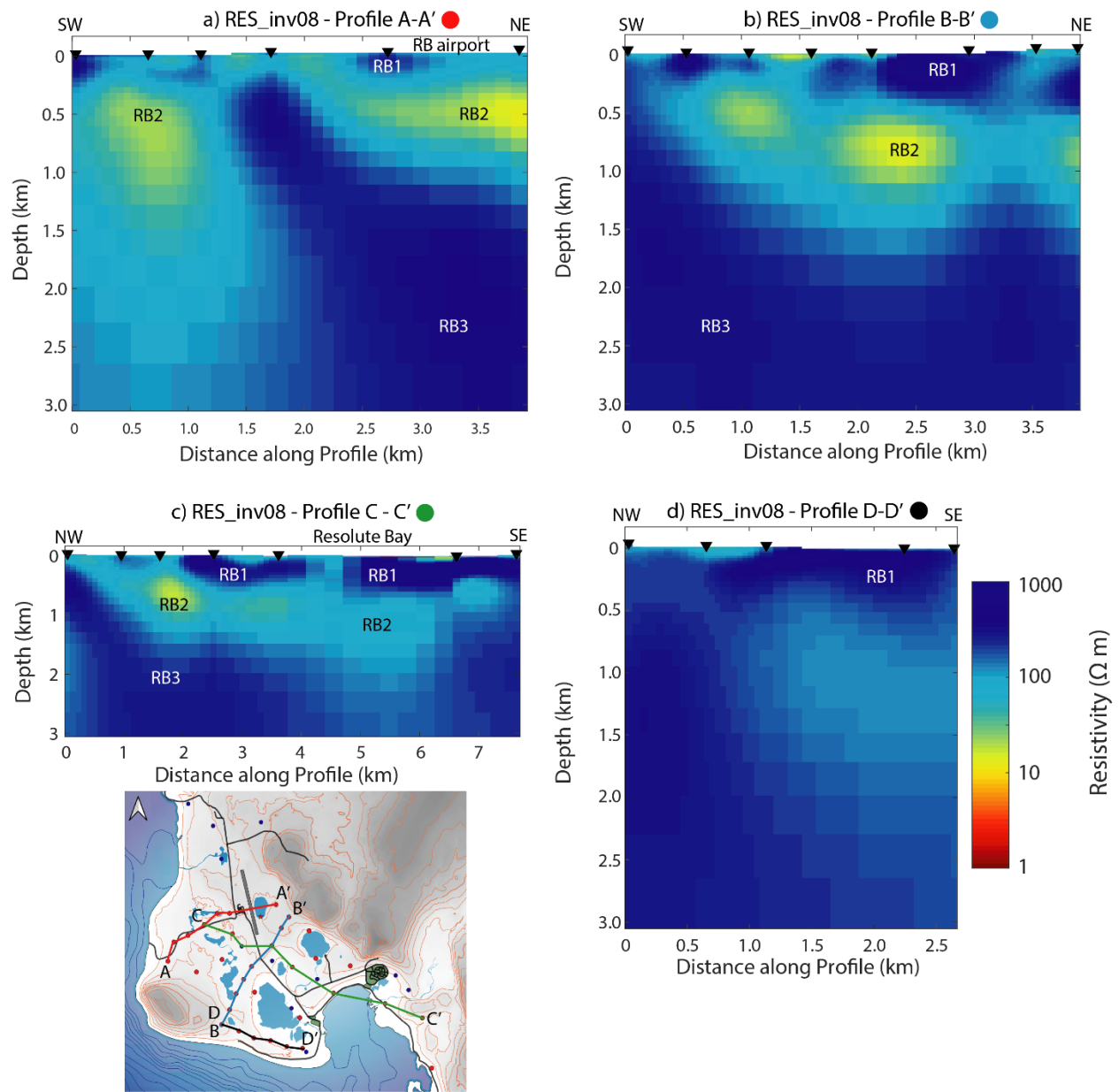


Figure 6.7: Vertical cross-sections of inversion model *RES_inv08* along profiles to a depth of 3 km: a) profile A-A'; b) profile B-B'; c) profile C-C'; d) profile D-D'. A reference map of the Resolute Bay survey is also shown.

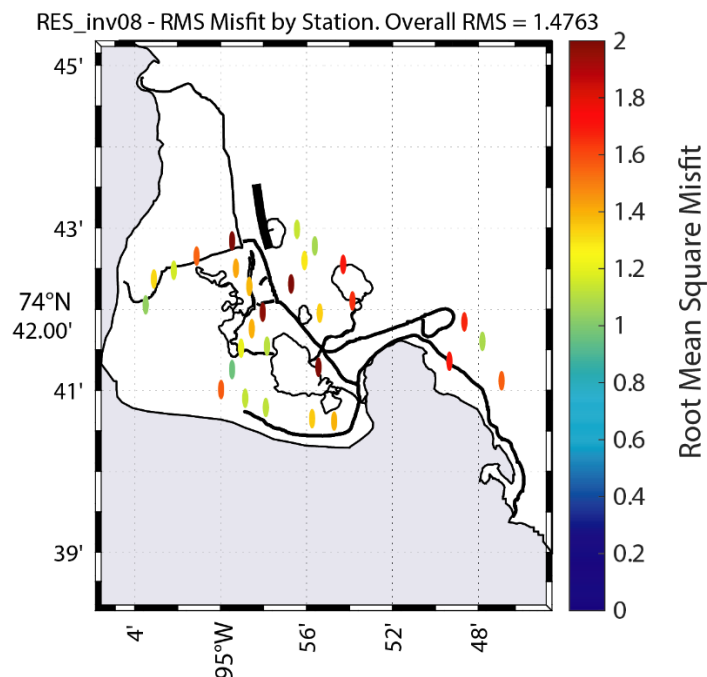


Figure 6.8: A map displaying the RMS misfit by station for inversion model RES_inv08.

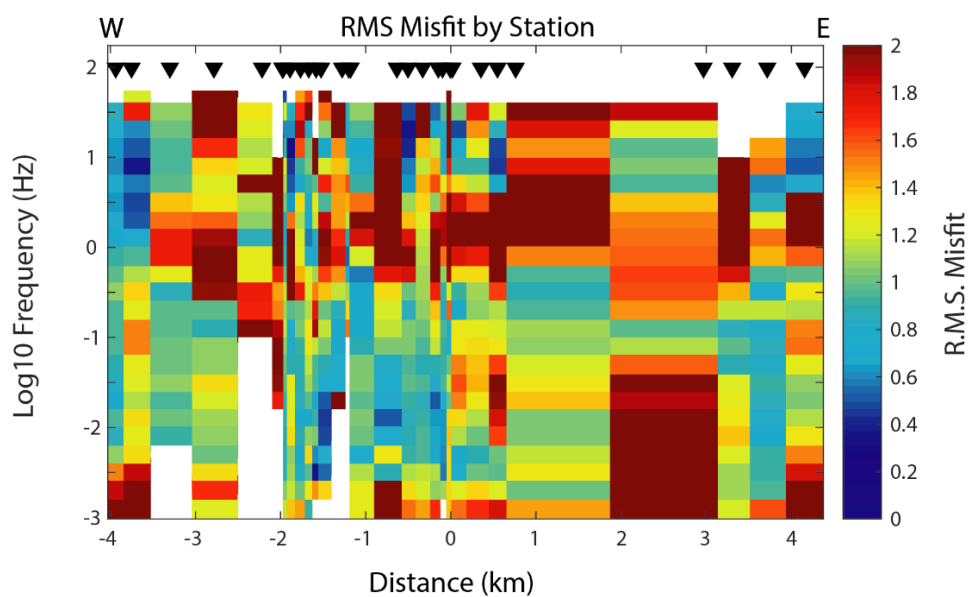


Figure 6.9: A pseudosection displaying the RMS misfit by station from W-E for inversion model RES_inv08. An RMS value of 1 indicates an ideal fit, with values greater than 1 indicating data that are being underfit by the model, and values less than 1 indicating data that are being fit too closely, given the estimated error. White blocks on the profile represent data that were removed during D+ editing of processed sounding curves.

3-D inversion models of the Resolute Bay dataset exhibit somewhat complex structures that are reasonably consistent along all four profiles. Descriptions of the preferred inversion model, *RES_inv08*, along the four profiles are as follows:

The near surface structure on profile A-A' is generally characterized by high resistivity in the range 100-1000 Ωm . A layer with lower resistivity is observed with its upper surface at a depth that varies along the profile. Towards the coastline, this layer varies significantly in depth and resistivity, but is generally shallower and thicker than sections of the model that are further inland. Beneath the airport, this low resistivity layer is in the depth range 500 - 1000 m and has a resistivity of 10-30 Ωm , and is generally similar to the model beneath profiles B-B' and C-C'.

Profile B-B' shows a similar resistivity structure to A-A' with a resistive surface layer (100 – 1000 Ωm) overlying a deeper conductive layer (10-30 Ωm). The conductive layer lies at depths between 250 – 750 m depth near the coastline, and deepening considerably to 500 – 1000 m inland.

Profile C-C' exhibits similar resistivity features as profiles A-A' and B-B', although the stratified features are less variable in depth and thickness than observed on the other profiles. The near surface layer has a resistivity of 100-1000 Ωm . The deeper conductive layer is located in the depth range 500-1000 m and has a resistivity of 10-30 Ωm . The conductive layer is poorly resolved along the profile where it crosses Resolute Bay. This is due to lack of MT stations in the bay.

The final profile, D-D', is distinct from the three other profiles, and is characterized by the absence of a pronounced low resistivity layer in the depth range 500-1000 m. This could be related to the fact that this profile is located on the edge of the grid of MT stations, and close to a major contrast in resistivity between Cornwallis Island and the seawater. In this inversion model, there is perhaps an indication of the expected lower resistivity layer, but this is not well resolved. This layer is generally less pronounced in other inversion models listed in Table 5.2.

It can be seen in Figures 6.8 and 6.9 that most stations that exhibit the highest RMS misfit with the predicted response are those situated near to the high voltage power line along the main highway or other infrastructure in the area. Aside from electromagnetic noise, vibrational noise from wind and the movement of vehicles along the highway could explain noise across a wider frequency band. Stations that were less affected by cultural noise were generally of better quality and were easier to fit with the 3-D inversion algorithm. Additionally, higher RMS misfits are more

typical at frequencies above ~ 0.3 Hz, possibly due to lower data quality, lower instrument sensitivity at high frequencies, and local noise.

Figure 6.10 shows the observed data and predicted data from model *RES_inv08* along a West-East transect across the survey area. Overall, the data predicted by *RES_inv08* is very similar to the observed data, however, there are visible differences, most notably at stations near to the settlement.

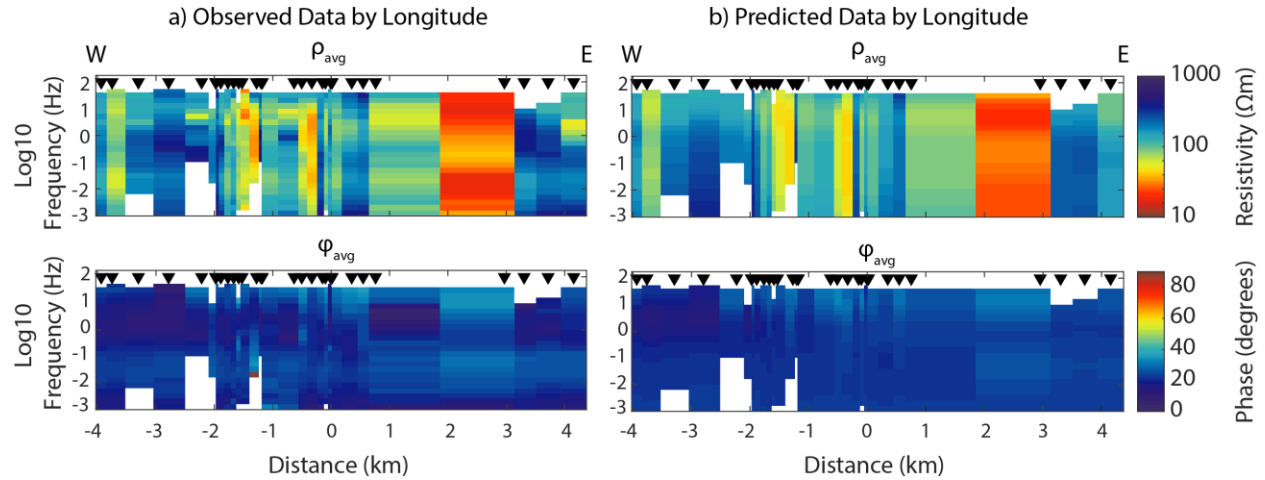


Figure 6.10: Pseudosections of apparent resistivity and phase projected along profile. a) observed data from W-E; b) the predicted response of inversion model *RES_inv08* from W-E.

6.5. Analysis of 3-D model sensitivity

The 3-D model produced by inversion was analyzed to determine the sensitivity of the main model features. Again, as for the analysis of Cambridge Bay, the model editing approach was used to analyze the sensitivity of model *RES_inv08*. Here, the resistivity model *RES_inv08* was edited to add a half-space to the model at various depths in the range 0.5 – 50 km. A summary of the sensitivity analysis is displayed in Figure 6.11.

The results of sensitivity analysis for *RES_inv08* show that the addition of a half-space affected the RMS misfit, even at depths as great as 50 km. Large changes to the RMS misfit at shallower depths show how the model sensitivity increases closer to the surface. The model was more sensitive to the addition of a 1000 Ωm half-space, compared to the 10 Ωm half-space. The RMS misfit began to start increasing rapidly when a 1000 Ωm half-space was added to the model between 20 – 30 km. The RMS misfit began to start increasing rapidly when a 10 Ωm half-space

was added to the model between 10 – 25 km. For shallower depths above 20 km, the RMS misfit increased significantly for both the 1000 Ωm and 10 Ωm half-spaces, with the most pronounced effects at depths below about 10 km. The key features of inversion model *RES_inv08* lie predominantly within the first 1-2 kilometers (Figure 6.7) suggesting that the MT data in the frequency band of 55 - 0.0016Hz were sensitive to depths greater than those identified in the previous section. Ultimately, the main features identified in *RES_inv08* appear to be well constrained by the MT data.

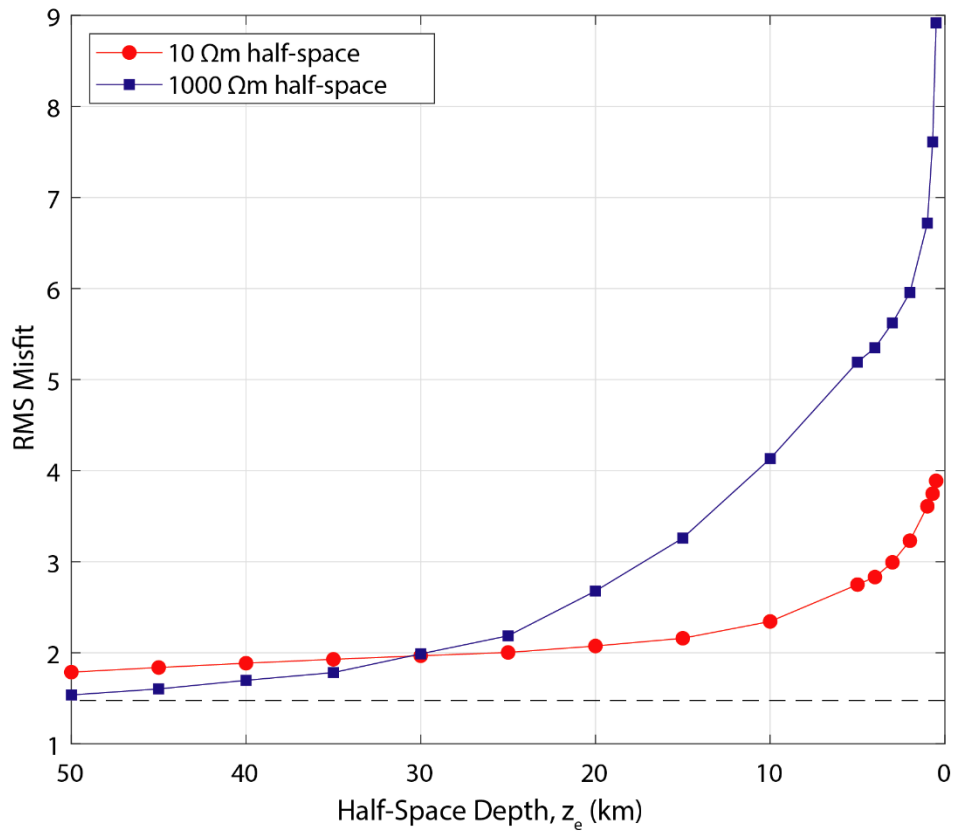


Figure 6.11: The RMS misfit when the resistivity of the inversion model, *RES_inv08* was edited to include a half-space below a depth, z_e . Analysis was done using a half-spaces with $\rho = 1000 \Omega\text{m}$ and $\rho = 10 \Omega\text{m}$. The dotted line shows the RMS misfit of *RES_inv08* without model editing.

6.6. Interpretation of MT inversion models

Analysis of the inversion models was undertaken to produce an interpretation of the subsurface resistivity structure at Resolute Bay. Resistivity models were examined to estimate the porosity

and temperature gradient in the subsurface to determine the geothermal potential of the region. Based on the resistivity models produced, a three-layered conceptual model was developed (Figure 6.12):

- RB1, a resistive surface layer with resistivity in range 100 – 1000 Ωm and approximately 500 m thick;
- RB2, a low-resistivity layer with resistivity in the range 10 - 30 Ωm in depth range 500 - 1000 m
- RB3, a resistive half-space with resistivity in the range 100 - 1000 Ωm at depths greater than 1000 m.

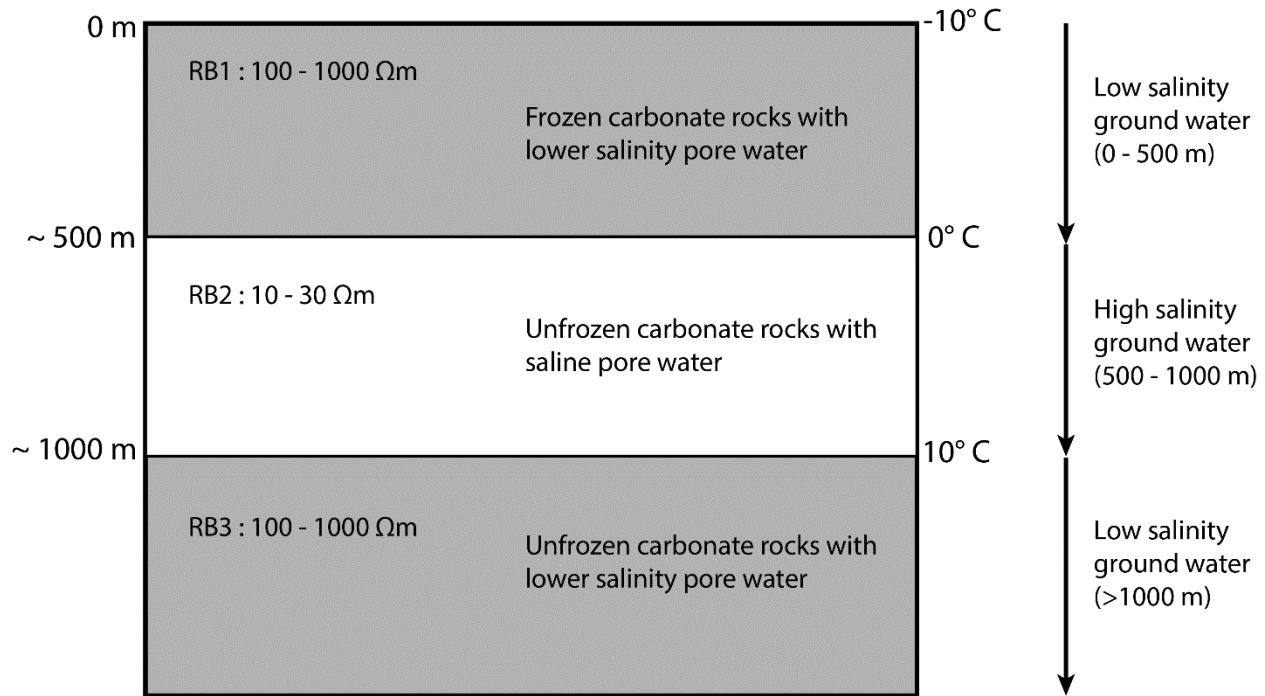


Figure 6.12: Summary of near surface resistivity structure at Resolute Bay, based on resistivity models derived from magnetotelluric data.

As was previously discussed in Chapter 3, the resistivity of rocks near the surface is dependent on multiple factors, including temperature, salinity, and whether they are frozen or unfrozen. Estimates on the thermal structure were used to assess the subsurface porosity and salinity. Minnick, et al., (2018) summarized the thermal structure at Resolute Bay. The MAGT at Resolute Bay is -10°C . A temperature gradient of around $20^{\circ}\text{C}/\text{km}$ is expected on the Arctic platform; however, previous measurements have suggested a higher local thermal gradient of

nearly 40°C/km (Misener, 1955). Both temperature gradients are considered in the analysis of the resistivity models produced by inversion of the 2022-2023 MT dataset to determine which gradient is more realistic (Figure 6.13).

If the subsurface temperature is assumed to be below the freezing point, the experimental results of Pandit and King (1979) can be used to interpret the resistivity of saturated frozen rocks. If the subsurface temperature is assumed to be above the freezing point, Archie's Law (Archie, 1942) can be used to assess the bulk resistivity of the rock from the porosity and pore water resistivity. Archie's law describes the behaviour of rock that is saturated with unfrozen fluid in the pore space, and the results of Pandit and King (1979) were measured from samples containing fluid that was frozen. In the case of a model with both frozen and unfrozen rock, neither approach can be used to completely describe the entire model.

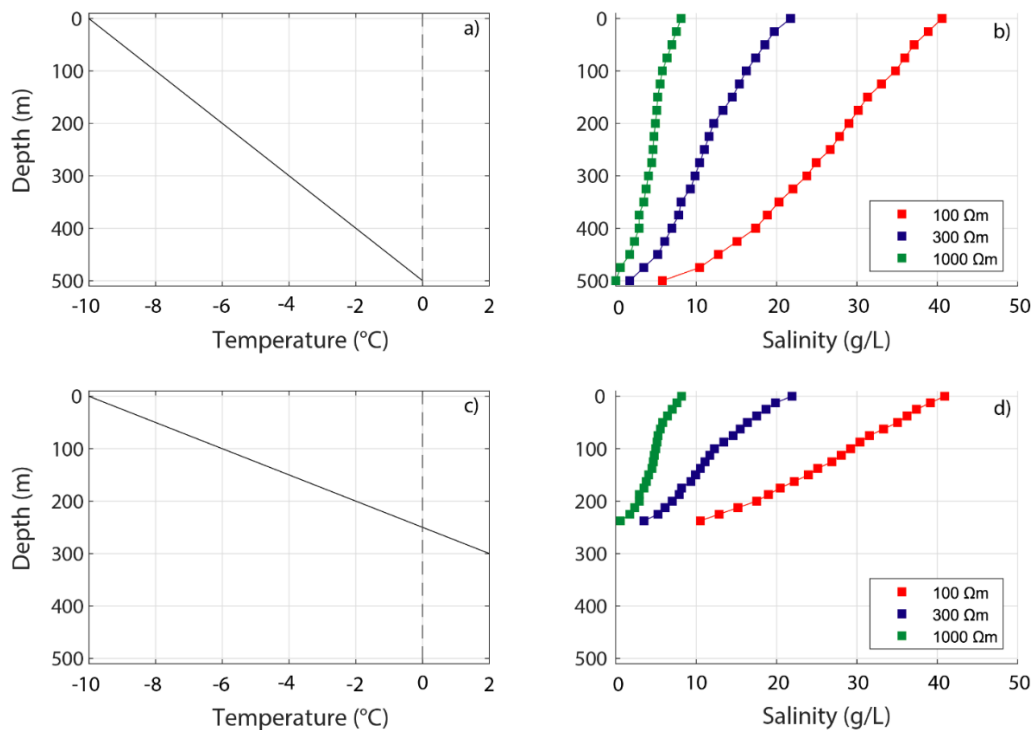


Figure 6.13: Interpretation of subsurface temperature, salinity and resistivity at Resolute Bay, Nunavut using the results of Pandit & King (1979). a) subsurface temperatures for a geothermal gradient of 20°C / km; b) salinity required to produce a specified resistivity with a geothermal gradient of 20°C/ km; c) subsurface temperatures for a geothermal gradient of 40°C / km as estimated by Misener (1955); d) salinity required to produce a specified resistivity with a geothermal gradient of 40°C/ km.schl

6.6.1. High-resistivity near-surface layer (RB1)

MT sounding curves and inversion modelling indicate that the surficial layer at Resolute Bay is relatively resistive ($\sim 100 - 1000 \Omega\text{m}$) and of varying thickness. Typical thickness of this layer is approximately 500 m, but varies significantly across the survey area. In particular, apparent thinning of RB1 along the profiles is of note, most strikingly along profiles A-A', B-B', and C-C'. In some regions close to the coastline, layer RB1 appears to thin to approximately 250 m or less.

The resistivity of RB1 is approximately $100 - 1000 \Omega\text{m}$, which is fairly consistent between inversion models. On average, the bulk resistivity of this layer is $300 \Omega\text{m}$ in the middle of the layer at a depth of 250 m. This value can be explained by a salinity of approximately 10 g/L, based on the experimental results of Pandit and King (1979). This salinity is less than the salinity of seawater ($\sim 30 \text{ g/L}$) suggesting that the fluids could have originated when the area was partially submerged in the Holocene and then been diluted. The geothermal gradient at Resolute Bay is poorly constrained. A typical value for a platform region in a continental interior is 20°C/km and this was used in the analysis described above. However, prior study has suggested that the geothermal gradient at Resolute Bay could be as high as 40°C/km . This higher geothermal gradient would predict a thinner layer with sub-zero temperatures as illustrated in Figure 6.13. The decrease in resistivity observed at a depth of 500 m is more consistent with a relatively low geothermal gradient (20°C/km) than a higher geothermal gradient (40°C/km). Overall, RB1 is likely to represent the region of frozen carbonates (i.e. permafrost) at the surface, with variation in thickness in the model either due to real insulating effects of the coast, or data sparsity and/or inconsistency.

6.6.2. Low-resistivity layer (RB2)

The intermediate layer, RB2, begins at a depth of around 500 m and has a bulk resistivity in the range of $10 - 30 \Omega\text{m}$. Even when a lower geothermal gradient is assumed the temperature within this layer is above the freezing point. The geology in this region is known to be a sequence of carbonate formations several kilometres thick; therefore, this layer is interpreted to be a region of unfrozen carbonate rocks.

Archie's Law was used to interpret the observed bulk resistivity using a value of $m = 2$ for the cementation exponent, reflecting a relatively isolated pore structure typically found in carbonate rocks (Schlumberger, 2013). Figure 6.14 shows how the bulk resistivity varies with porosity. A

range of pore fluid salinity values are considered in the range of 1 – 30 g/L. If a salinity of 10 g/L is assumed, based on the analysis of layer RB1, then this requires a porosity in the range 20 – 30%. This is reasonable for general studies of carbonates and could be verified if core samples from the study area become available.

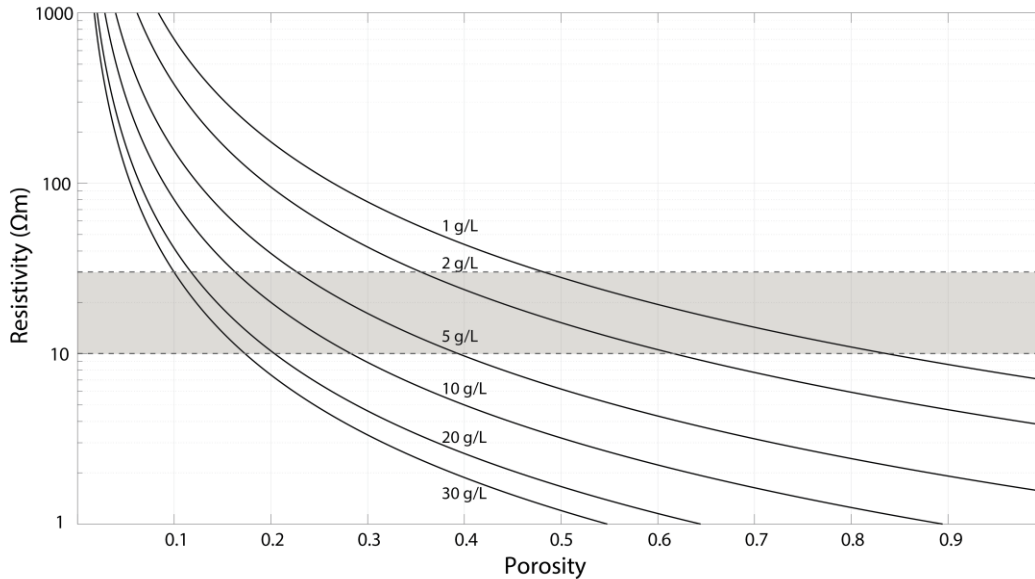


Figure 6.14: Resistivity of unfrozen limestone calculated with Archie's Law. Cementation exponent value was $m = 2$. Pore fluid resistivity was calculated using the equation of Block (2001). The grey box denotes the observed bulk resistivity in layer RB2.

6.6.2. Deep resistive layer (RB3)

Below a depth of 1000 m, a layer with a relatively high resistivity in the range 100 - 1000 Ωm is observed. A geothermal gradient of 20°C/km predicts a temperature of 10°C at this depth. A geothermal gradient of 40°C/km predicts a temperature of 30°C at this depth. Thus, the RB3 resistive layer will certainly not be frozen. At these depths, the geology is still considered to be characterized by carbonate rocks; however, at greater depths the porosity will begin to decrease due to increasing pressure, thus increasing the bulk resistivity of the rock. The resistivity of this layer can be explained by (a) a lower salinity than RB2 (b) a lower porosity than RB2 or (c) a combination of these two effects.

6.7. Discussion and implications for geothermal development

The magnetotelluric data collected at Resolute Bay have revealed a near surface structure that is significantly different to that at Cambridge Bay. At Resolute Bay, layer RB1 is identified as a frozen permafrost layer with a thickness that is consistent with the thickness expected for a 20°C/km temperature gradient. The lower surface of this layer occurs at a depth of 500 m at most locations along the profiles, which is the depth at which the temperature is expected to reach 0°C using a geothermal gradient of 20°C/km. This change in temperature can explain decrease in resistivity from RB1 to RB2. Apparent thinning of layer RB1 towards the coastline is visible along profiles A-A', B-B' and C-C', and can be explained by the insulating effects of the seawater (Figure 6.15) (Lachenbruch, 1957). A thinner RB1 layer could be explained by an elevated geothermal gradient of 40°C/km, however, the thinner regions are most obvious towards the coastline. In the absence of additional evidence of an elevated geothermal gradient, insulation of the frozen layer by the ocean provides a simpler explanation for the observed thinning. Overall, the resistivity models produced by MT modelling mirror that which is expected in a region of significant permafrost, that is, a resistive surface layer overlying a less resistive layer at depths where the permafrost thaws.

Quantitatively, the near surface layer (RB1) at Resolute Bay can be interpreted as a layer of frozen carbonate rocks with a salinity of 10 g/L. This salinity is significantly lower than in the surface layer at Cambridge Bay, and may reflect the differing sea level histories or saltwater drainage behaviour of the two locations over the Holocene (Simon et al., 2016). A layer of unfrozen carbonate rocks containing saline pore fluids is inferred to be present in the depth range 500 – 1000 m in the area located between the airport and the hamlet. This could represent a layer of rocks suitable for heat extraction and storage. The heat contained in this layer will be highly dependent on the geothermal gradient, and this quantity should be determined through temperature gradient drilling. Confirmation of the porosity and salinity is also required before any geothermal development can proceed. The decrease in resistivity from layer RB1 to layer RB2 at a depth of 500 m is consistent with a change from frozen to unfrozen subsurface conditions. This would correspond to a geothermal gradient 20°C/km at Resolute Bay. However, there are a number of uncertainties in this analysis, which must be considered.

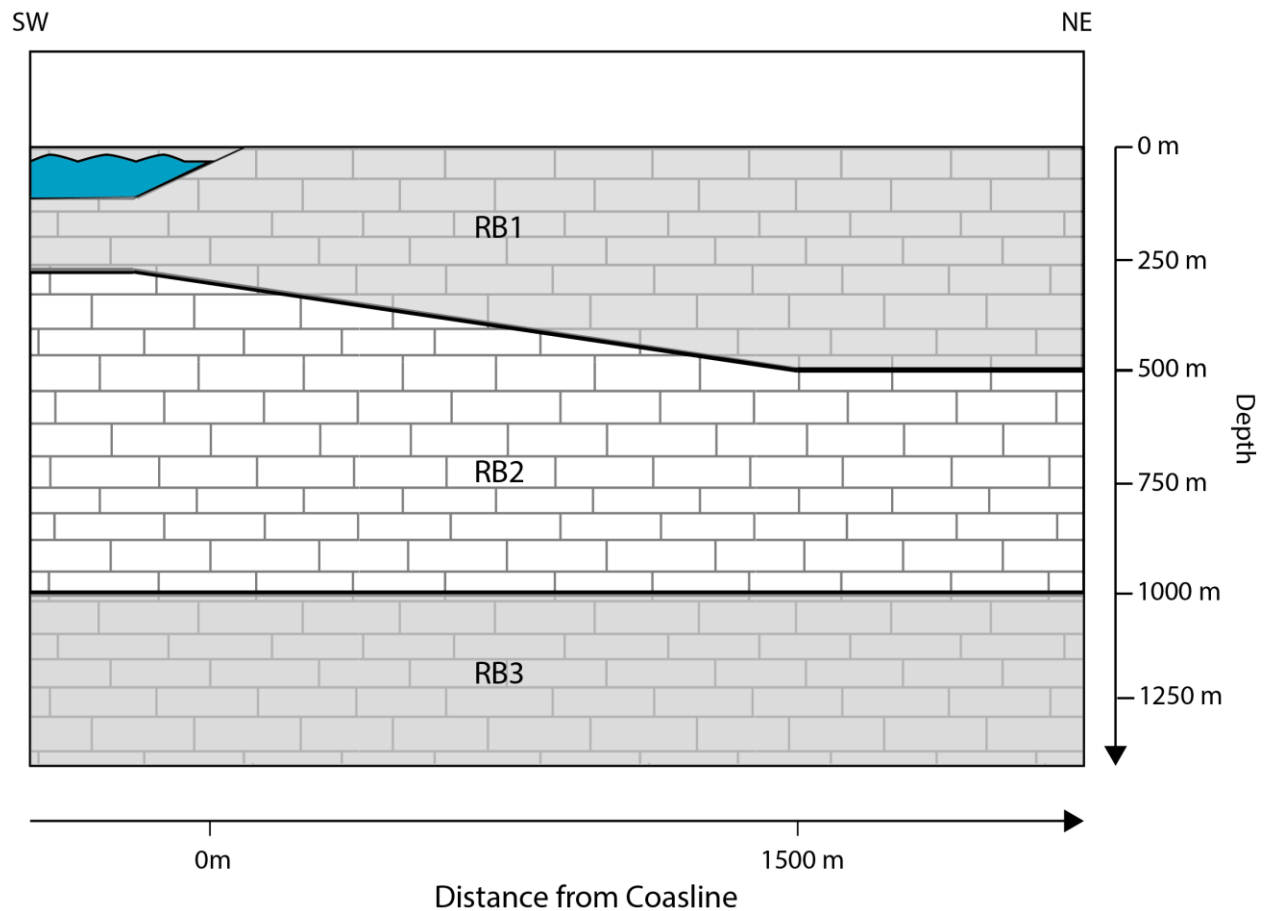


Figure 6.15: An illustration showing how the thickness of RB1 varies with distance from the coastline, as observed from subsurface resistivity models derived from MT data. RB1 attains maximum thickness of ~ 500 m approximately 1.5 km away from the coastline.

Lower grade geothermal resources, as inferred by the MT data, would suggest that direct use heating applications would be more appropriate at Resolute Bay. A $20^{\circ}\text{C}/\text{km}$ gradient would imply temperatures of approximately 10°C at 1 km depth, and 30°C at 2 km depth. The potential presence of elevated porosity at depths between 500 m and 1000 m could provide an ideal region for a geothermal reservoir, but the character of this region should be verified by drilling. The presence of a ~500 m permafrost layer would significantly increase the complexity and cost of drilling programs, and would need to be factored into cost estimates. Direct use geothermal applications could significantly reduce the reliance of the community on fossil fuels, and lower the overall carbon footprint. An example of such an application is the use of subsurface reservoirs to store solar-heated water. Industrial applications such as geothermal heating for greenhouses could also provide an additional usage for geothermal energy

Chapter 7: Numerical Modelling of Near Surface Salinity at Cambridge Bay

7.1. Introduction

Chapter 5 described how magnetotelluric data were collected to study the near-surface hydrogeology as part of an investigation of the geothermal energy potential at Cambridge Bay. A prominent feature of the resistivity model was a low resistivity layer extending from the surface to a depth of approximately 100 m, with resistivity in the range 30 - 100 Ωm . Below this layer, the resistivity increased rapidly with depth to values in excess of 1000 Ωm .

The low resistivity of this layer was inferred to be due to the presence of salt water that had diffused vertically downwards in the near surface sedimentary rocks during the period that Victoria Island was submerged following the melting of the Laurentide ice sheet over an ~ 10 kyr period during the Holocene (Ullman et al., 2016).

The interface observed at a depth of 100 m is too shallow to be associated with the base of the permafrost layer. Additionally, a transition from high resistivity to low resistivity is expected from MT data at the base of the permafrost layer, which is not seen in the Cambridge Bay MT data. Based on typical geothermal gradients, the permafrost thickness at Cambridge Bay is estimated to be approximately 500 m. Rather, the increase in resistivity is hypothesized to be the transition from saline to fresh pore water within the frozen layer. Numerical modelling is needed to determine if this hypothesis is valid: is it realistic that saltwater would have diffused 100 m in a time frame similar to transgression-regression periods? The numerical modelling used to evaluate this hypothesis is outlined in this chapter.

The process to be modelled is explained in detail in Section 7.3; however, a brief summary is as follows:

- The simulation considers a time period of 20,000 years (20 kyr) for a single glacial cycle. The initial time ($t = 0$ kyr) represents a point 20 kyr in the past, and the final time ($t = 20$ kyr) represents the current time.
- $t = 0 - 10$ kyr: In this period Victoria Island was covered by the Laurentide ice sheet. The rock surface temperature was fixed at -10°C , and the temperature at the base of the model

was fixed at 190°C at a depth of $z = 10$ km. This was calculated from the geothermal gradient that was estimated to be 20°C/km.

- $t = 10 - 19$ kyr: This was the period where the overlying ice had melted and the surface of Victoria Island was submerged in seawater. The surface temperature increased to 0°C and the surface was covered with seawater with salinity of 35 g/L.
- $t = 19 - 20$ kyr: This was the period where Victoria Island emerges from the sea due to post-glacial rebound. The surface temperature and salinity were returned to their original values of -10°C and 10g/L.

Inversion models obtained from MT data collected at Cambridge Bay were shown in Figures 5.7 and 5.8. An interpretation of these results was shown in Figure 5.14.

7.2. Finite difference scheme for temperature and salinity as a function of depth

To investigate the role of salt-water diffusion at Cambridge Bay a simple 1-D algorithm was implemented. Here, the overlying assumptions and methods are discussed.

7.2.1. Assumptions of numerical modelling

In this modelling, only 1-dimensional diffusion was considered. Heat diffusion and salt diffusion were assumed to occur in the vertical z -direction, with the rock surface corresponding to $z = 0$ m.

The modelling algorithm needed to simulate two distinct processes. The first was the thermal change associated with changes in surface temperature, which leads to melting or freezing of the permafrost. The second was the diffusion of salt within the groundwater of the unfrozen material. These processes are linked: the movement of salt is impeded by frozen material, and salinity depresses the freezing point of pore water and alters the thermal conductivity of the rock.

7.2.2. Finite difference scheme for temperature and salinity as a function of depth

A finite difference code was written to simulate spatio-temporal changes in temperature and salinity. This simulation used a finite difference scheme to solve the 1-D diffusion equation for both temperature and salt-water concentration. Both the forward time-centered space (FTCS) and Crank-Nicolson (Crank & Nicolson, 1947) schemes were tested for this algorithm. The schemes produced similar solutions for most time and space discretizations that were tested. However, the Crank-Nicolson scheme was more attractive in this case as it has no stability conditions on the

model discretization. The Crank-Nicolson scheme was more numerically intensive, but this did not pose any substantial problems for the tested discretizations.

The 1-D diffusion equation is given by:

$$\frac{\partial u}{\partial t} = k \frac{\partial^2 u}{\partial z^2} \quad (7.1)$$

where u can represent either the temperature or salinity of the medium, and k is a positive diffusion coefficient.

The forward time-centered space (FTCS) scheme discretizes this equation as:

$$\frac{u_i^{j+1} - u_i^j}{\Delta t} = k \frac{u_{i+1}^j - 2u_i^j + u_{i-1}^j}{(\Delta z)^2} \quad (7.2)$$

or combining the parameters Δt , Δz , and k :

$$u_i^{j+1} = \alpha u_{i+1}^j + (1 - 2\alpha)u_i^j + \alpha u_{i-1}^j \quad (7.3)$$

where $\alpha = \frac{k\Delta t}{(\Delta z)^2}$. The final model space consists $i = 1, \dots, M$ spatial layers in the z -direction, and $j = 1, \dots, N$ steps in time.

It can be shown through von Neumann stability analysis that the above scheme is conditionally stable for $\Delta t \leq \frac{(\Delta z)^2}{2k}$. A major drawback of this scheme is that for sufficiently small spatial discretization, time steps must be reduced significantly, increasing computational requirements. Challenges were encountered with this method in modelling layer thicknesses less than 10 m, as the time-step must be lowered significantly to maintain stability.

Alternatively, the Crank-Nicolson numerical scheme can be implemented, which approximates the derivative in time similarly to the FTCS scheme, but applies an average of second centered differences for the derivatives with respect to space:

$$\frac{u_i^{j+1} - u_i^j}{\Delta t} = \frac{1}{2}k \left[\frac{u_{i+1}^{j+1} - 2u_i^{j+1} + u_{i-1}^{j+1}}{(\Delta z)^2} + \frac{u_{i+1}^j - 2u_i^j + u_{i-1}^j}{(\Delta z)^2} \right] \quad (7.4)$$

$$-\alpha u_{i+1}^{j+1} + (2 + 2\alpha)u_i^{j+1} - \alpha u_{i-1}^{j+1} = \alpha u_{i+1}^j + (2 - 2\alpha)u_i^j + \alpha u_{i-1}^j \quad (7.5)$$

This forms a linear system that can be solved implicitly for the non-boundary values of the next time step:

$$\begin{bmatrix} 2 + 2\alpha & -\alpha & \cdots & 0 & 0 \\ -\alpha & 2 + 2\alpha & -\alpha & 0 & \vdots \\ 0 & \ddots & \ddots & \ddots & 0 \\ \vdots & \ddots & -\alpha & 2 + 2\alpha & -\alpha \\ 0 & \cdots & 0 & -\alpha & 2 + 2\alpha \end{bmatrix} \begin{bmatrix} u_2^{j+1} \\ u_3^{j+1} \\ u_4^{j+1} \\ \vdots \\ u_{M-1}^{j+1} \end{bmatrix} = \begin{bmatrix} \alpha u_3^j + (2 - 2\alpha)u_2^j + \alpha u_1^j \\ \alpha u_4^j + (2 - 2\alpha)u_3^j + \alpha u_2^j \\ \vdots \\ \alpha u_M^j + (2 - 2\alpha)u_{M-1}^j + \alpha u_{M-2}^j \end{bmatrix} \quad (7.6)$$

or

$$\mathbf{A}\mathbf{U}^{j+1} = \mathbf{U}^j \quad (7.7)$$

The Crank-Nicolson scheme allows for the calculation of model parameters at the next time step in their entirety; however, the $\mathbf{U}^j$ matrix needs to be constructed at each step. It can be shown that the Crank-Nicolson scheme is unconditionally stable for any choice of Δt and Δz , which makes this an attractive scheme for this modelling. In the case where α is constant throughout the model, the matrix $\mathbf{A}$ must be computed only once.

The modelling consisted of solving for the temperature and salinity values from one time step to the next. Since the salt is dissolved in the pore water, it was assumed that the salt content was unable to diffuse to cells that are frozen. In each iteration, the new cell temperatures were computed first and the salinity in a given cell was unchanged if the temperature of the cell was below the freezing temperature. Currently, the effects of salinity on the thermal conductivity of the rocks were not addressed in this model.

Since the freezing temperature of water is affected by the salinity, the freezing point depression was calculated before determining if the salinity changed in a model cell. The adjusted freezing point is then compared to the temperature of the cell to determine if the salinity changes. The freezing point depression due to salinity is calculated with:

$$dT = k_f \cdot b \cdot i \quad (7.8)$$

where dT is the change in freezing point, k_f is the cryoscopic constant (1.86 for water), and i is the molality of the salt water.

7.2.3. Boundary conditions

An overview of the boundary conditions used is described below and shown in Figure 7.1:

1. For $t < 10$ kyr, the surface temperature was fixed at -10°C , and the model base temperature was fixed to the expected gradient temperature at the model base depth. In this case, 190°C at 10 km depth for a geothermal gradient of $20^{\circ}\text{C}/\text{km}$ (Figure 7.2, panel 1).
2. For $10 \text{ kyr} \leq t < 19 \text{ kyr}$ the surface temperature was fixed to 0°C , and the model base temperature fixed to the same temperature as initially set. This was chosen to model the increase in temperature as the surface heats and the Laurentian ice sheet melts, submerging the surface of Victoria Island in seawater. Any changes in temperature at the surface are unlikely to propagate down to the bottom of the model in the time frame of the simulation so the model base temperature is unchanged.

The surface salinity was fixed to a value of 35 g/L to model the salinity of seawater covering the surface. (Figure 7.2, panels 2-3).

3. For $t \geq 19$ kyr the surface temperature returned to a value of -10°C , and the model base temperature was again kept the same as the previous stages. The surface temperature was reverted to the original value to model the emergence of Victoria Island in relatively recent times where the surface is expected to have refrozen (Figure 7.2, panel 4).

The surface salinity was reverted to the original value of 10g/L to simulate the surface rising out of the sea (Figure 7.2, panel 4).

In all cases, the model base salinity was fixed to the starting model salinity, 10 g/L. The diffusion coefficient used for NaCl in water was $\sim 1.5 \times 10^{-9} \text{ m}^2/\text{s}$ (Vitagliano & Lyons, 1956). The thermal diffusion coefficient for carbonates used was $\sim 1.25 \times 10^{-6} \text{ m}^2/\text{s}$ (Hanley, et al., 1977; Shen, et al., 2018). As the diffusion for salt through rock occurs primarily through pore water, the diffusion coefficient for NaCl was multiplied by the porosity of the rock to obtain an effective diffusion coefficient.

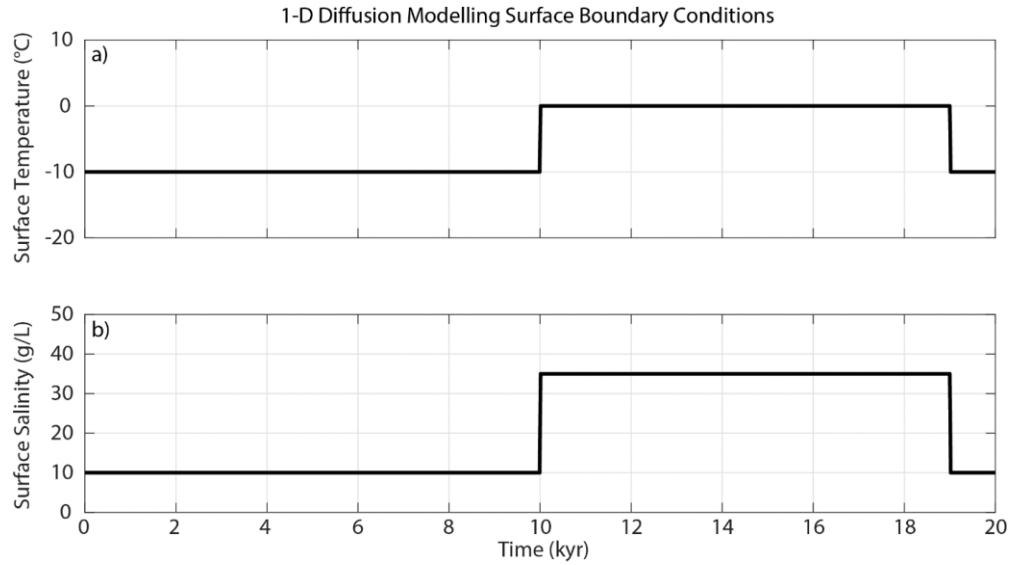


Figure 7.1: An overview of the surface boundary conditions applied to the diffusion modelling. a) The surface temperature was set to -10°C from $t = 0$ kyr to $t = 10$ kyr. The surface temperature rose to 0°C from $t = 10$ kyr to $t = 19$ kyr. The temperature returned to -10°C from $t = 19$ kyr to $t = 20$ kyr (present day). b) The model salinity began at 10 g/L from 0 kyr to 10 kyr. The salinity rose from 10 g/L to 35 g/L from 10 kyr to 19 kyr. The temperature returned to 10 g/L from 19 kyr to 20 kyr.

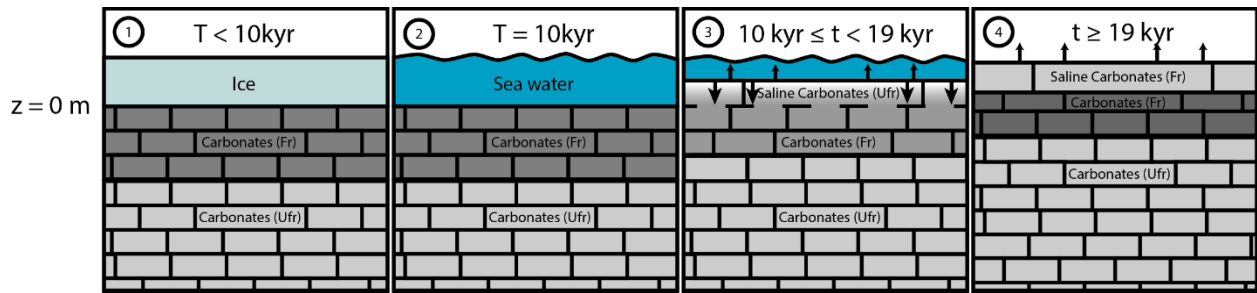


Figure 7.2: A diagram depicting the assumptions used in the diffusion modelling. (1) The carbonate platform is covered by the Laurentide ice sheet. (2) The ice sheet quickly melts, submerging the surface with seawater. (3) Elevated temperatures thaw the frozen rock. Over time, seawater diffuses into the carbonate rock from above. In the absence of the overlying ice sheet, the surface rises under the effects of post-glacial rebound. (4) The surface emerges from the ocean and re-freezes.

7.2.4. Estimating model resistivity

The subsurface hydrogeology can be imaged by measuring resistivity with MT. Thus, it was important to convert temperature and salinity profiles to resistivity to compare the estimated resistivity profile of the model with observations in the MT data. This was done using two methods:

1. The results of Pandit and King (1979) were used to estimate the resistivity of frozen sedimentary rocks. This was implemented in cells that were below the estimated depressed freezing points, which were calculated using the model salinity values. A model of the resistivity as a function of temperature and salinity was produced by linearly interpolating the results of Pandit and King (1979). This model was used to assign a resistivity value to any model cell that had a temperature below the freezing point.

This model was based on studies of a sample of Salem limestone by Pandit and King (1979), which had a porosity of 13%. This porosity value was also assumed in this modelling, where the diffusion coefficient for salt in water was multiplied by the porosity to obtain an effective diffusion coefficient. The porosity was also an important parameter in estimating resistivity of unfrozen rock using Archie's Law. Rock porosity is also likely to play a significant role in the resistivity of frozen rock; however, as the Pandit and King (1979) experiment only describes the behavior of a single limestone sample the extent of this is poorly understood.

2. Archie's Law (Eq. 3.5) was used to calculate the resistivity of model cells that were above the depressed freezing point. Here, the rocks were assumed to be completely saturated such that $S = 1$. Typically, $m = 2$ for carbonate rocks and reflects the type of porosity in this rock type.

7.3. Numerical experiments

Several numerical experiments were run to validate the simulation. In Model 1, a simple steady-state model, where the boundary conditions were kept constant for the entire duration, was used to verify that the algorithm was working as intended. Then, an experiment was performed to simulate a warming event at $t = 10$ kyr, but without a return to the original surface state. A third experiment

simulated a single glacial cycle over a 20 kyr period, with warming at $t = 10$ kyr and a return to the original surface temperature at $t = 19$ kyr. The final experiment simulated a series of successive glacial cycles, repeating the surface conditions of the previous model over a period of 200 kyr.

7.3.1. Model 1: constant temperature and salinity

The first test of the numerical model was used to confirm that under unchanging conditions, the starting model would also remain unchanged. This scenario applied no perturbation to the starting model and expected the model to remain unchanged throughout the entire 20 kyr duration. Temperature, salinity, and resistivity models are shown below in Figure 7.3. The starting model and final model for the Model 1 simulation are shown in Figure 7.4 for spatial step sizes of 10 m and 5 m. The percent difference between the starting model and final model is shown in Figure 7.5 for spatial step sizes of 10 m and 5 m.

In both cases of step size, $\Delta z = 10$ m and $\Delta z = 5$ m, the temperature and salinity models exhibited no change over the 20 kyr simulation when there were no changes to the boundary conditions of the model. This result validated the algorithm used for the modelling as this result was consistent with the behaviour expected for a steady-state diffusion model.

For the Model 1 scenario, the starting model and the final model were identical, showing no change throughout the simulation. These results confirmed that the simulation was stable for vertical spatial discretizations of $\Delta z = 10$ m and $\Delta z = 5$ m and suggest that these spacings are sufficient to accurately describe changes near the surface.

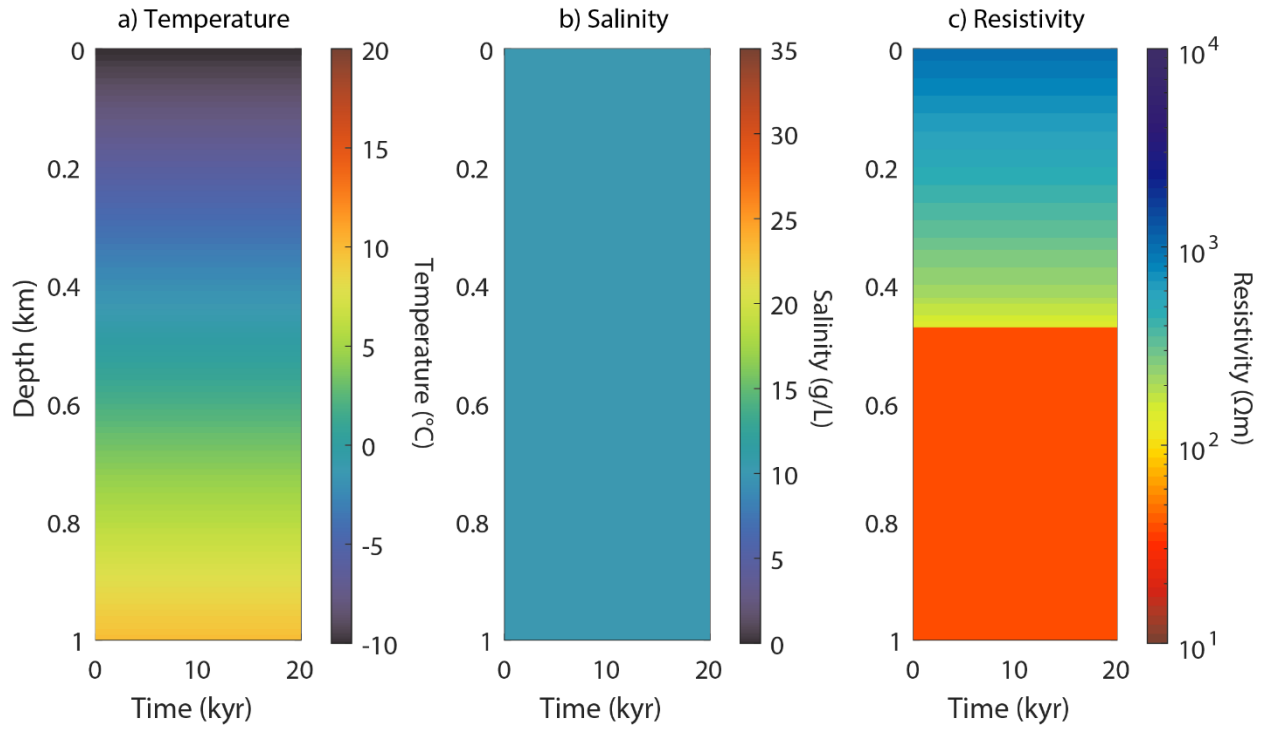


Figure 7.3: Results of the constant temperature and salinity (Model 1) simulation with $\Delta z = 10$ m: a) a plot of the temperature-depth profile at all time steps; b) plot of the salinity-depth profile at all time steps; c) a plot of the resistivity-depth profile predicted by the calculated temperature and salinity at all time steps.

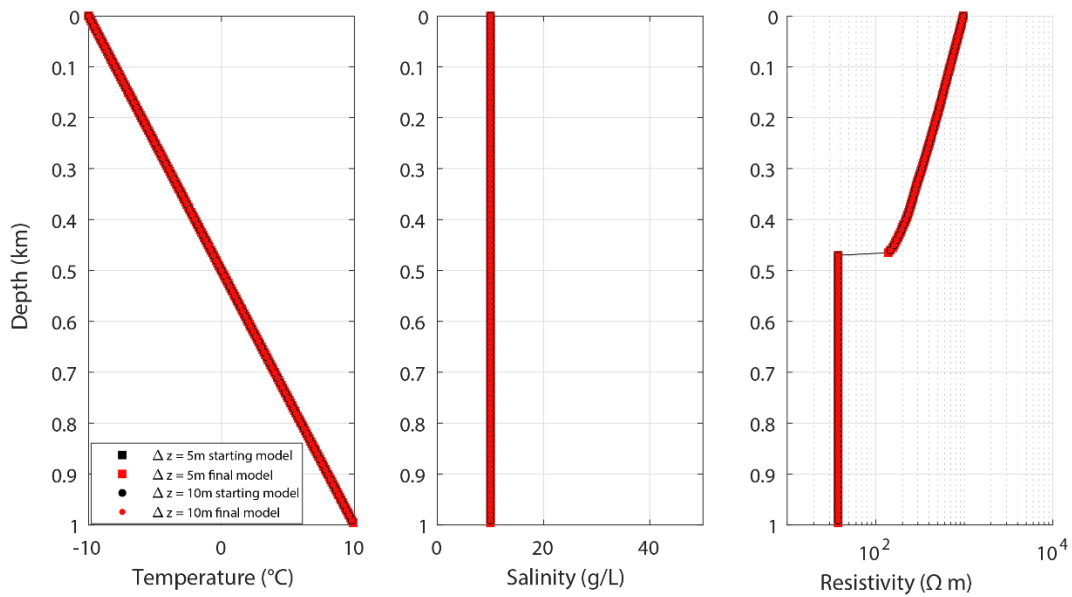


Figure 7.4: a) The starting and final temperature, salinity, and resistivity models for the constant temperature and salinity scenario (Model 1) for $\Delta z = 10$ m and $\Delta z = 5$ m.

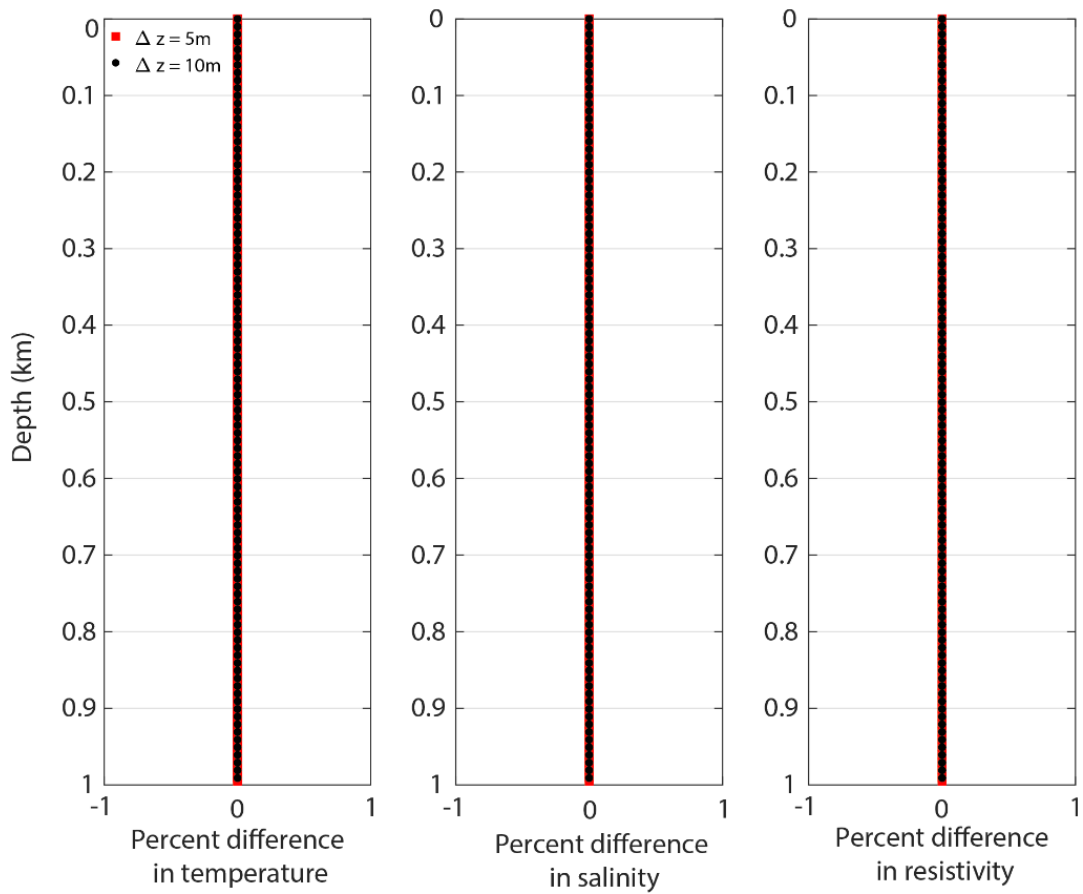


Figure 7.5: The percent difference between the starting and final models for the constant temperature and salinity scenario (Model 1) are shown for vertical grid spacings of $\Delta z = 10 \text{ m}$ and $\Delta z = 5 \text{ m}$.

7.3.2. Model 2: end of ice age with no temperature recovery

To further test the modelling algorithm, the starting model was subject to an increase in surface temperature to 0°C , and an increase to the surface salinity to 35g/L at $t = 10 \text{ kyr}$. This experiment subjects the model to conditions that simulate the end of the ice age, where the surface temperature increases and the land surface is submerged in seawater. In this experiment; however, the boundary conditions do not return to their original states and the simulation is run for a much longer time of 50 kyr . In this scenario, the model moves out of equilibrium where it will readjust to the new conditions. After some time the model is expected to return to a steady state.

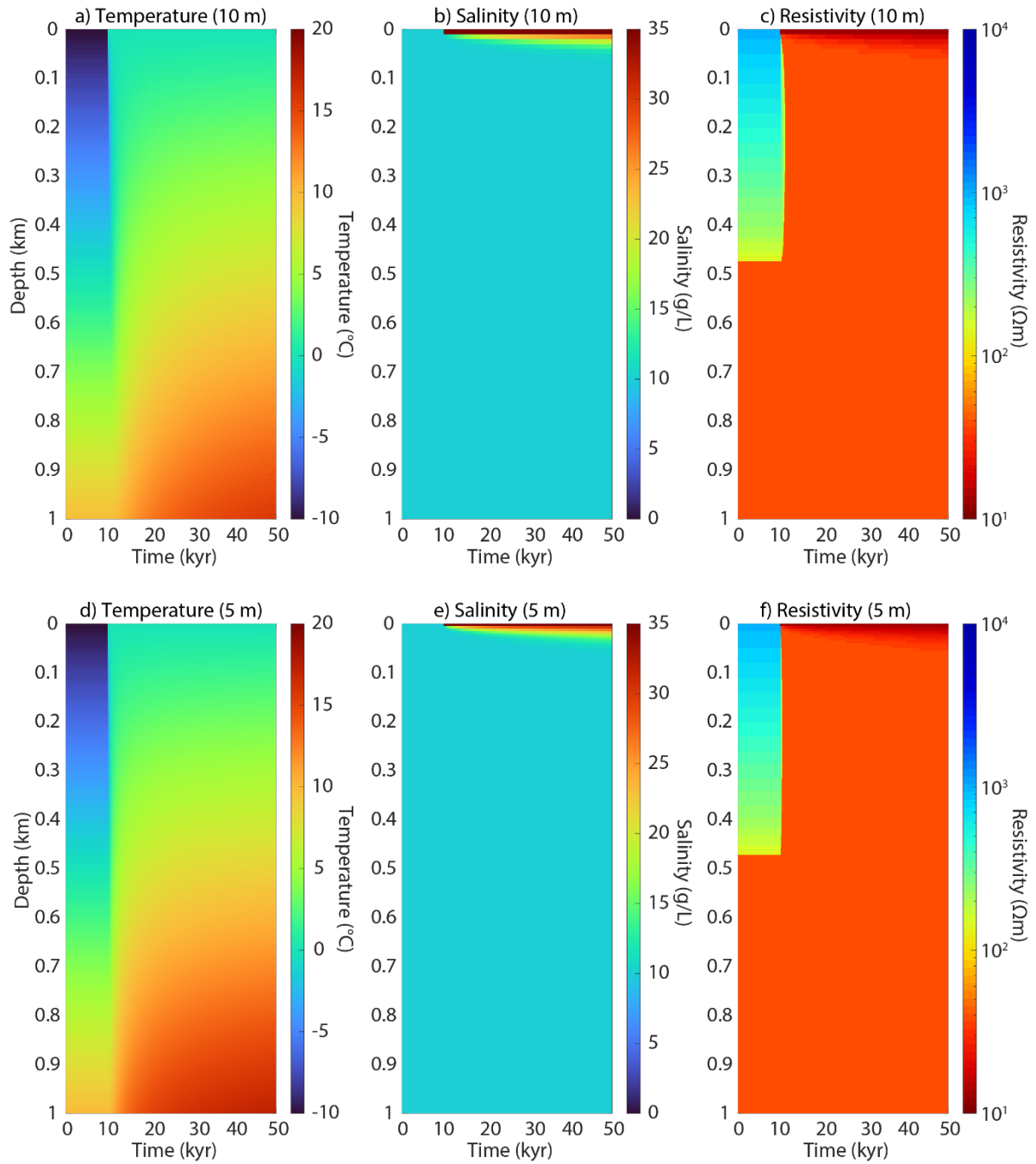


Figure 7.6: Simulation of the Model 2 scenario at all time steps down to depths of 1km. Results are shown for simulations using $\Delta z = 10$ m and $\Delta z = 5$. The Model 2 simulation used a total duration of 50 kyr.

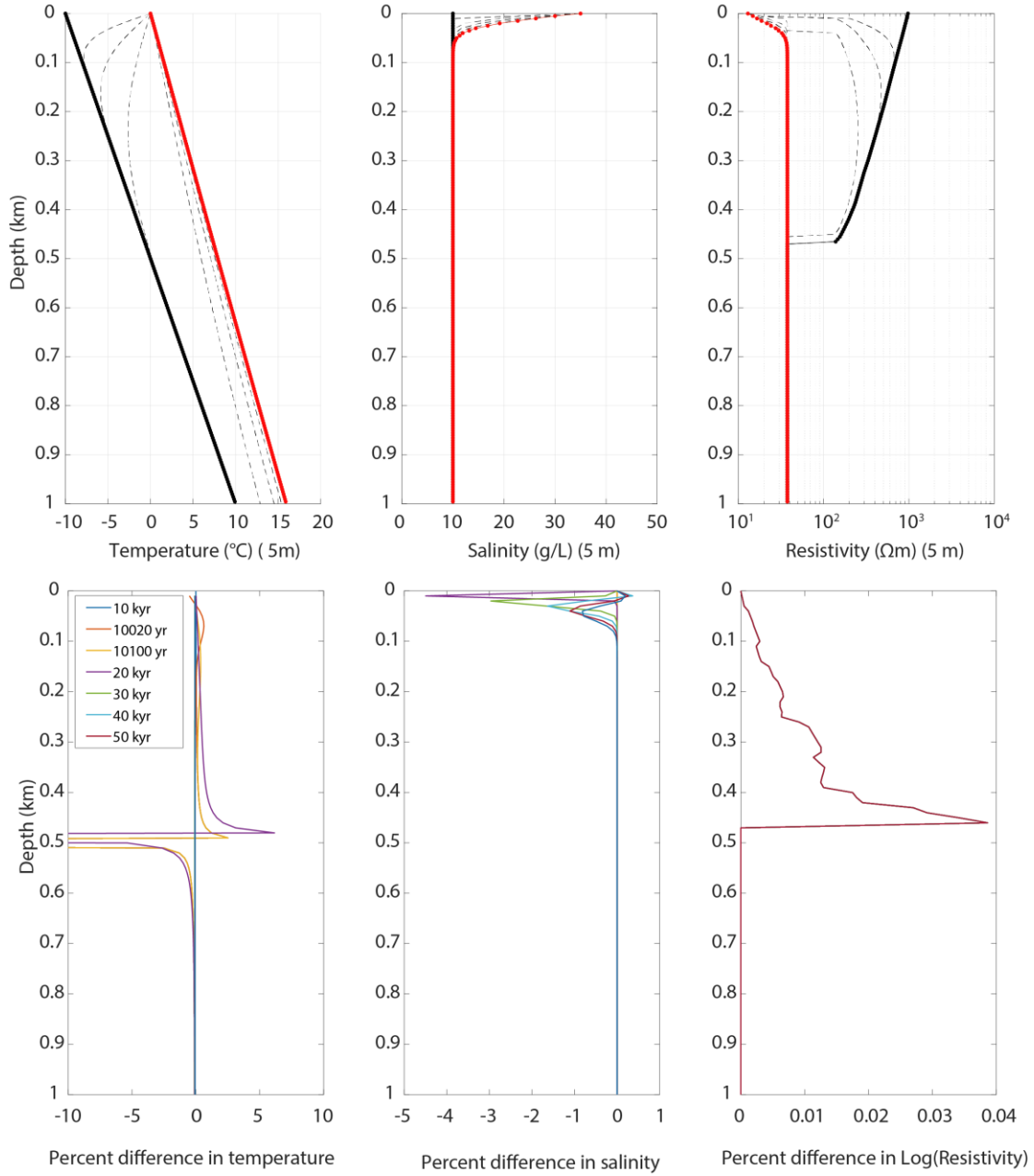


Figure 7.7: The starting and final models, as well as intermediate models, are shown for Model 2 for depths down to 1 km, using $\Delta z = 5$ m. The beginning and final models are represented by solid lines in black and red, respectively. Intermediate models are shown as dotted black lines for $t = 10.02$ kyr, $t = 10.1$ kyr, $t = 10.5$ kyr, $t = 20$ kyr, $t = 30$ kyr, $t = 40$ kyr. The percent differences between the models using $\Delta z = 5$ m and $\Delta z = 10$ m are also shown.

It is clear from the results of the experiment with no temperature decrease at $t = 19$ kyr that the effects of changes in the surface temperature propagate much faster than the changes in surface salinity. This was expected when considering the orders of magnitude difference in diffusion coefficients between the two. Computed differences between models using different step sizes showed that there was little difference between models that used $\Delta z = 10$ m and $\Delta z = 5$ m which supports the notion that the discretization was sufficient in the vertical direction to successfully model the diffusive effects. Differences between $\Delta z = 10$ m and $\Delta z = 5$ m were less than 5% between salinity models and were much smaller when comparing temperature and resistivity models.

As expected, the temperature profile showed convergence towards a new linear gradient following the sudden change in temperature at the surface. The salinity changes were much slower and were expected to converge towards the surface salinity at all model cells, but this would likely take a far longer time. Interpreted resistivity changes were largely dominated by the differences between the Archie's Law and Pandit and King (1979) methods. Since the temperature quickly rose above 0°C near the surface, the algorithm began to interpret the resistivity near the surface using the Archie's law method. The salinity began to lower the estimated resistivity near the surface as the salt diffused downwards.

7.3.3. Model 3: end of ice age with temperature recovery

The algorithm was tested with similar conditions as in Model 2, but with reversal of the surface boundary conditions to their original state at $t = 19$ kyr. This step simulated the rise of Victoria Island from the seawater where the surface has refrozen. Results of modelling using these conditions are shown below.

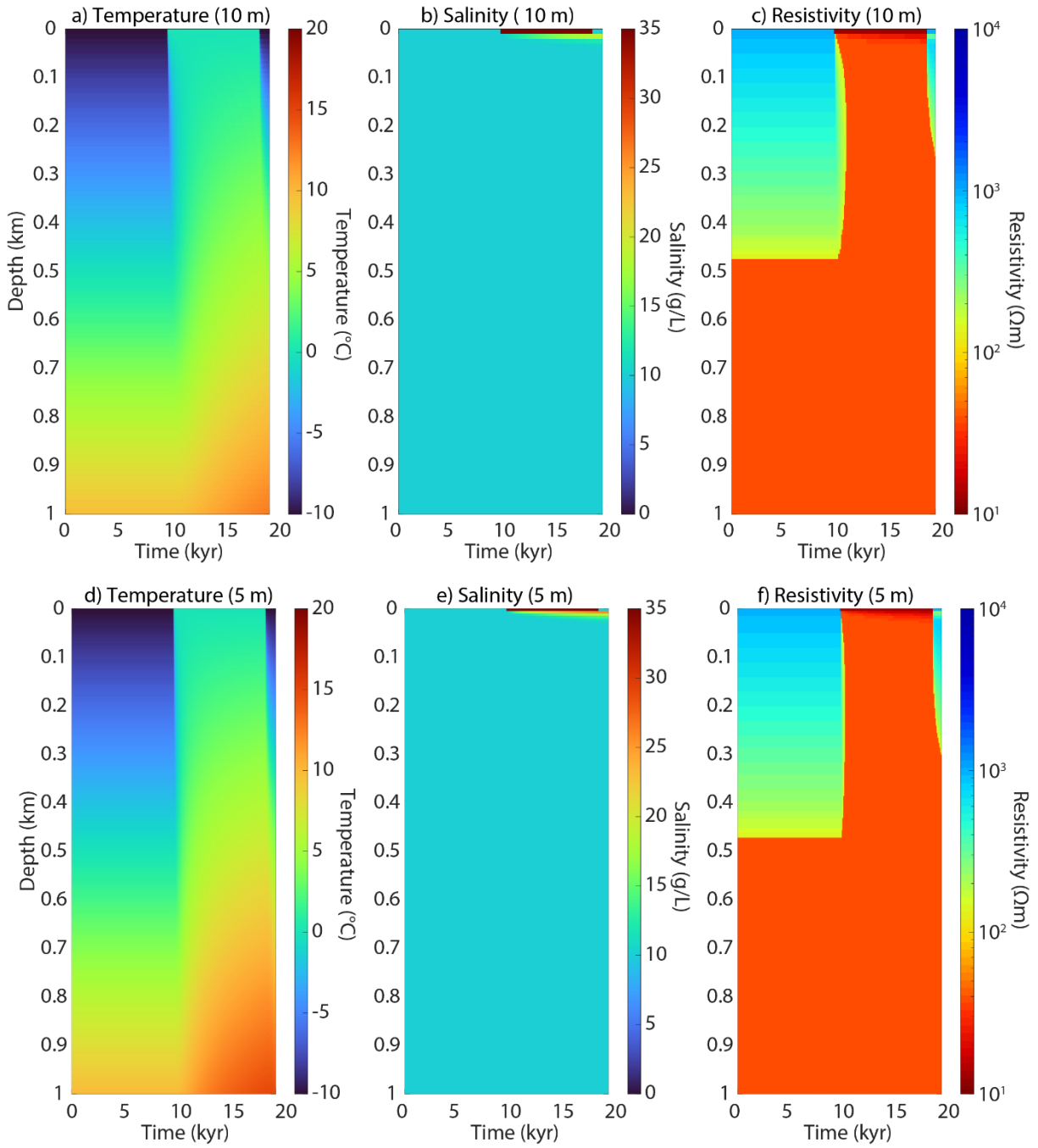


Figure 7.8: Simulation of the single glacial cycle scenario (Model 3) at all time steps down to depths of 1 km. Results are shown for simulations using $\Delta z = 10$ m and $\Delta z = 5$.

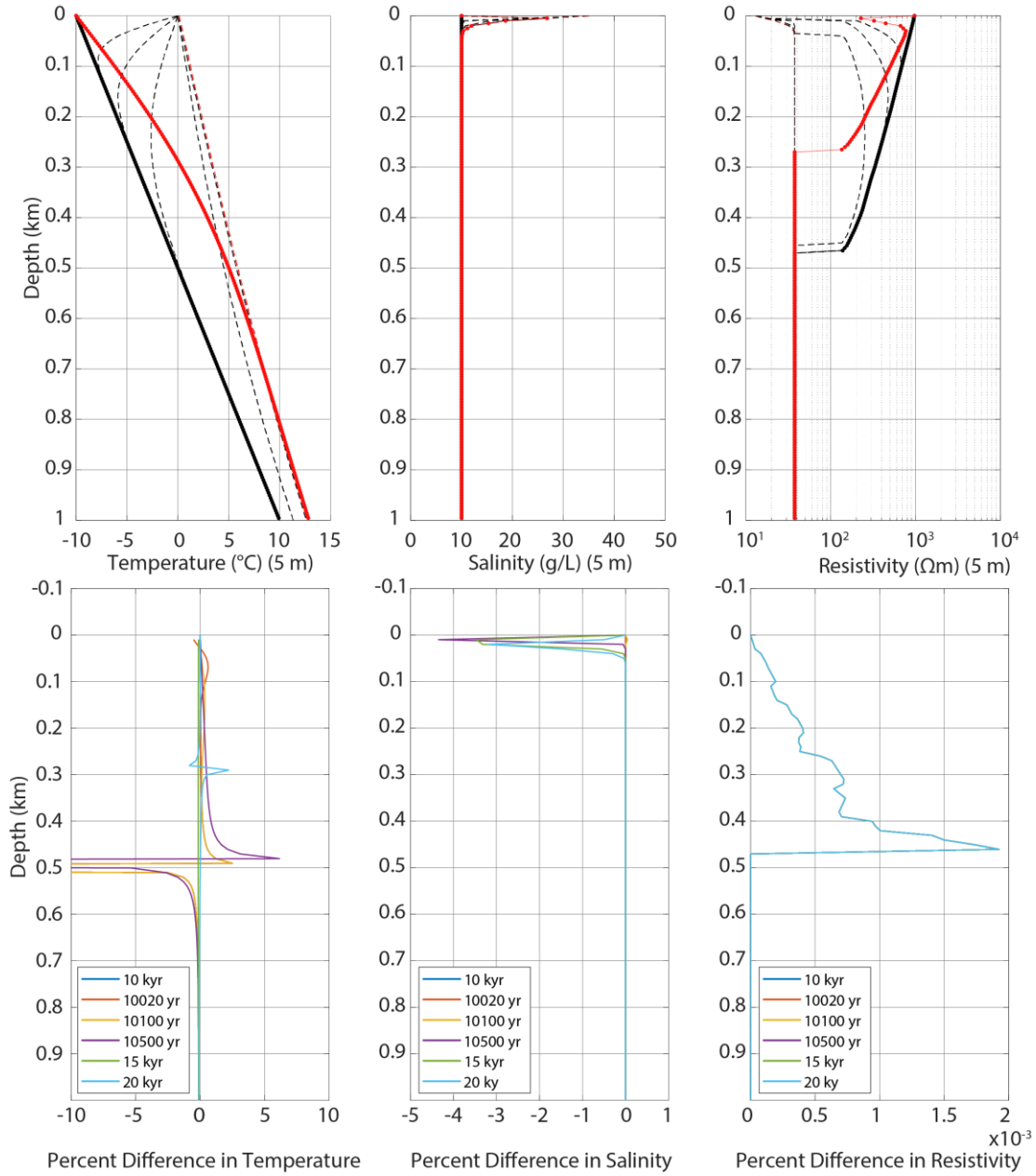


Figure 7.9: The starting and final models, as well as intermediate times, are shown for Model 3 for depths down to 1 km, using $\Delta z = 5$ m. The beginning and final models are represented by solid lines in black and red, respectively. Intermediate models are shown as dotted black lines for $t = 10.02$ kyr, $t = 10.1$ kyr, $t = 10.5$ kyr, $t = 15$ kyr. The percent differences between the models using $\Delta z = 5$ m and $\Delta z = 10$ m are also shown.

The results of the Model 3 experiment show that the temperature in the upper 1 km of the model changed very quickly, while the salinity was slow to respond to the change in conditions. It is clear from Figures 7.8 and 7.9 that 1000 years was not long enough for the temperature conditions to return to their original state. However, in the period from 19 – 20 kyr the layers near the surface refroze very quickly, limiting the diffusion of salt during this period. In a similar fashion to the Model 2 experiment, the differences between simulations using $\Delta z = 10$ m and $\Delta z = 5$ m are generally less than 5%. These results also suggest that changes in temperature at greater depths were limited and modelling could likely be accomplished using a shallower maximum depth for similar results. This would decrease the number of cells in the model space, reducing model size and computational costs.

This scenario was also tested using different time steps. Simulations were run using $\Delta t = 0.5$ yr and $\Delta t = 0.25$ yr, with no significant change to the results indicating that the temporal discretization of $\Delta t = 1$ yr is sufficient for this modelling.

The final resistivity model in Figure 7.9 shows three distinct behaviors: from the surface down to a depth of approximately 30 m the resistivity increased sharply due to the increasing salinity. At depths between ~30 - 250 m the resistivity gradually decreased due to increasing temperatures. At depths below ~250 m the resistivity sharply increased as the temperature rose above freezing and the Archie's Law method was used.

7.3.4. Model 4: multiple cycles of glaciation

To model the effects of a sequence of successive glaciation cycles, the simulation was run a number of times, each time using the previous final model as the starting model in the next simulation. The results of 10 successive cycles over 200 kyr are shown below. Figures 7.10 and 7.11 show the model results at all time steps. Figures 7.12 and 7.13 show how the effects of repeated cycles have diminishing effects on the downward diffusion of salt through the rock.

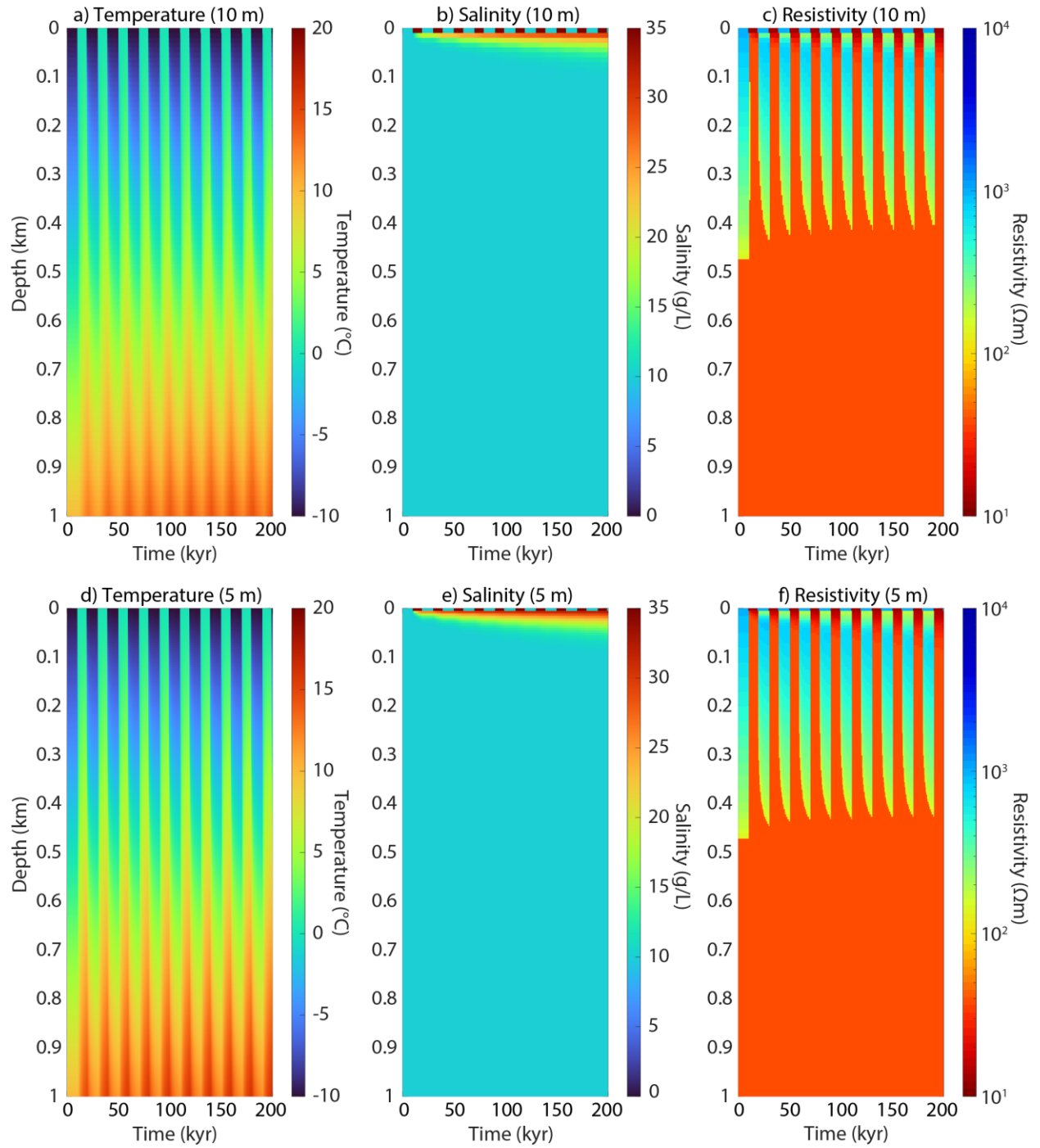


Figure 7.10: Simulation of the single glacial cycle scenario (Model 4) at all time steps down to depths of 1 km. Results are shown for simulations using $\Delta z = 10$ m and $\Delta z = 5$.

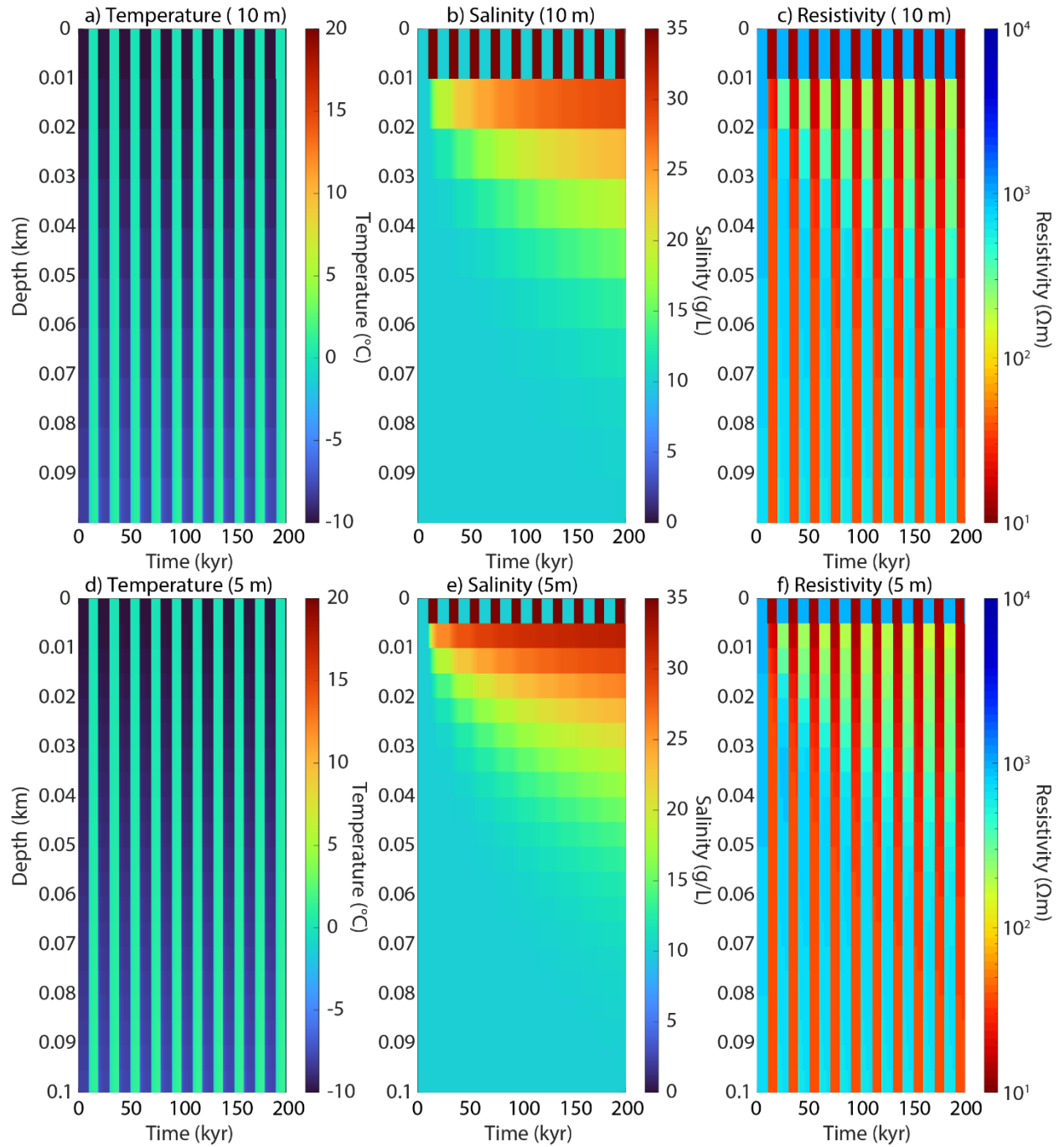


Figure 7.11: Simulation of the single glacial cycle scenario (Model 4) at all time steps down to depths of 100 m. Results are shown for simulations using $\Delta z = 10$ m and $\Delta z = 5$.

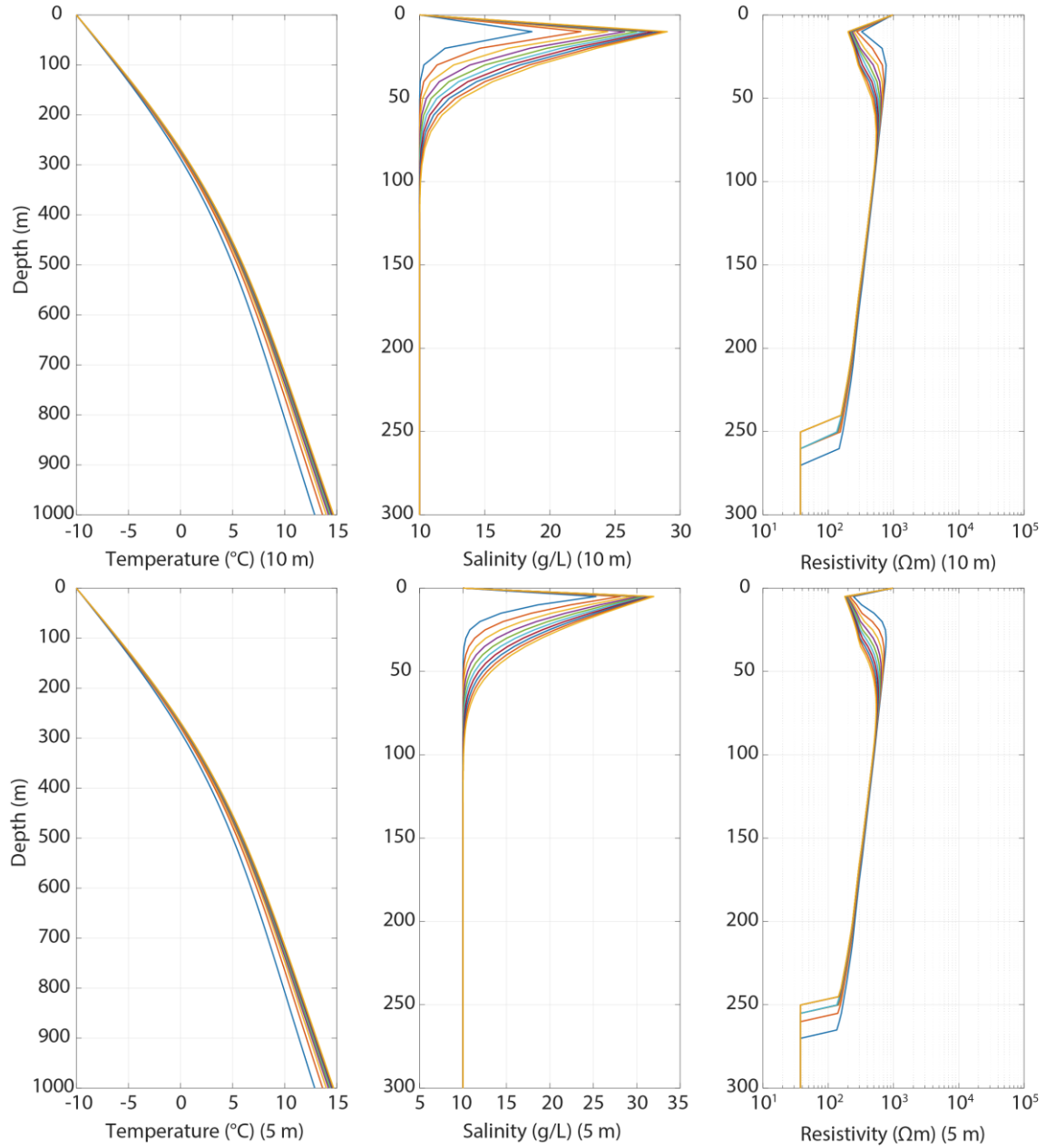


Figure 7.12: The final models for each of the 10 cycles are plotted for $\Delta z = 10$ m and $\Delta z = 5$. The model temperature values are shown to depths of 1 km. The model salinity and resistivity values are shown to depths of 300 m.

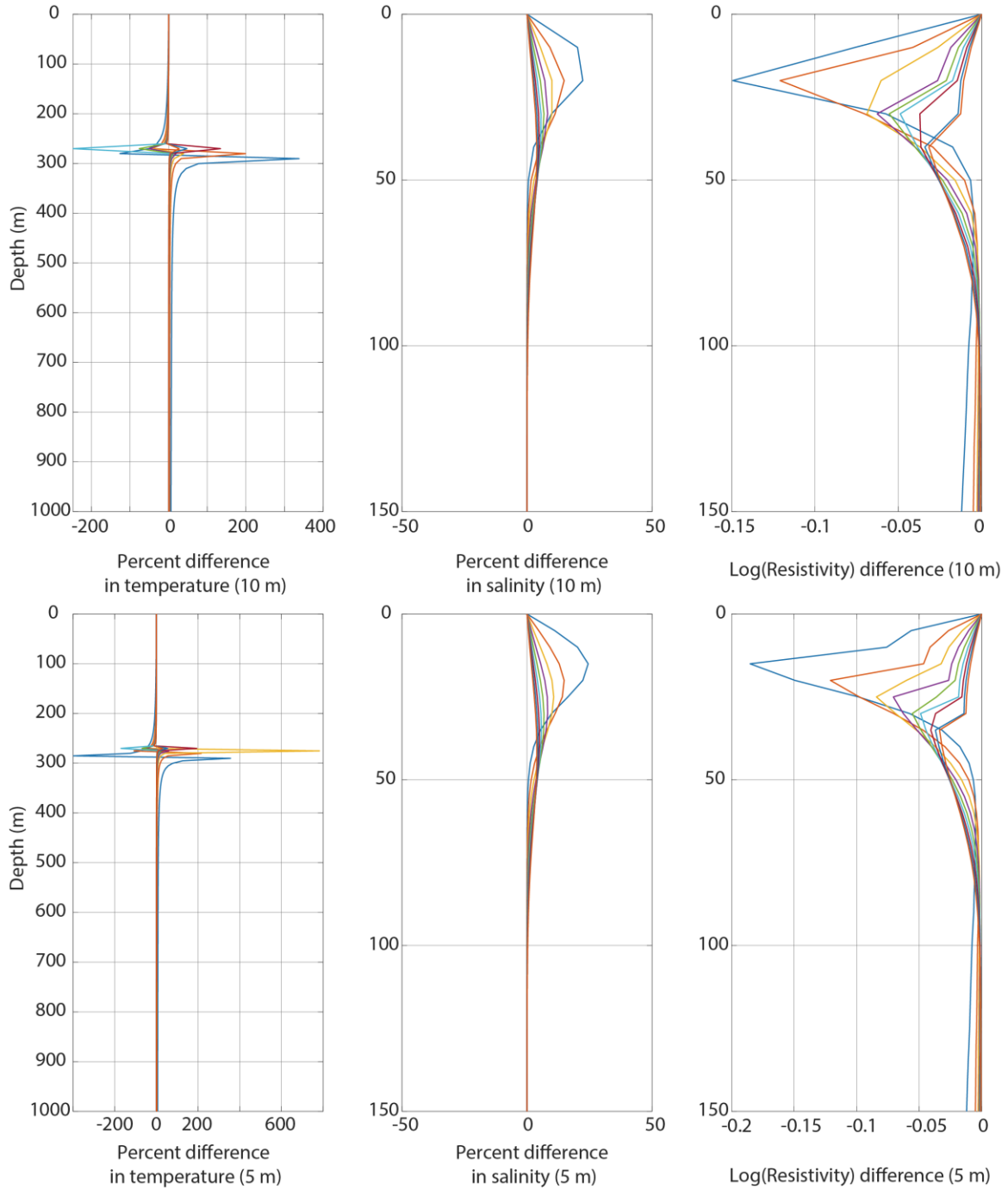


Figure 7.13: The percent difference between each cycle is shown for $\Delta z = 10$ m and $\Delta z = 5$. The percent difference in temperature is shown to depths of 1 km. The percent difference in salinity and $\text{Log}_{10}(\text{resistivity})$ are shown to depths of 300 m.

As can be seen in Figures 7.10 - 7.12, subsequent cycles allowed for further diffusion of salt through the rock. After 10 cycles the effects of high surface salinity began to approach approximately 100 m, although the changes at these depths are small. This would manifest as lower electrical resistivity values in rocks with higher salinity content. The differences between cycles (Figure 7.13) became smaller with additional iterations, which would be expected from this diffusion. If continued, one would expect diffusion to converge after many cycles as changes at the surface with sufficiently high frequency will not propagate through the entire model. The method of Pandit and King (1979) resistivity estimation shows relatively small changes in resistivity in the near-surface layers; however, these could become substantial given enough cycles

7.4. Comparison with MT models

It was important to compare the interpreted resistivity models from the numerical simulation with those obtained from MT modelling. Accordingly, the results of Occam 1-D inversion modelling and the preferred 3-D inversion model, *CB_inv13* were compared with the Model 4 simulation results. To simplify the comparison, the average of all 26 1-D inversion models, as well as the average resistivity of all stations on *CB_013* were used for comparison.

As is seen in Figure 7.14, the average 1-D and 3-D models were in good agreement with one another, as discussed in Chapter 5. These models started at $\sim 200 \Omega\text{m}$ at the surface, rising to $\sim 6000 \Omega\text{m}$ in the first kilometer. The average of the 3-D inversion models showed a small-scale decrease in resistivity to less than $100 \Omega\text{m}$ in the first 100 m, before rising again. This feature was not observed in the average of the 1-D models. The interpreted resistivity derived from Model 4 showed similar resistivity values at the surface, but a different resistivity pattern. Within the upper 100 m, the resistivity of the interpreted model rose sharply, due to the gradient in salinity. After approximately 100 m, the model decreased to values of approximately $150 \Omega\text{m}$, due to rising temperatures. At $\sim 250 \text{ m}$, where the ground thawed, the resistivity of the interpreted model was calculated using Archie's Law and decreased sharply to values of $\sim 40 \Omega\text{m}$, which remained constant at all depths beyond this point.

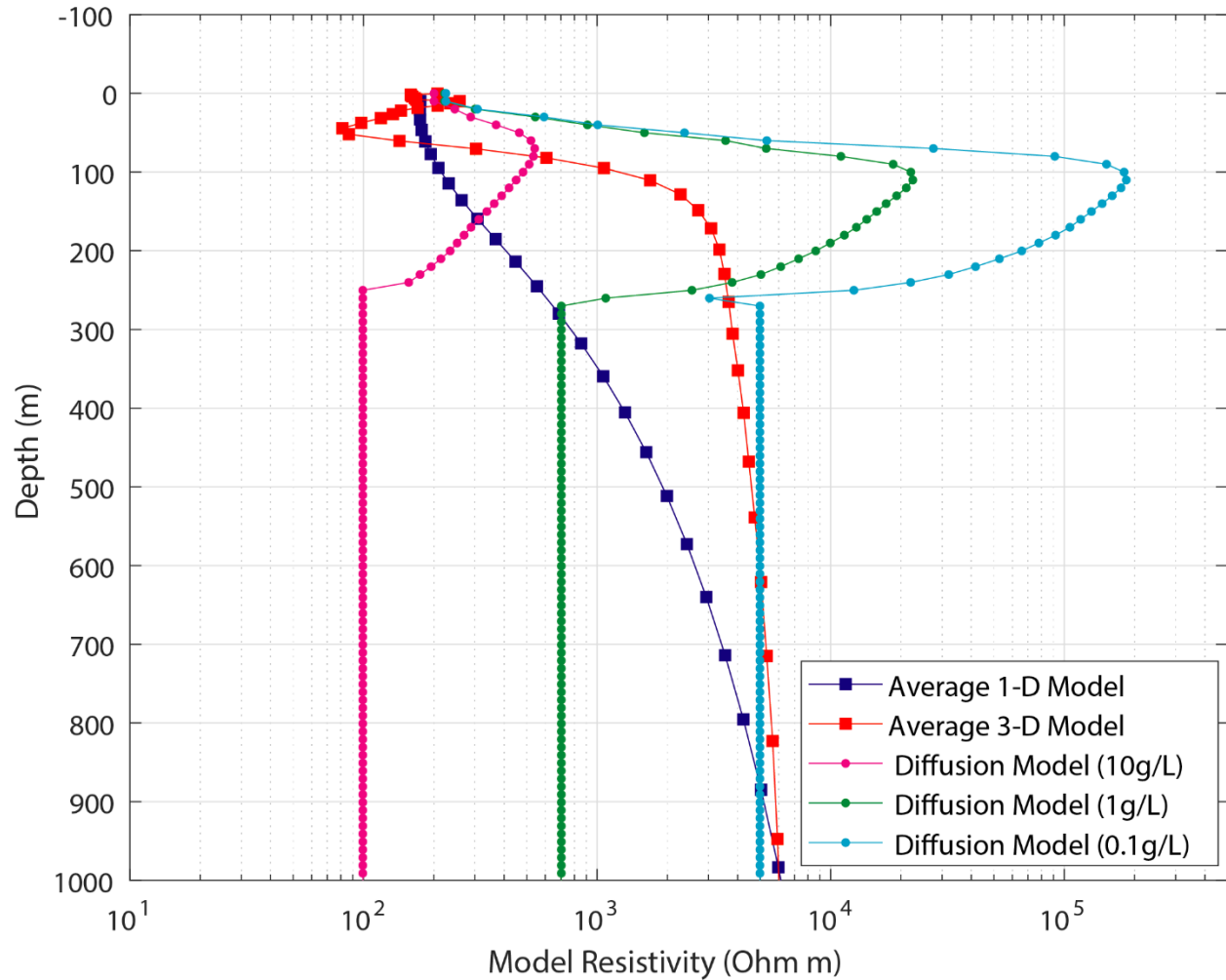


Figure 7.14: A plot comparing the diffusion model interpreted resistivity model with the resistivity models produced via MT inversion. The average of all 26 Occam 1-D inversion models for the Cambridge Bay MT dataset and the average of the model cells between each station of CB_inv13 are shown. The interpreted resistivity at the final time step of Model 4 is shown, as well as for additional Model 4 results using lower background salinity values of 1 g/L and 0.1 g/L.

Overall, the interpreted resistivity model produced by Model 4 fit the data reasonably within the first 250 m of the model where the rocks were assumed to be unfrozen. However, at depths that the salt did not reach by diffusion, the model was unable to accurately match the MT inversion models. In the final resistivity model obtained from the Model 4 simulation, the rocks were considered to be carbonate rocks, such as limestone and dolostone. Generally, the resistivity of such rocks that are unfrozen lie in the range of $\sim 10^3 - 10^5 \Omega\text{m}$ (Figure 3.5), which is much greater than the values that were interpreted using Archie's Law. One possible explanation for this

difference in resistivity values is that the quantities for rock porosity and salinity used may be too high to accurately reflect rocks at depth. According to Archie's Law (Eq. 3.5), reducing the porosity and/or salinity of rocks in the unfrozen layer can drastically increase the interpreted resistivity to levels that match the expected values more closely. To test this, the simulation was performed twice more, once using a background salinity of 1 g/L, and once with a background salinity of 0.1 g/L. The results of these simulations are also shown in Figure 7.14.

Decreasing the background salinity has multiple effects on the final interpreted resistivity. For one, increasing the contrast between the surface salinity and background salinity increased the rate of diffusion, as evidenced by the resistivity peaks at slightly greater depths for the models with lower background salinity. This effect was expected, as the rate of diffusion is related to both the diffusion coefficient of the material, and the contrast in salinity between neighboring cells in the simulation. The extent of this effect on the thickness of the saline layer was minimal, only altering the thickness of the layer by a few meters. More pronounced is the effect of decreasing the background salinity on the interpreted resistivity. In the 1 g/L model, the resistivity peak around 100 m reached approximately 20,000 Ωm , and in the 0.1 g/L model it reached nearly 200,000 Ωm . At depths directly below the maximum resistivity, both models exhibited decreasing interpreted resistivity due to increasing temperature with depth, similar to the original 10 g/L model. At ~ 250 – 300 m, where the models transitioned from frozen to unfrozen rock, the 1 g/L model assumed a constant resistivity profile of 700 Ωm to the base of the model, and the 0.1 g/L model assumed a constant resistivity profile of 5000 Ωm .

The resistivity values in the frozen layer that were estimated by the Model 4 were in relatively good agreement with the values produced in MT modelling, however there are some differences. The temperature-dependent effects that were visible on the interpreted resistivity from 100 m – 250 m were not strongly seen in the MT models, especially the 1-D models. When the background salinity used in the simulation was decreased to 0.1 g/L, the interpreted resistivity profile matched those of the MT inversion models quite well, although the peak resistivity between 50 – 250 m was much higher than either inversion model. Due to the overlying low-resistivity layer, it is unlikely that MT would be able to clearly resolve the thin high-resistivity layer between 50 and 250 m. It is likely that performing an extended simulation in which the salinity in the first 100 m

becomes more uniform, and then decreases between 100m – 250 m, could produce a resistivity model that more closely matches the resistivity from the MT models.

7.5. Discussion

The diffusion modelling experiments performed were successful in showing that salt can vertically diffuse distances of ~ 100 m under conditions that are expected of Cambridge Bay over the last few hundred thousand years. Final interpreted resistivity values using multiple simulated glacial cycles were able to reasonably match those obtained by MT inversion models in the shallow depths that were interpreted to be frozen, and values of porosity and salinity could be found that explained the resistivity values observed by the MT data in the deeper, unfrozen rock. Ultimately, the diffusion algorithm has a number of uncertainties and could likely be improved through several means.

Simple dimensional analysis of the diffusion equation allows one to estimate the timescale for a diffusion process. In the case of the temperature diffusion being considered in this experiment, one can estimate that the time, T for diffusion across a distance $L = 1$ km will be approximately

$$T = \frac{L^2}{k} \approx 25 \text{ kyr} \quad (7.9)$$

which is an approximation but is of similar magnitude to the effects seen in the diffusion modelling simulations.

The boundary conditions used in this diffusion modelling are also rough approximations of the true conditions over glacial and inter-glacial periods. Greater knowledge of these conditions would enable modelling that more accurately describes the surface conditions that existed due to the geological processes being studied. For instance, the duration of inundation of Victoria Island is not precisely known, and more accurate estimates could lead to a more accurate simulation of past events.

Currently, the algorithm considers diffusion as the only mechanism of heat and salt transport, but this is likely not the case. In the model that was considered the presence of water of varying salinity would likely introduce advective behavior, caused by buoyancy variations between water of different salinity. Even in a 1-D environment, these affects could enhance the transport of salt from the surface and accelerate the movement of salt from the surface.

Further, detailed knowledge of the Cambridge Bay geology and the resistivity of frozen rocks is scarce. Porosity values of 13% used in the simulations were not based off actual measurements of rocks at Cambridge Bay, but instead by the Salem limestone samples used by Pandit & King (1979). Using more accurate porosity values could significantly change both the rate of diffusion of salt, and the interpreted resistivity using Archie's Law. However, the resistivity of frozen rocks is likely dependent on the rock porosity, the extent of which is unknown. An extension of the results by Pandit and King (1979) to more completely describe how resistivity is affected by temperature, salinity, and porosity would be helpful in modelling the resistivity of frozen rocks. Currently, the algorithm uses a limited set of observations of frozen resistivity, which must be interpolated.

Finally, Model 4, the scenario that considered multiple successive glacial cycles, used 10 cycles over a period of 200 kyr. This model length was an arbitrary limit and was used only used because at this point the salt began to reach depths of approximately 100 m. Additional simulations could be run using a greater number of glacial cycles, which would enhance the salinity at greater depths, and increase the salinity further in the first 100 m, to a maximum value bounded by the surface. As was previously discussed, the rate of diffusion decreases over time such that the addition of more cycles to the simulation has diminishing effects on the depth that the salt diffuses.

While a number of improvements can be made to the algorithm used to model salt transport at Cambridge Bay, the existing model can be used to reasonably describe, to first order, the observations made using MT.

Chapter 8: Conclusion

8.1. Introduction

Geothermal energy resources provide a promising avenue for Canada to reduce its reliance on fossil fuels technology. In particular, geothermal technology could provide a clean, reliable source of energy in the Canadian Arctic, where diesel power is currently the dominant source of electricity and heating. While the Arctic sedimentary basin does not present extremely high geothermal temperatures that are typically required for conventional geothermal electricity production, the communities of Cambridge Bay and Resolute Bay, Nunavut could adopt geothermal for direct use applications.

An ideal setting for geothermal development requires the following: (1) significant subsurface temperatures, (2) subsurface rocks with significant permeability to enable the storage and circulation of fluids, and (3) fluids such as water or brines that can be used as a working fluid for heat extraction. Both Cambridge Bay and Resolute Bay lie within the Arctic sedimentary basin, which is a region that has generally low geothermal gradients, $\sim 20^{\circ}\text{C}/\text{km}$ (Minnick et al., 2018). However, previous data gathered during drilling at Resolute Bay has suggested the presence of an anomalously high local geothermal gradient of approximately $40^{\circ}\text{C}/\text{km}$ (Misener, 1955). Additionally, both locations are situated on the coast, providing potential for seawater to have intruded into the local rock. A combination of elevated subsurface temperatures, as well as substantial aquifer volume would be conducive for the development of economic geothermal infrastructure. To investigate the subsurface conditions at Cambridge Bay and Resolute Bay, MT data were collected at both communities in 2022 and 2023 to produce subsurface resistivity models. Resistivity is a property that is sensitive to temperature and water content, and so generated resistivity models can be used to assess these characteristics. The MT results were used to evaluate whether geothermal is feasible at these communities. To follow is a summary of the results of this investigation, as well as suggestions for future study.

8.2. Summary of MT investigations at Cambridge Bay, Nunavut

In July 2023, broadband MT data were collected at 26 stations at Cambridge Bay, Nunavut in the frequency band 500 – 0.001 Hz. The data were collected in a $\sim 17 \times 8$ km area surrounding the community. Observations of apparent resistivity and phase curves, induction vectors, and phase

tensors were used to determine the most appropriate techniques for modelling. The induction vectors at Cambridge Bay were dominated by the coast effects of the Arctic Ocean; however, the apparent resistivity and phase curves, and phase tensors suggested that the data were relatively 1-D at most frequencies sampled. The dataset was used to produce a series of simple forward models to assess the subsurface conditions, and then inverted in both 1-D and 3-D to produce more complex subsurface resistivity models. Analysis and interpretation of the modelling results revealed a simple 1-D resistivity structure consisting of: (1) CB1, a persistent 100 m thick surface layer with low resistivity in the range 30 – 100 Ωm that was interpreted to represent frozen limestone with partially frozen saline pore water and (2) CB2, a deeper high-resistivity (1000 – 10,000 Ωm) layer that was identified as limestone containing low-salinity pore water. No subsurface aquifers were identified in the MT models. Analysis of these results required knowledge of the electrical properties of frozen rocks. Estimates of the porosity at Cambridge Bay were low, indicating that there is poor natural reservoir potential in the rocks below the settlement.

The thin low-resistivity layer, CB1, was attributed to seawater that had diffused into the surface rock during past periods of inundation. Additional diffusion modelling was performed to verify this hypothesis, and included a number of simulations of the subsurface temperature and salinity over periods of changing surface conditions. The finite difference algorithm simultaneously solved the diffusion equation for temperature and salinity over a 20 kyr period. Boundary conditions were chosen to simulate the effects of a glacial-interglacial cycle. The first stage used a 10 kyr period with a surface temperature of -10°C and a surface salinity of 10 g/L. The second stage used a 9 kyr period with higher surface temperatures of 0°C and a surface salinity of 35 g/L to simulate the inundation of the surface with seawater. Finally, the surface conditions were returned to their initial state for a period of 1 kyr. Modelling results showed that over a single cycle, salt was able to diffuse from the surface through limestone to depths of approximately 30 m. Using 10 consecutive cycles, the salt was able to diffuse from the surface to depths of approximately 100 m over a period of 200 kyr. The resistivity of the model was interpreted using two methods: the experimental results of Pandit and King (1979) for model cells that were below the freezing point, and Archie's Law for model cells that were above the freezing point.

The interpreted resistivity model that was generated from the simulation results was compared to the values observed in the 1-D and 3-D MT inversion models. It was found that simulation using

an initial model salinity of 10 g/L produced resistivity values that matched the MT models at the surface, but were much lower at greater depths. Additional simulations were performed that used background salinity values of 1 g/L and 0.1 g/L, finding the 0.1 g/L model to reasonably match the MT models. Thus, the diffusion modelling results were consistent, to first order, with the interpretation of the MT resistivity models produced for Cambridge Bay.

8.3. Summary of MT investigations at Resolute Bay, Nunavut

At Resolute Bay, data were collected at 44 MT stations over two deployments from 2022 – 2023. Data were collected in the frequency band of 909 – 0.0016 Hz. The 11 stations collected in 2022 exhibited generally poorer data quality than those collected the following year. Due to significant snow cover during the 2022 data collection, different equipment was used for the collection of electric field data in each deployment, which may have influenced the differences in observed data quality. Significant human structures around the community introduced significant cultural noise and further impacted the quality of the collected data.

Dimensional analysis of this dataset suggested a relatively complex subsurface resistivity structure at Resolute Bay. The induction vectors were dominated by coast effects; however, the apparent resistivity and phase curves, and phase tensors suggested that the subsurface structure was 2-D or 3-D. Inversion modelling was performed on the dataset to generate a series of 1-D and 3-D inversion models. At Resolute Bay, the preferred resistivity model was characterized by three layers: (1) RB1, a high-resistivity (100 – 1000 Ωm) limestone layer that was 250 – 500 m thick, containing partially frozen saline pore water (2) RB2, a low-resistivity (10 – 30 Ωm) layer that was ~ 500 m thick, containing unfrozen saline pore fluids, and (3) RB3, a deeper high-resistivity (100 - 1000 Ωm) layer, containing low salinity pore water. RB1 was observed to thicken with distance away from the coast, and was generally consistent with what would be observed of permafrost if the local geothermal gradient were ~ 20°C/km, lower than previously suggested. Assuming a lower geothermal gradient, the resistivity modelling was used to estimate rock porosities of ~ 20 – 30% in RB2. Consequently, the subsurface below Resolute Bay could have significant reservoir potential at depths of 500 – 1000 m, but should be verified by drilling.

8.4. Implications for geothermal development at Cambridge Bay and Resolute Bay

As addressed above, the MT data collected at Cambridge Bay showed no evidence of elevated geothermal temperatures or subsurface aquifers. These results, as well as the presence of permafrost at this location would make the development of geothermal less appealing. Overall evidence suggests that Cambridge Bay is poorly suited for conventional geothermal development. Engineered geothermal systems, heat pumps, and other direct-use applications could provide suitable alternatives for the community.

The magnetotelluric data collected at Resolute Bay have revealed a near surface structure that is significantly different to that at Cambridge Bay. The near surface layer (RB1) at Resolute Bay can be interpreted as a layer of frozen carbonate rocks with a salinity of 10 g/L. This salinity is significantly lower than in the surface layer at Cambridge Bay, and may reflect the differing sea level histories of the two locations over the Holocene (Simon et al., 2016). A layer of unfrozen carbonate rocks containing saline pore fluids is inferred to be present in the depth range 500 – 1000 m. This could represent a layer of rocks suitable for heat extraction or storage, such as with a solar augmented geothermal system. The heat contained in this layer will be highly dependent on the geothermal gradient, and this quantity should be determined through drilling. Confirmation of the porosity and salinity is also required before any geothermal development can proceed.

Future development of geothermal infrastructure at Cambridge Bay and Resolute Bay is largely an economic problem. While the potential for conventional geothermal at these communities is low, temperatures that could be used for alternative, direct use geothermal systems likely exist. Further studies are required to assess whether the installation of geothermal at these communities is economic, particularly with the significant costs associated with such development in the Arctic environment.

8.5. Future studies

To further investigate the near-surface layer (CB1) observed at Cambridge Bay, observations of the shallow subsurface could be performed at additional locations to learn the extent of this feature, and whether it is regional, or confined to the area around Cambridge Bay. Airborne EM data exists for other regions on Victoria Island and could provide additional insight to the geology of the island. Additionally, Cambridge Bay has essentially no varying topography. Additional

observations at nearby locations with varying surface elevation could provide information on whether the structure is topographically controlled. Locally, other electrical methods, such as DC resistivity, induced polarization, or TDEM could be utilized for further investigation of the first 100 – 200 m of the subsurface. Physical sampling of the near-surface salinity at Cambridge Bay could be done to validate the results obtained from MT data.

Diffusion modelling of the processes that may have resulted in the formation of CB1 were successful in validating the hypothesis formed during interpretation of the MT models. To first order, these simulations were able to explain how geological processes may have resulted in what was observed. The diffusion algorithm could be refined and extended to more accurately simulate these processes. For one, additional mechanisms of heat and salt transport could be implemented. Additionally, a 1-D model cannot completely describe the processes that occur in a 3-D environment, thus extension of the modelling to 2-D or 3-D may result in more accurate results.

At Resolute Bay, the results of the MT study suggest that although geothermal gradients are likely relatively low, subsurface porosities may be sufficient for development of certain geothermal systems. However, these conclusions are subject to a number of uncertainties, and it is recommended that these results are verified using additional means, such as drilling.

The study of geothermal potential at both Cambridge Bay and Resolute Bay were localized to the immediate areas around the settlements. Additional studies could extend this investigation to regions that are further away. However, the cost of developing geothermal infrastructure will increase with distance from the communities.

Lastly, the experimental results of Pandit and King, (1979) were used in the interpretation of resistivity models for rocks that were thought to be frozen. While these results were helpful, the original study was limited in scope, and is unable to describe certain effects – the effect of changing porosity on the resistivity of the rocks, for instance. A more complete set of observations about the resistivity of frozen rocks could provide a powerful tool for interpreting measurements of the subsurface resistivity.

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Appendix

A. Supplementary material for Chapter 5

Station	NS (m)	EW (m)	Latitude	Longitude	Configuration	Tipper
CB001	55	60	69.11366	-104.9755	2E-2H	N
CB002	57	55	69.11171	-105.0037	2E-2H	N
CB003	59	59	69.15492	-104.9074	2E-2H	N
CB004	53	61	69.14189	-104.9434	2E-2H	N
CB005	65	51	69.14784	-104.9199	2E-2H	N
CB006	57	61	69.11175	-105.2171	2E-2H	N
CB007	58	57	69.11338	-105.1885	2E-2H	N
CB008	52	63	69.13758	-105.1651	2E-3H	Y
CB009	57	62	69.1059	-105.2425	2E-2H	N
CB010	58	57	69.13296	-105.1364	2E-3H	Y
CB011	64	50	69.11549	-105.1408	2E-2H	Y
CB012	61	58	69.14431	-105.181	2E	N
CB013	50	47	69.12696	-104.9914	2E-3H	Y
CB014	56	61	69.13464	-105.0692	2E-3H	Y
CB015	50	63	69.13312	-104.9725	2E-3H	Y
CB016	60	58	69.12623	-105.1078	2E-3H	Y
CB017	66	50	69.12608	-105.1254	2E	N
CB018	61	50	69.13121	-105.036	2E-3H	Y
CB019	61	50	69.1034	-105.2665	2E-3H	Y
CB020	63	64	69.1291	-105.0113	2E-3H	Y
CB021	62	50	69.10006	-105.2919	2E-3H	Y
CB022	62	47	69.10705	-105.0976	2E	N
CB023	47	60	69.11587	-105.0875	2E	N
CB024	63	62	69.1063	-105.0302	2E-3H	Y
CB025	53	46	69.12653	-105.0779	2E	N
CB026	47	61	69.11993	-105.0066	2E	N

Table A1: A summary of the data collected at Cambridge Bay in July 2023. The location and configuration of each station is shown.

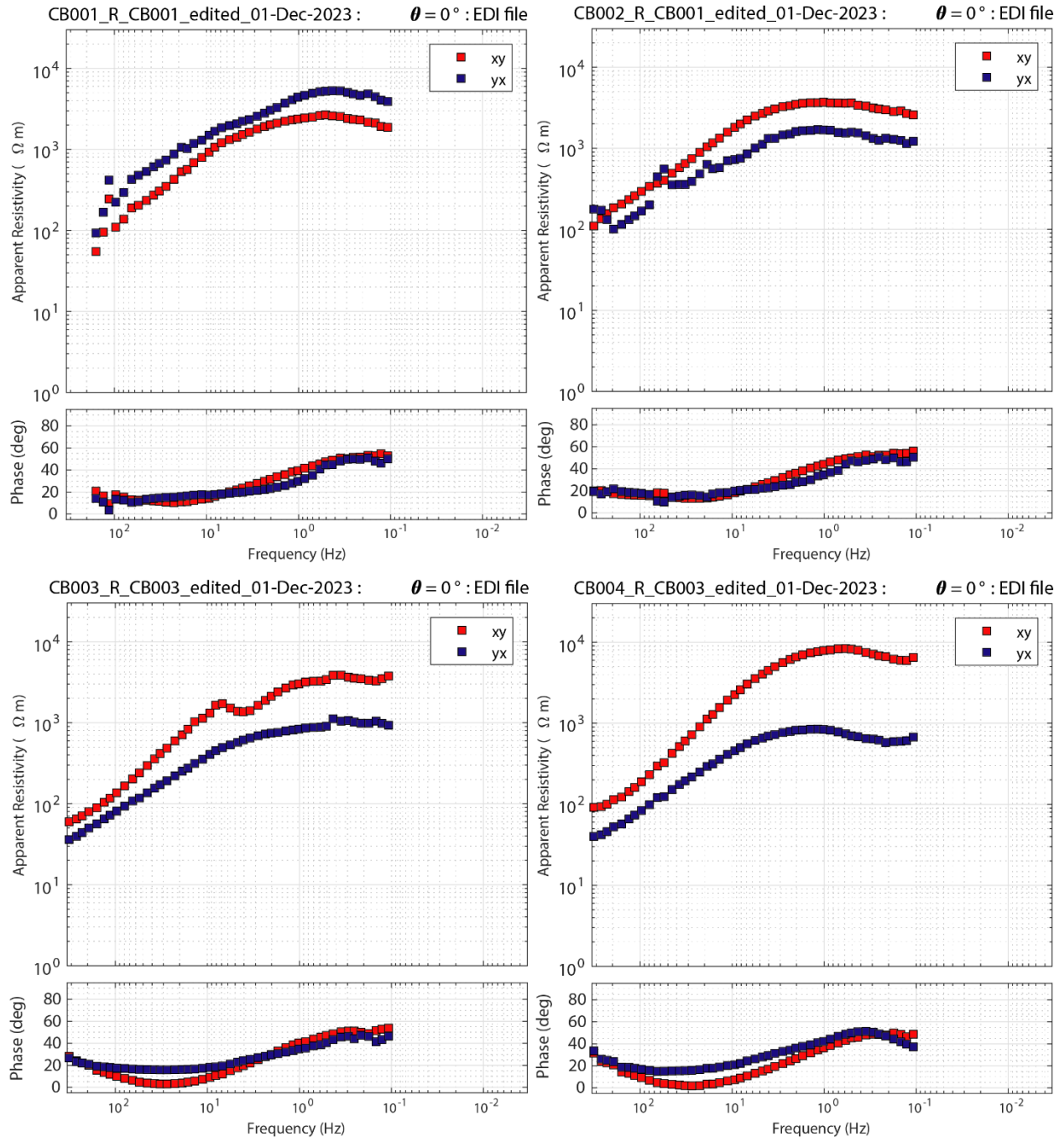


Figure A1: MT data at stations CB001, CB002, CB003, and CB004.

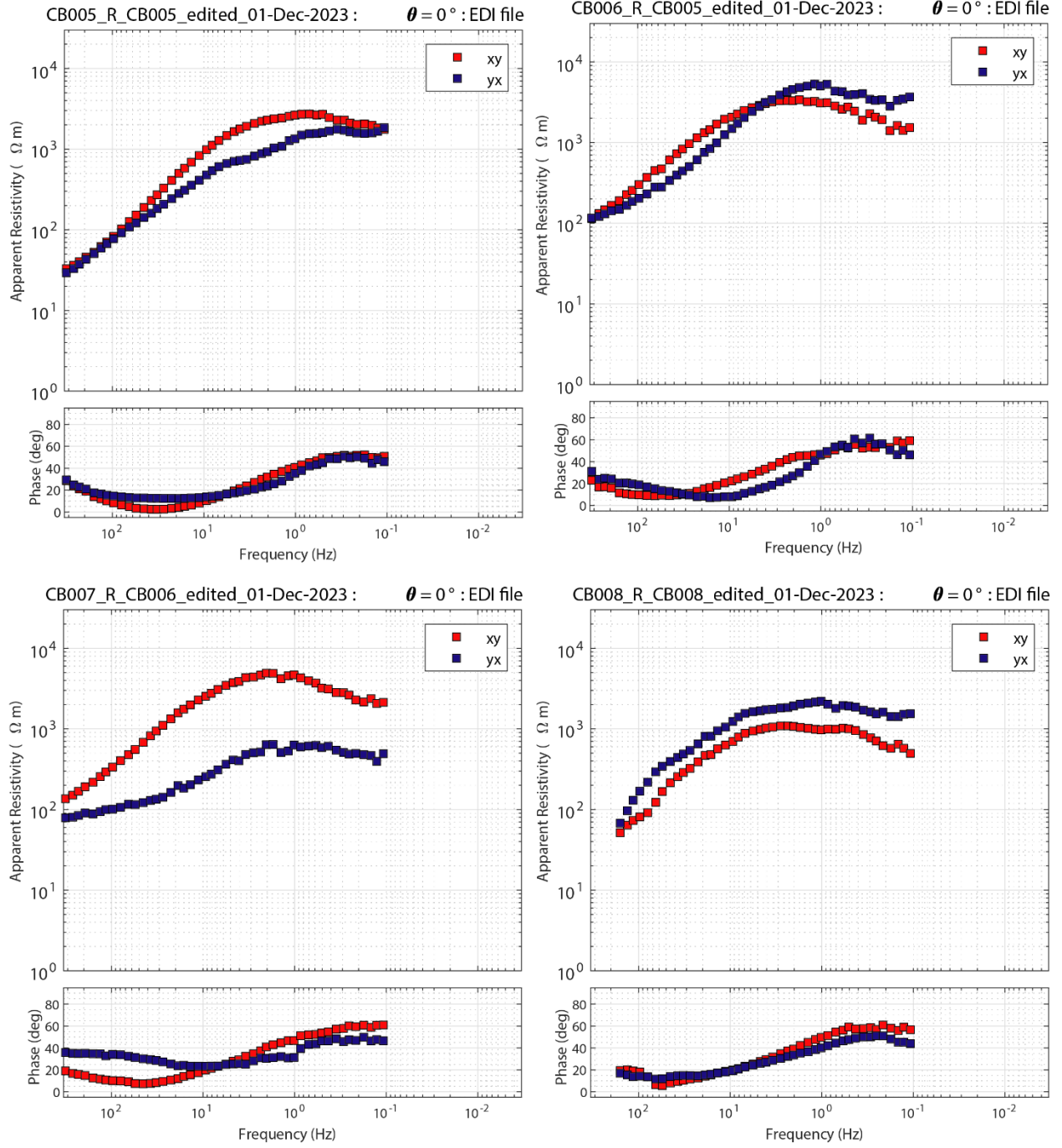


Figure A2: MT data at stations CB005, CB006, CB007, and CB008.

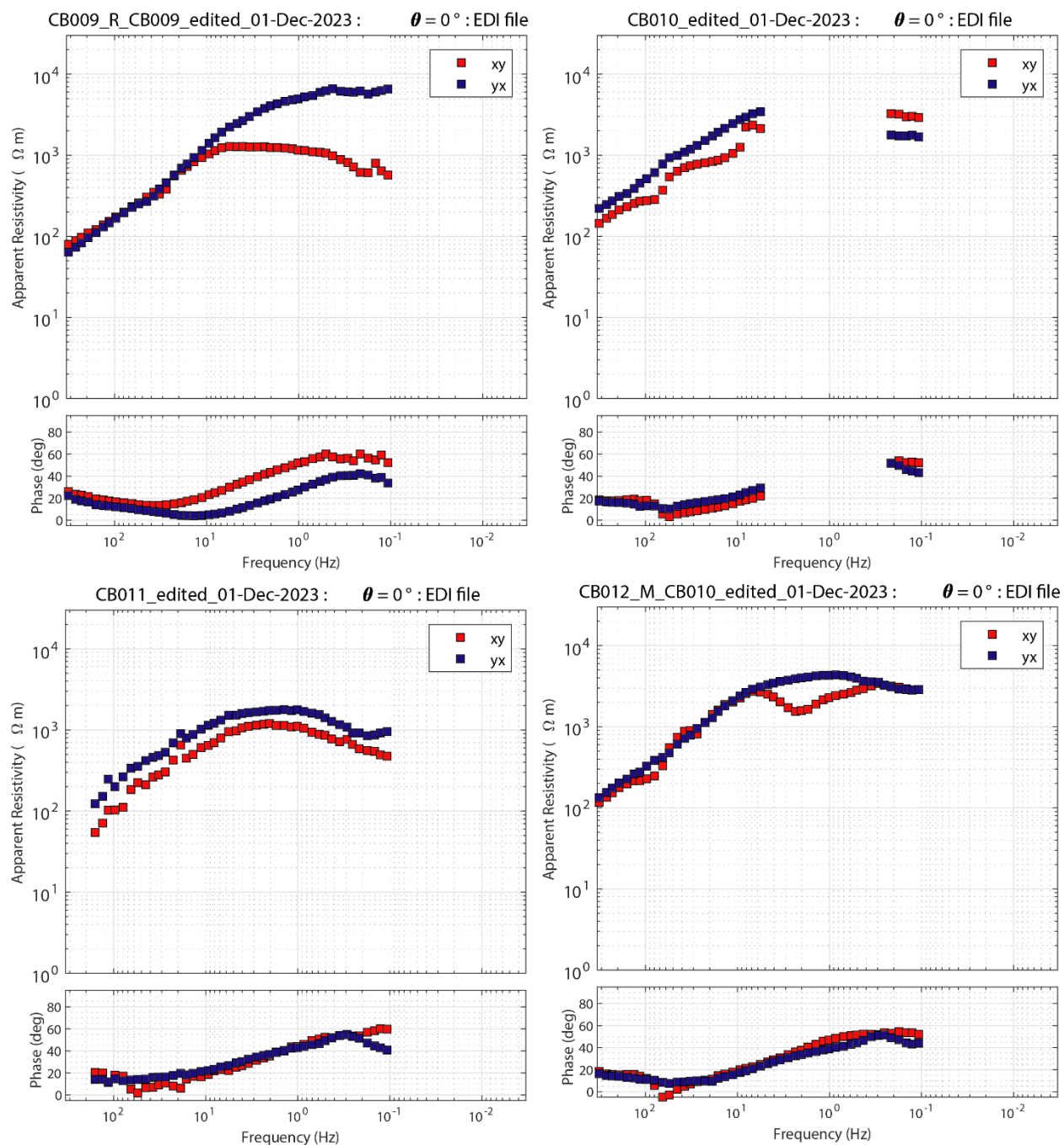


Figure A3: MT data at stations CB009, CB010, CB011, and CB012.

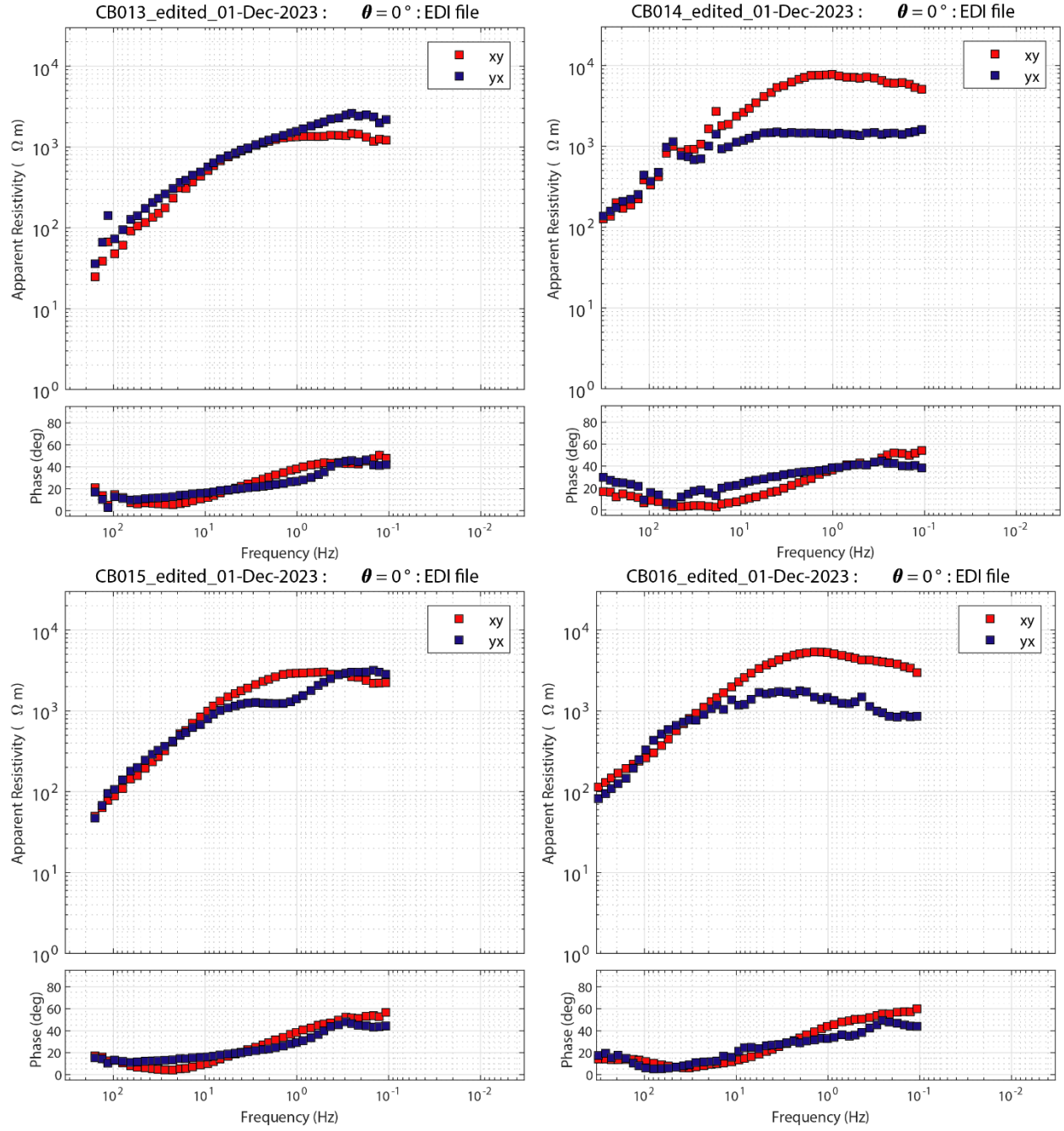


Figure A4: MT data at stations CB013, CB014, CB015, and CB016.

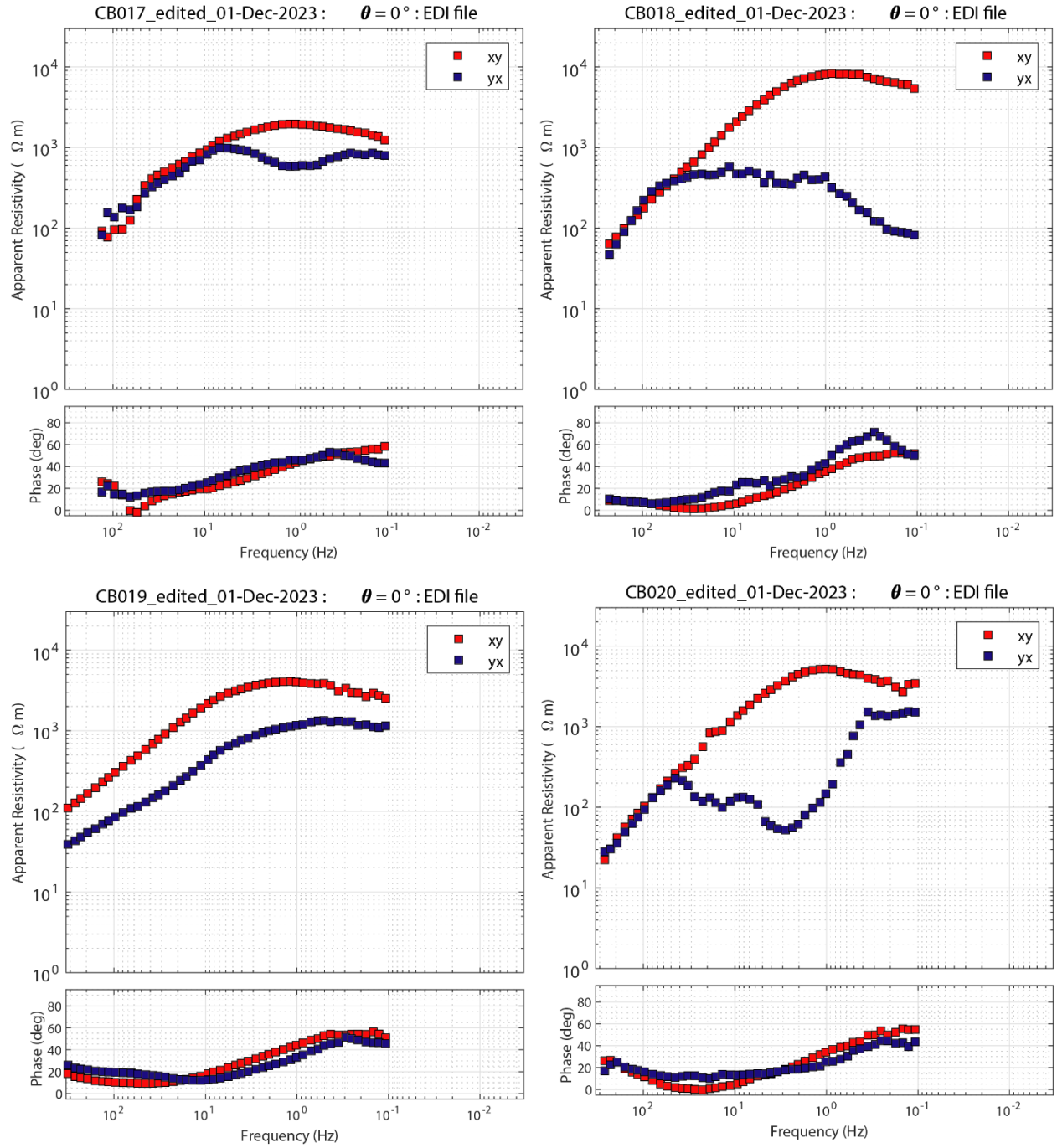


Figure A5: MT data at stations CB017, CB018, CB019, and CB020.

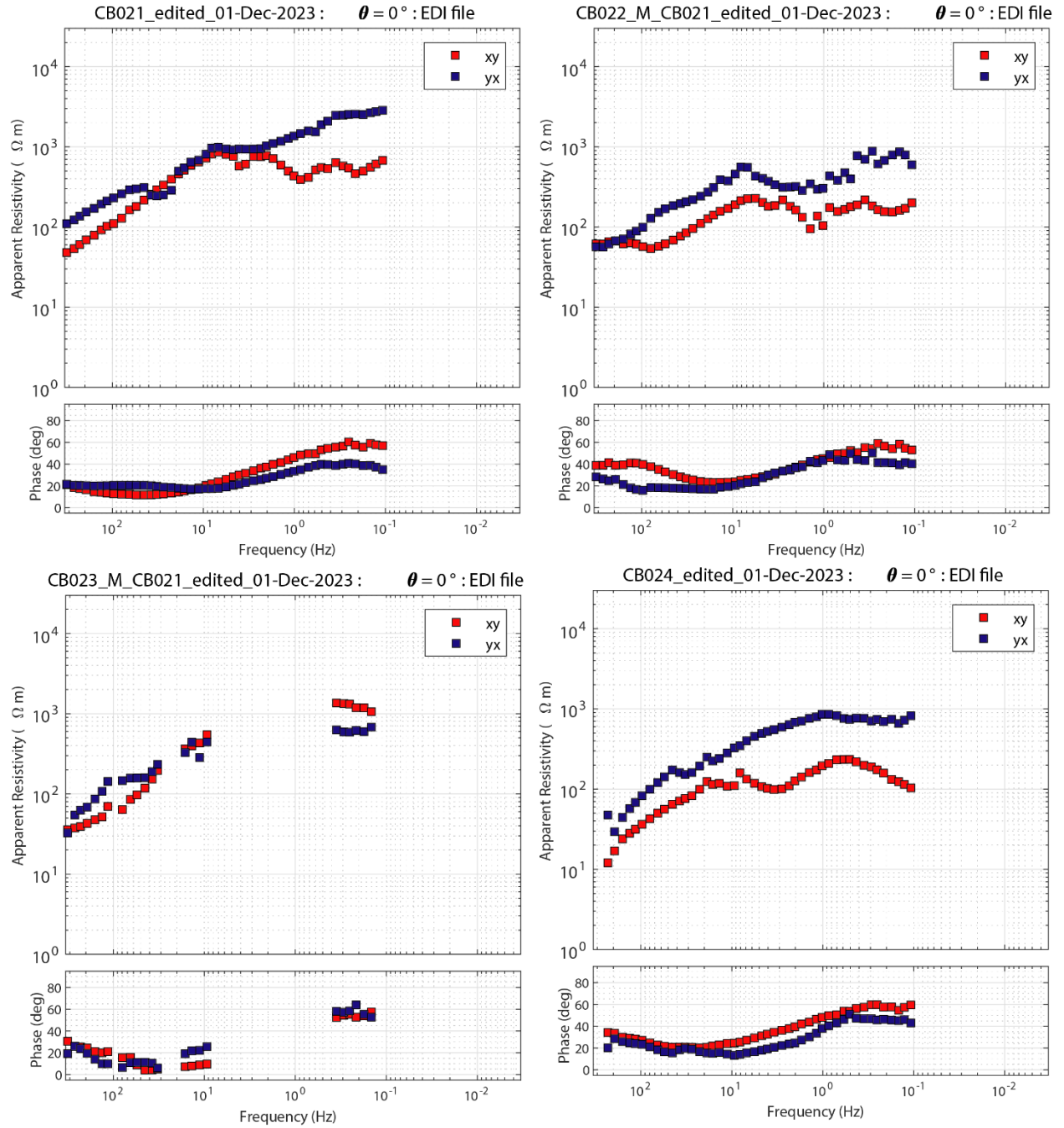


Figure A6: MT data at stations CB021, CB022, CB023, and CB024.

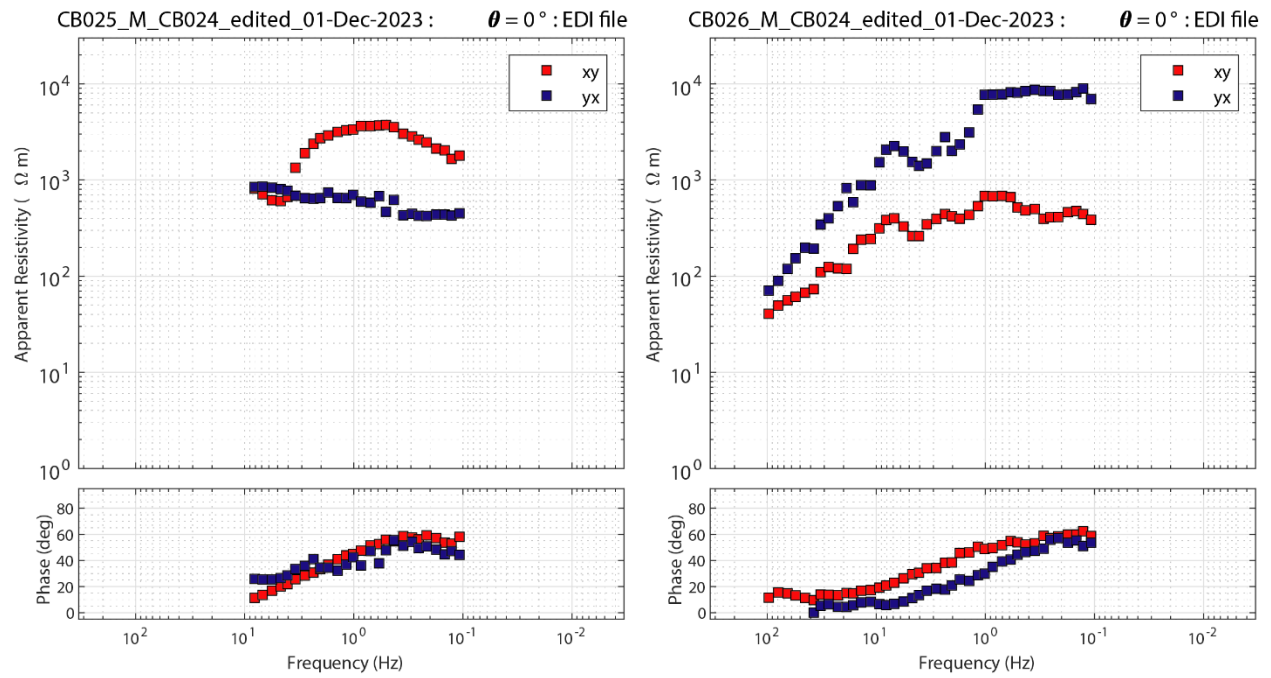


Figure A7: MT data at stations CB025 and CB026.

Station	Error Floor	Final RMS Misfit
CB001	5%	1.1986
CB002	5%	0.99723
CB003	10%	0.99853
CB004	5%	0.99968
CB005	5%	0.99835
CB006	5%	0.99795
CB007	5%	0.99942
CB008	5%	0.9995
CB009	5%	0.99784
CB010	5%	0.99769
CB011	5%	0.9998
CB012	5%	0.9994
CB013	5%	1.0795
CB014	5%	0.99851
CB015	5%	0.99939
CB016	5%	1.00
CB017	5%	0.99953
CB018	5%	0.99926
CB019	5%	0.99618
CB020	5%	0.99972
CB021	10%	0.99997
CB022	5%	1.904
CB023	10%	0.99795
CB024	10%	0.99655
CB025	5%	1.5464
CB026	10%	3.3343

Table A2: Summary of 1-D inversions for Cambridge Bay using the OCCAM inversion algorithm of Constable et al., (1987).

Name	# Stations	NX	NY	NZ	Topography	Bathymetry	Starting model	Regularization Parameter, λ	covariance (default 0.5)	start RMS	end RMS	Iterations
CB_inv01	26	54	100	40	n	n	100 Ω m h.s.	1	0.5	6.42	2.018	107
CB_inv02	26	54	100	40	n	n	100 Ω m h.s.	1	0.5	4.59	1.089	146
CB_inv03	26	54	100	40	n	n	100 Ω m h.s.	1	0.5	4.65	17.085	10
CB_inv04	26	54	100	62	n	n	100 Ω m h.s.	1	0.5	4.65	2.145	94
CB_inv05	26	54	100	65	n	n	100 Ω m h.s.	1	0.5	4.63	0.955	104
Cb_inv06	26	54	100	65	n	n	(100 m , 100 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	0.5	4.56	1.042	135
CB_inv07 2	26	54	100	65	n	n	(100 m ,10 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	0.5	4.60	0.965	108
CB_inv08	26	54	100	75	n	n	(100 m ,10 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	0.5	4.30	1.016	126

Table A3: Summary of 3-D inversions CB_inv01 – CB_inv08 for Cambridge Bay using the ModEM inversion algorithm of Kelbert et al., (2014).

Name	# Stations	NX	NY	NZ	Topography	Bathymetry	Starting model	Regularization Parameter, λ	covariance (default 0.5)	start RMS	end RMS	Iterations
CB_inv09	25	54	100	75	n	n	(100 m ,10 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	x,y 0.4	4.30	1.18	106
CB_inv10	25	54	100	75	n	n	(100 m ,10 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	x,y 0.6	4.30	1.377	115
CB_inv11	25	84	176	90	y	y	(100 m ,100 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	0.5	4.11	1.657	40
CB_inv12	25	54	100	90	y	y	(100 m ,100 Ω m); (9.9 km, 1000 Ω m); (100 Ω m h.s.)	1	x,y 0.8	4.15	0.955	141
CB_inv13	25	54	100	90	y	y	100 Ω m h.s.	1	0.5	4.57	0.919	122
CB_inv14	25	54	100	90	y	y	100 Ω m h.s.	1	0.3	4.57	0.865	107
CB_inv15	25	54	100	90	y	y	100 Ω m h.s.	1	0.5	4.57	0.893	111

Table A4: Summary of 3-D inversions CB_inv09 – CB_inv15 for Cambridge Bay using the ModEM inversion algorithm of Kelbert et al., (2014).

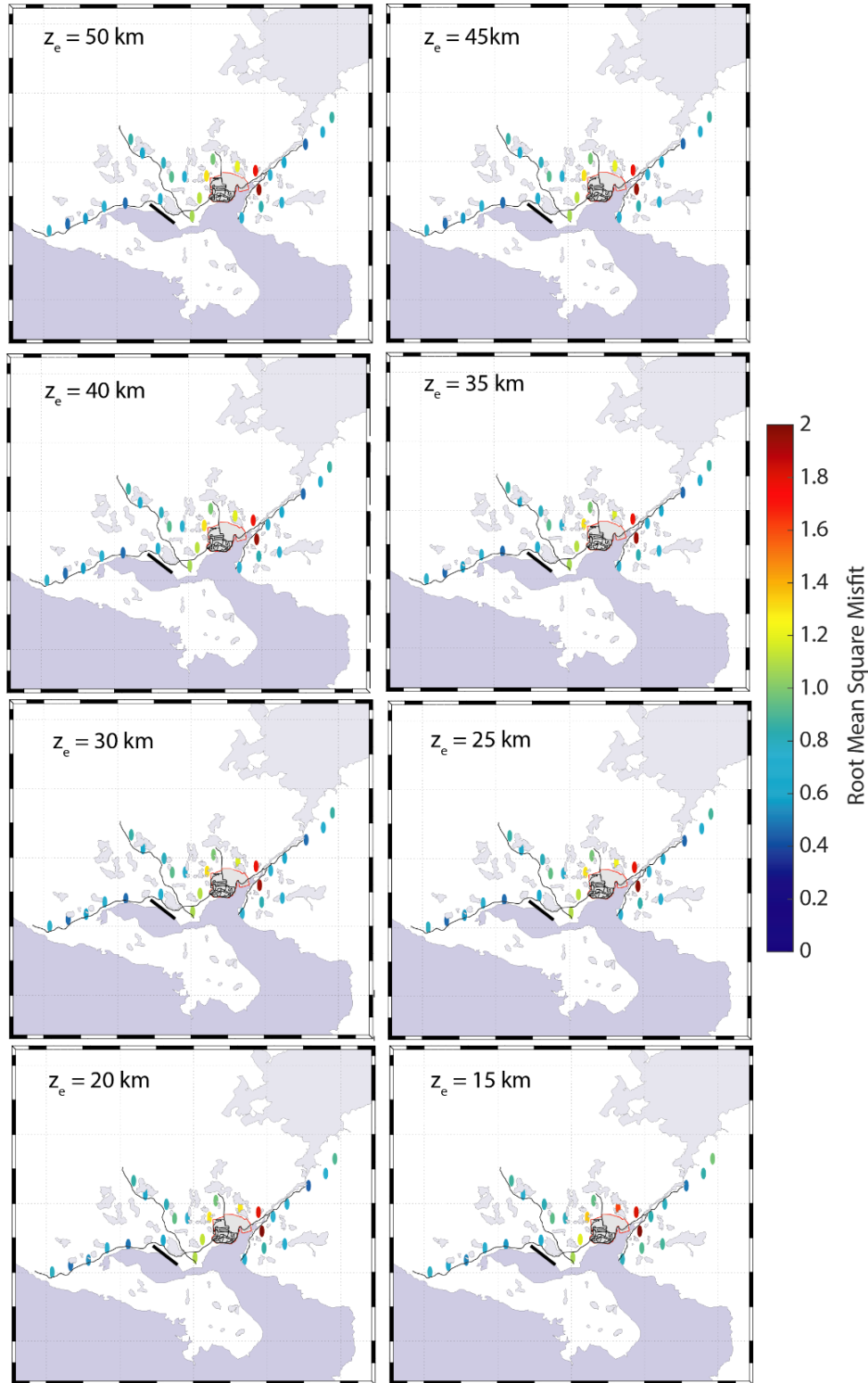


Figure A8: RMS misfit maps for edited models. The model *CB_inv13* was edited to a $1000 \Omega m$ half-space below a depth at depths between 50 km and 15 km.

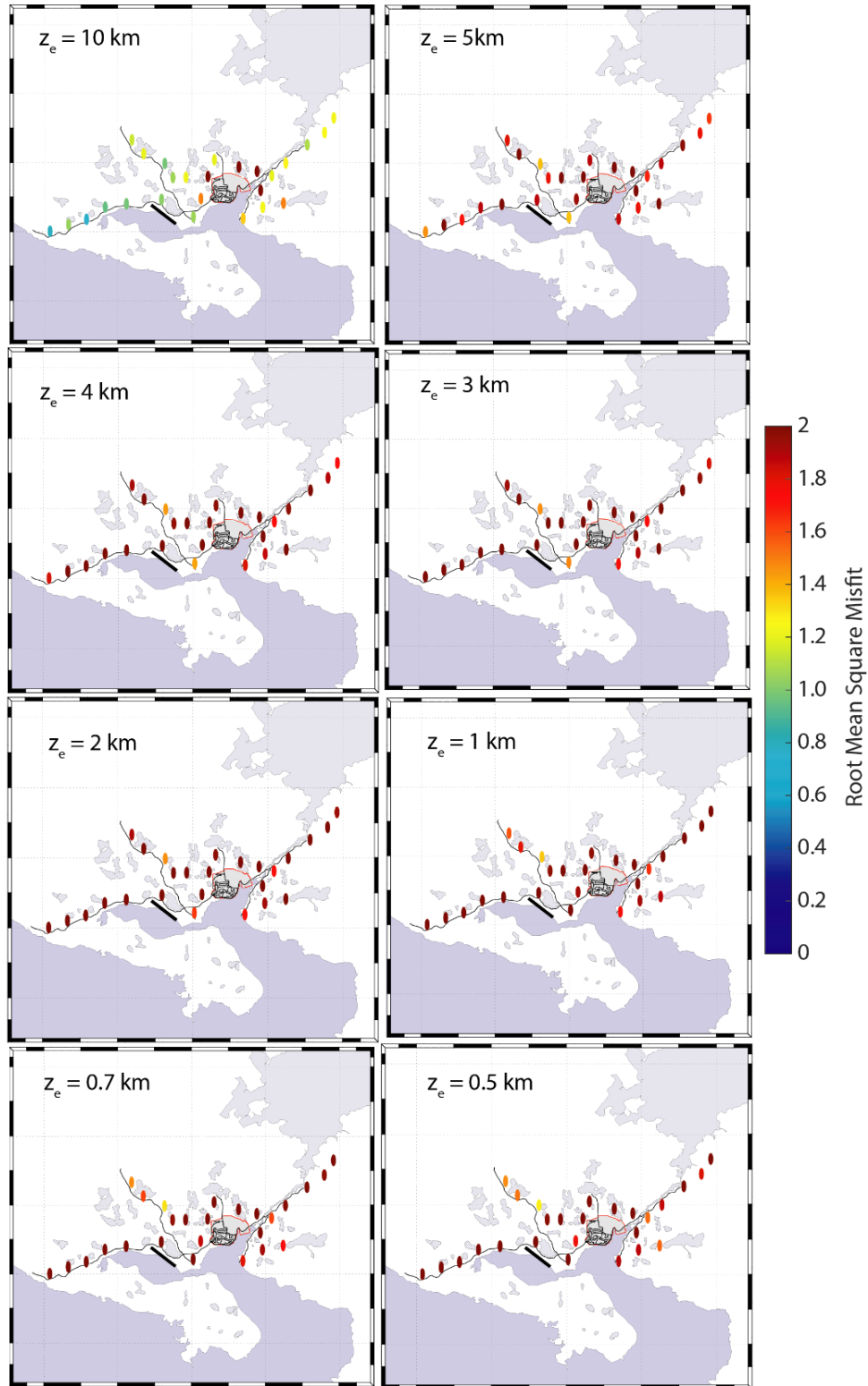


Figure A9: RMS misfit maps for edited models. The model *CB_inv13* was edited to a $1000 \Omega m$ half-space below at depths between 10 km and 0.5 km.

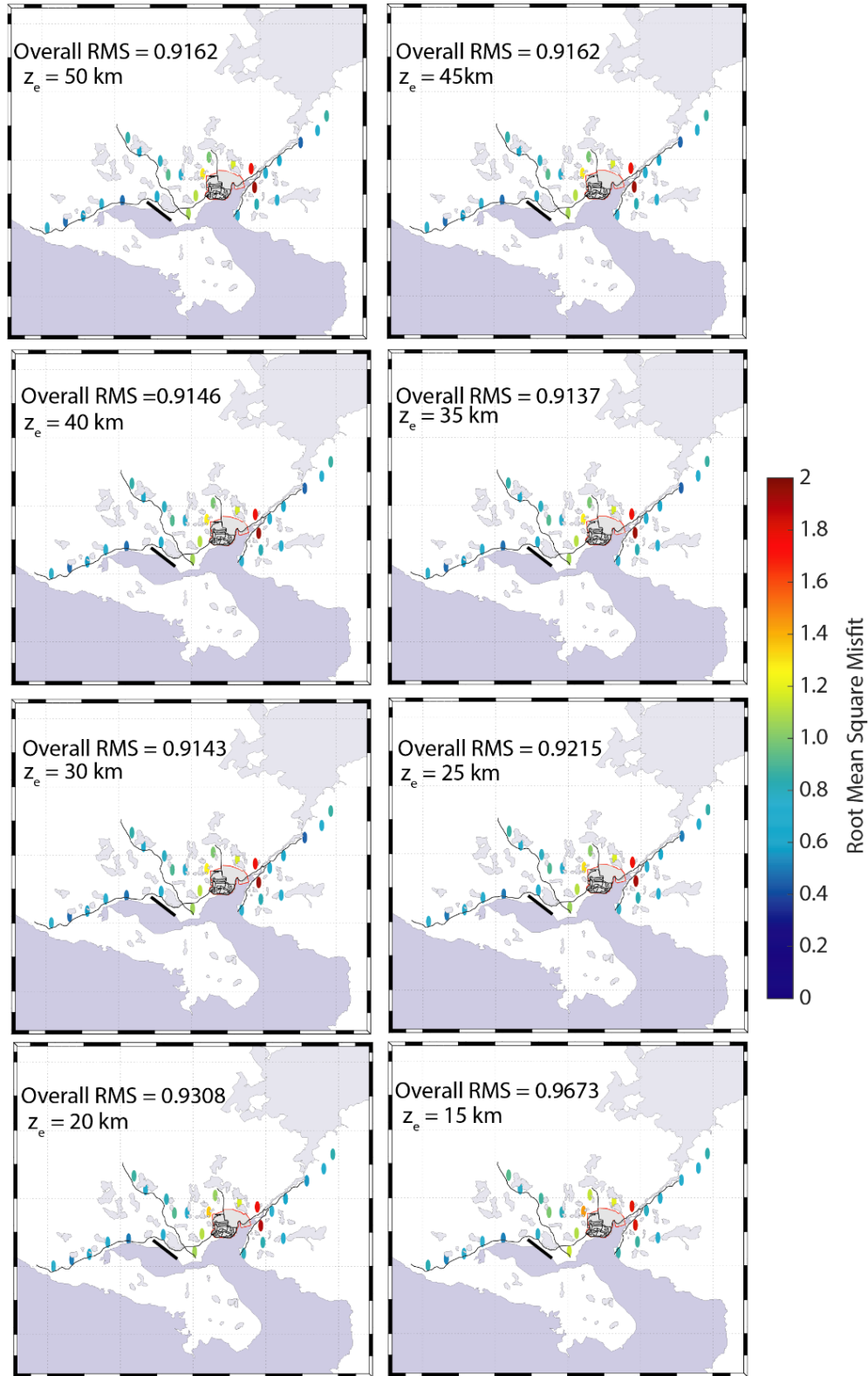


Figure A10: RMS misfit maps for edited models. The model *CB_inv13* was edited to a $10 \Omega m$ half-space below at depths between 50 km and 15 km.

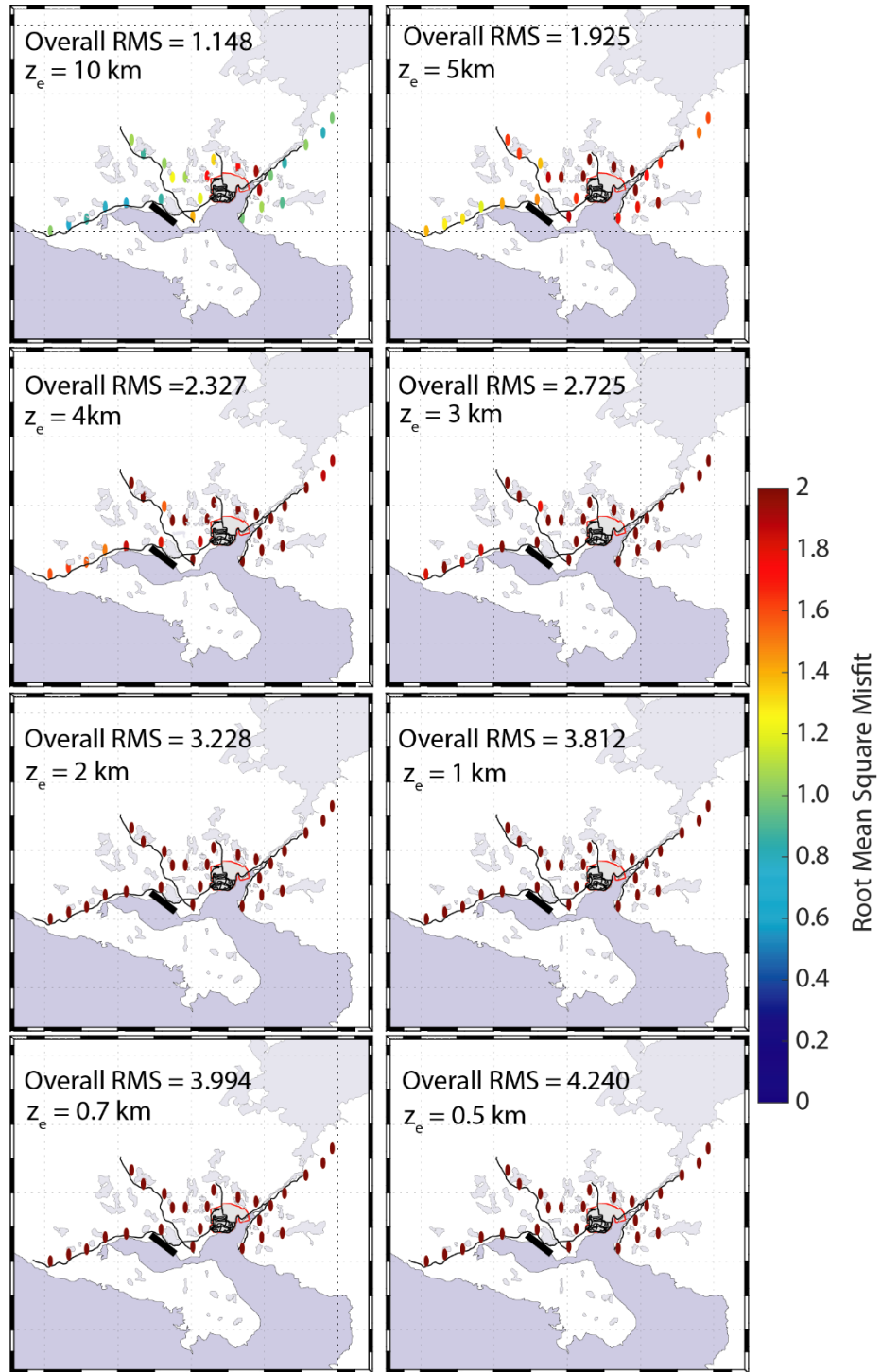


Figure A11: RMS misfit maps for edited models. The model *CB_inv13* was edited to a 10 Ωm half-space below at depths between 10 km and 0.5 km.

B. Supplementary material for Chapter 6

Station	Data run	N-S (m)	E-W (m)	Latitude	Longitude	Access	Configuration	Runs
RES001	10351_2022-06-06-135806, 10351_2022-06-05-213552, 10351_2022-06-06-005308	96	96	74.701	-94.945	Truck	2E-2H	4
RES002	10379_2022-06-06-164939, 10379_2022-06-06-204512	96	96	74.738	-95.009	Truck	2E-2H	3
RES003	10351_2022-06-07-163950	96	96	74.696	-94.896	Truck	2E-2H	2
RES004	10379_2022-06-07-211006	96	96	74.739	-94.957	Truck	2E-2H	2
RES005	10351_2022-06-08-163624	96	96	74.676	-94.908	Truck	2E-2H	2
RES006	10379_2022-06-08-210618, 10379_2022-06-09-155105	96	96	74.729	-95	Truck	2E-2H	4
RES007	10351_2022-06-09-202655	96	96	74.697	-94.811	Truck	2E-2H	2
RES008	10379_2022-06-10-163813	96	96	74.688	-94.924	Truck	2E-2H	2
RES009	10351_2022-06-10-205930	96	96	74.693	-94.797	Truck	2E-2H	2
RES010	10379_2022-06-11-163219	96	96	74.705	-94.978	Truck	2E-2H	2
RES011	10351_2022-06-11-220350	96	96	74.744	-95.036	Truck	2E-2H	2

Table B1: A summary of the data collected at Resolute Bay in 2022. The location and configuration of each station is shown.

Station	Data File	N-S (m)	E-W (m)	Latitude	Longitude	Elevation (m)	Access	Configuration
RES101	10430_2023-08-04-162827	59	61	74.70091	-95.0581	26.832366	Truck	2E-2H
RES102	10352_2023-08-04-200204	49	-64	74.70615	-95.0517	20.440709	Truck	2E-2H
RES103	10352_2023-08-05-163720	51	-65	74.70797	-95.0363	27.547182	Truck	2E-2H
RES104	10352_2023-08-05-203714	50	65	74.7109	-95.0186	32.20712	Truck	2E-2H
RES105	10352_2023-08-06-162301	48	50	74.71401	-95.0046	36.532628	Truck	2E-2H
RES107	10352_2023-08-06-203706	50	48	74.69198	-94.9841	19.286831	Truck	2E-2H
RES106	10379_2023-08-07-142334	-64	-59	74.714	-94.9911	36.576432	Truck	2E
RES108	10352_2023-08-07-162355	51	48	74.69597	-94.9758	29.276943	Truck	2E-2H
RES109	10379_2023-08-07-192145	-63	-63	74.69941	-94.9673	29.207302	Truck	2E
RES110	10352_2023-08-07-210058	46	-50	74.68758	-94.9911	28.206472	Truck	2E-2H
RES111	10379_2023-08-08-152619	65	63	74.68351	-94.9996	39.444381	Truck	2E
RES112	10352_2023-08-08-165851	48	49	74.68174	-94.9809	25.692781	Truck	2E-2H
RES113	10379_2023-08-08-195059	65	65	74.67983	-94.9649	32.340011	Truck	2E
RES114	10352_2023-08-08-212535	48	51	74.67912	-94.9466	8.829209	Truck	2E-2H
RES115	10379_2023-08-09-164500	-57	64	74.67749	-94.929	10.370828	Truck	2E
RES116	10352_2023-08-09-195347	-51	-47	74.677	-94.9119	13.282481	Truck	2E-2H
RES117	10379_2023-08-09-205545	64	-54	74.68542	-94.9151	3.213184	Truck	2E
RES118	10352_2023-08-10-163854	51	-48	74.70169	-94.8977	30.763603	Truck	2E-2H
RES119	10379_2023-08-10-191104	63	64	74.69924	-94.9231	16.302515	Truck	2E
RES120	10352_2023-08-10-204115	48	50	74.70922	-94.9049	28.599396	Truck	2E-2H
RES121	10379_2023-08-11-151346	64	62	74.70515	-94.9453	44.147682	Truck	2E
RES122	10352_2023-08-11-162846	49	48	74.71303	-94.9269	63.980317	Truck	2E-2H

RES123	10379_2023-08-11-195512	62	-64	74.70997	-94.9349	58.394429	Truck	2E
RES124	10352_2023-08-11-214203	47	48	74.71638	-94.9407	60.045081	Truck	2E-2H
RES125	10352_2023-08-12-162815	63	62	74.69795	-95.027	35.178554	Truck	2E-2H
RES126	10379_2023-08-12-185800	47	44	74.70137	-94.9993	25.387291	Truck	2E
RES127	10352_2023-08-12-202125	-65	-60	74.70841	-94.9881	34.161358	Truck	2E
RES128	10352_2023-08-13-160320	59	-62	74.69242	-94.964	10.838917	Truck	2E-2H
RES129	10379_2023-08-13-164718	43	46	74.69215	-94.8783	4.030648	Truck	2E
RES130	10352_2023-08-13-195517	59	48	74.68927	-94.8225	-0.859588	Truck	2E-2H
RES131	10379_2023-08-13-205949	47	-46	74.68534	-94.782	32.047575	Truck	2E
RES132	10352_2023-08-14-153904	59	48	74.67154	-94.7718	6.538584	Truck	2E-2H
RES133	10379_2023-08-14-164822	-48	47	74.69963	-94.8602	47.434909	Truck	2E

Table B2: A summary of the data collected at Resolute Bay in 2023. The location and configuration of each station is shown.

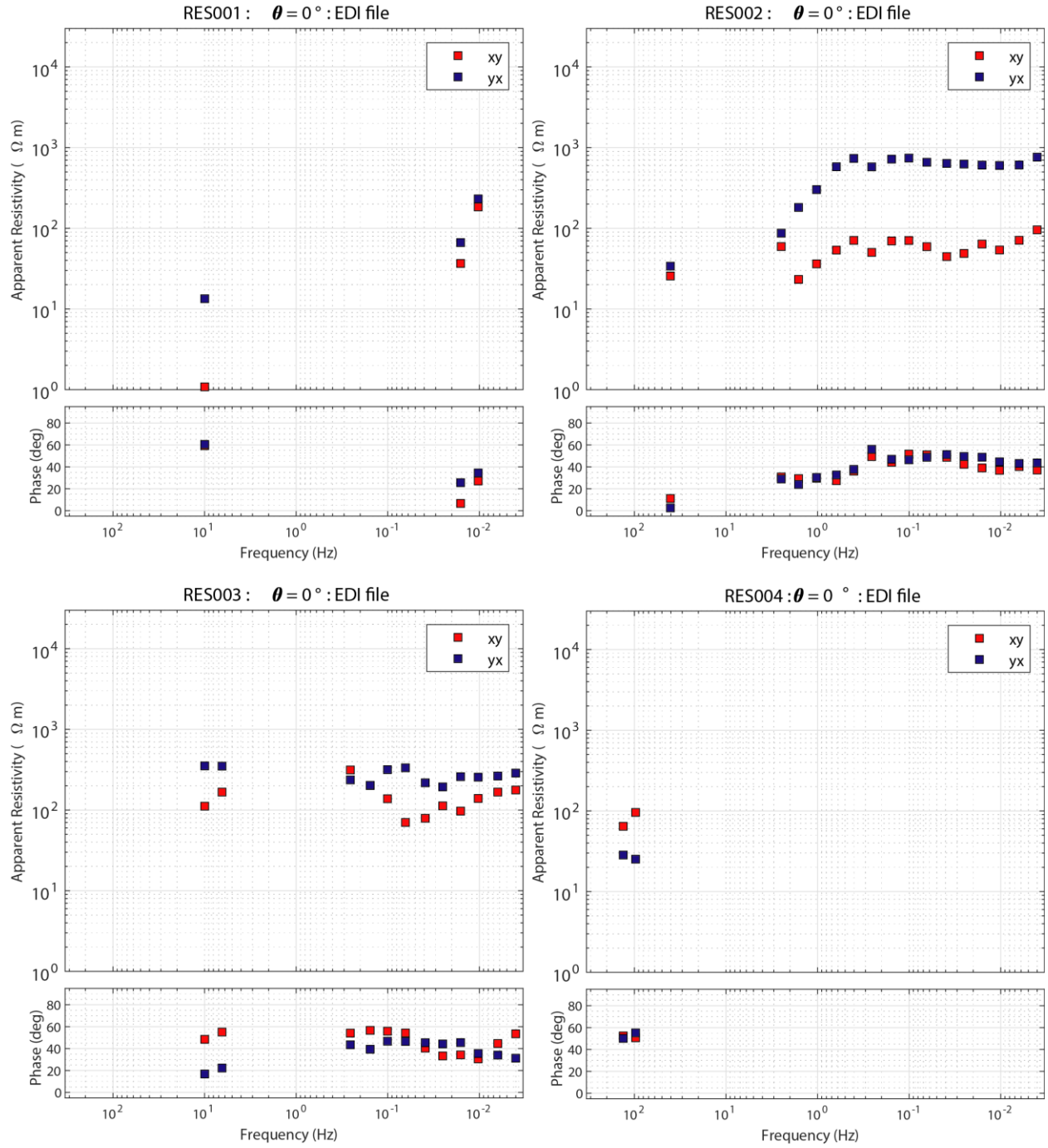


Figure B1: MT data at stations RES001, RES002, RES003, and RES004.

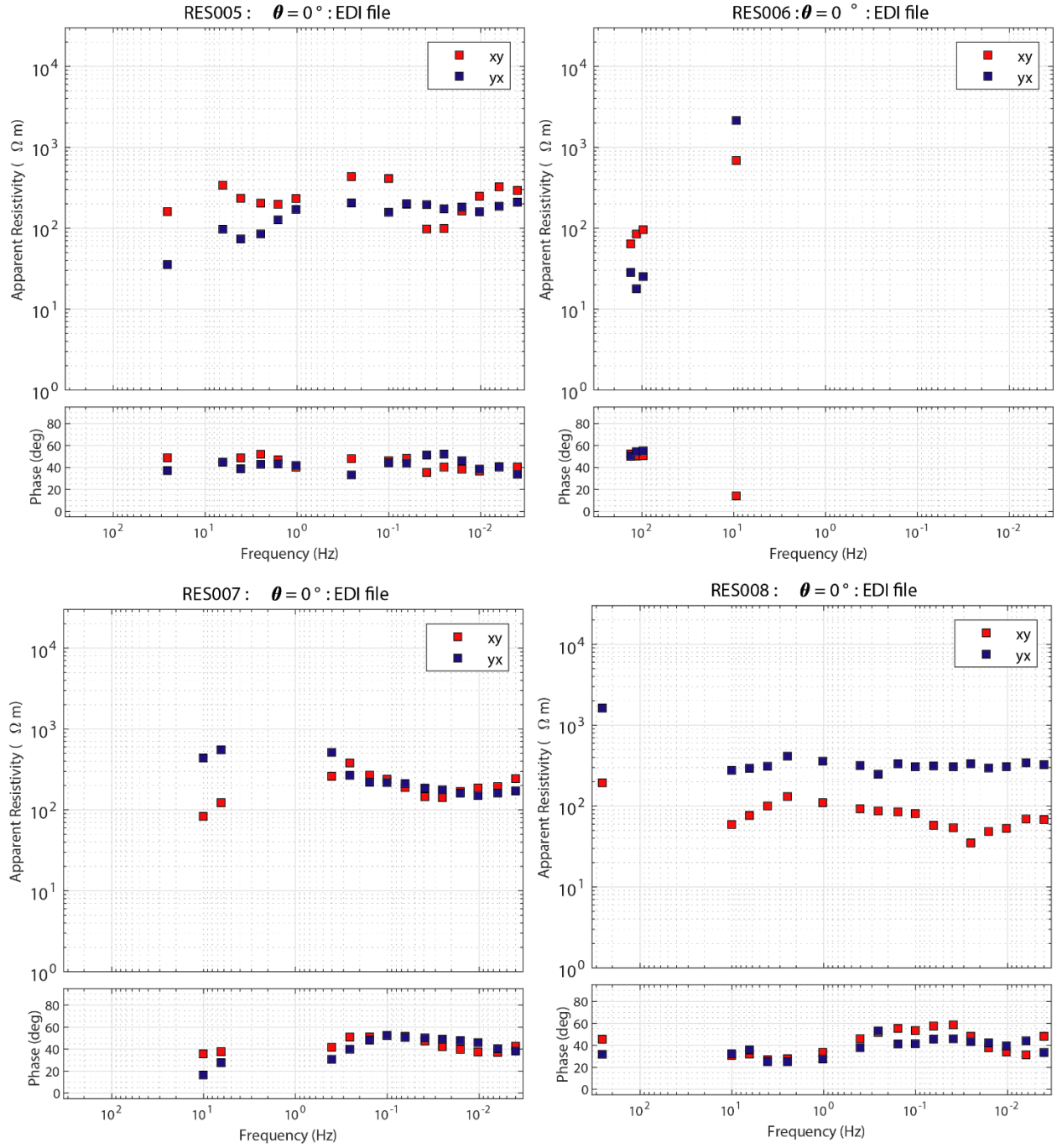


Figure B2: MT data at stations RES005, RES006, RES007, and RES008.

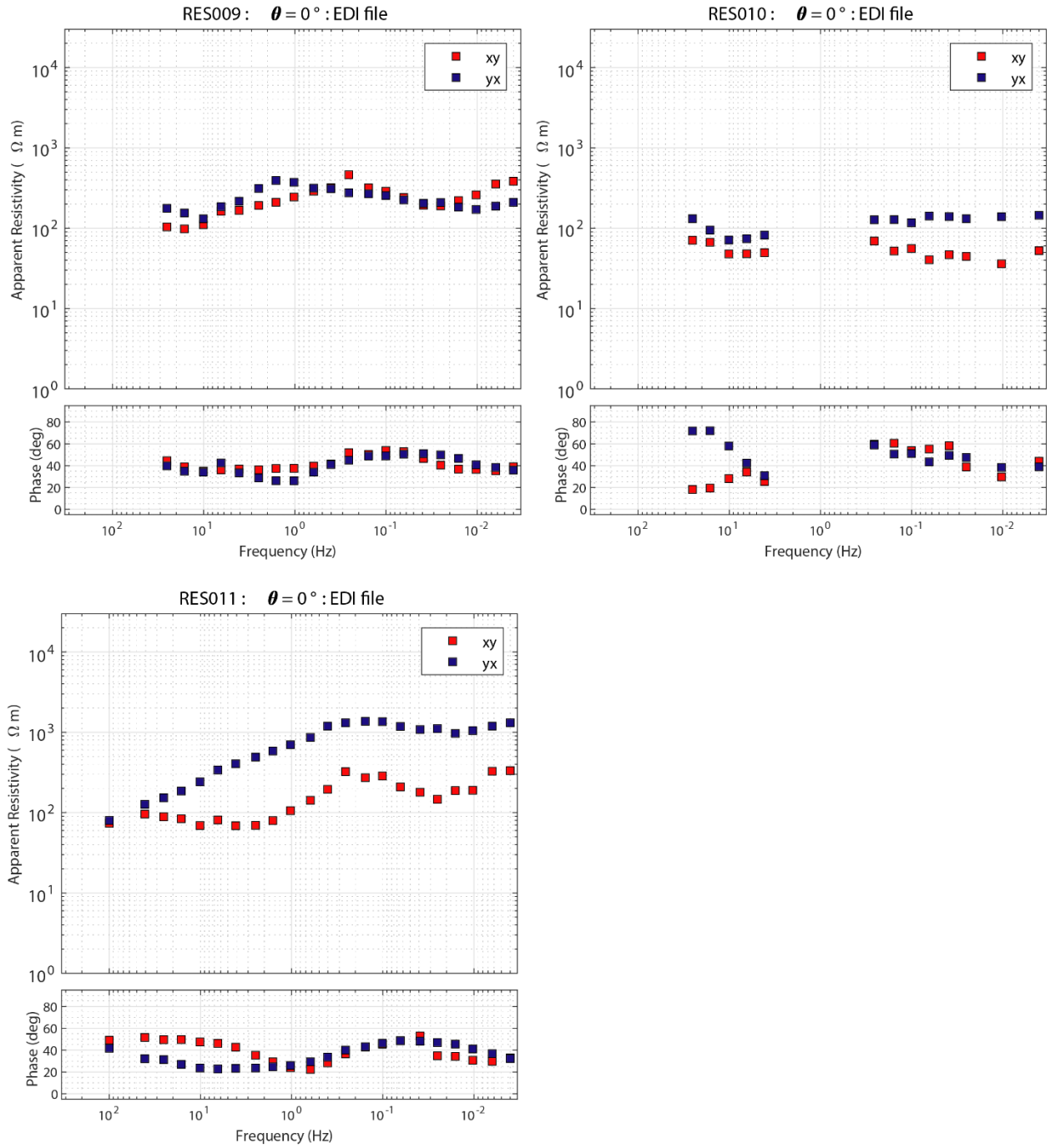


Figure B3: MT data at stations RES009, RES010, and RES011.

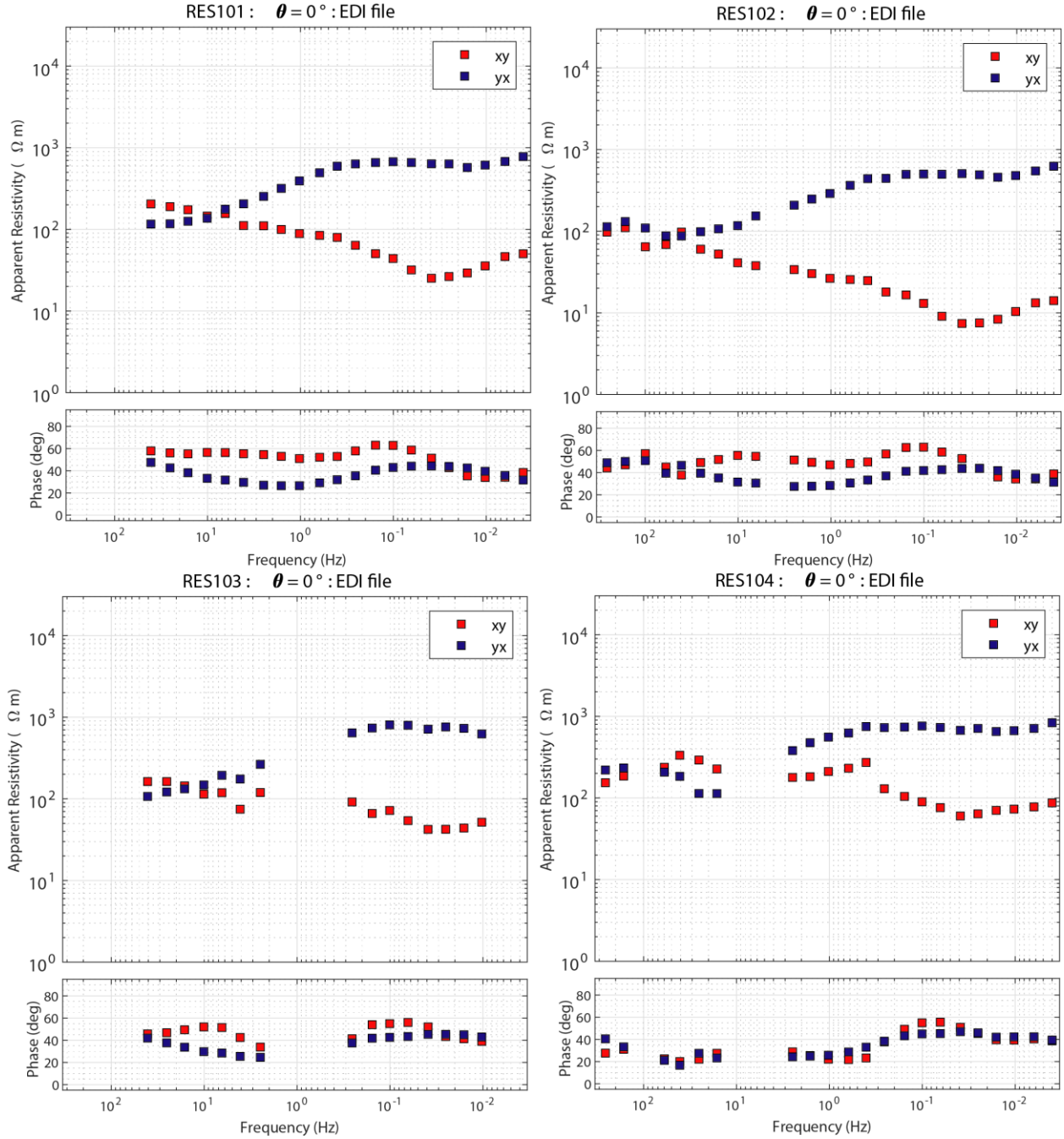


Figure B4: MT data at stations RES101, RES102, RES103, and RES104.

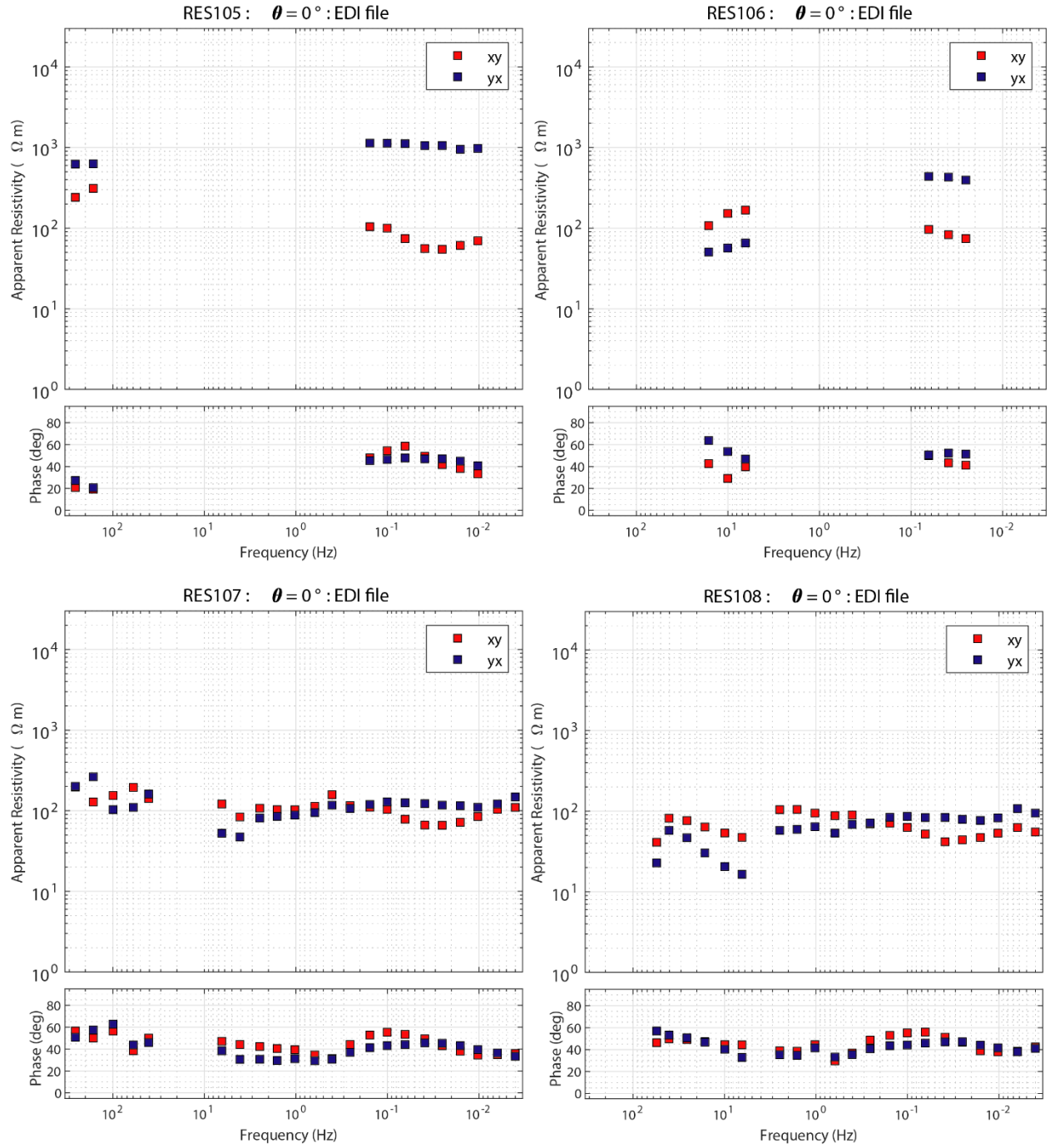


Figure B5: MT data at stations RES105, RES106, RES107, and RES108.

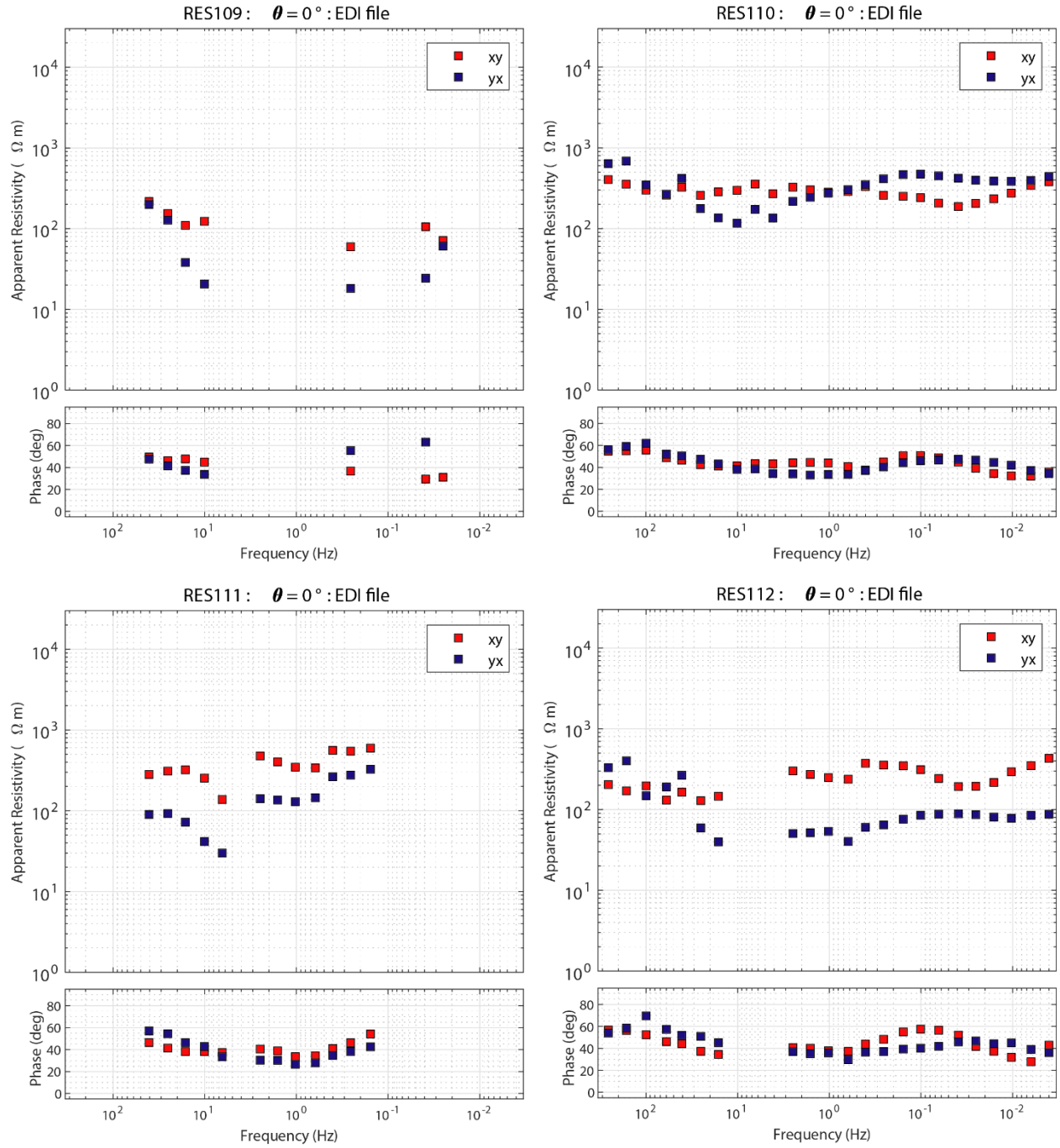


Figure B6: MT data at stations RES109, RES110, RES111, and RES112.

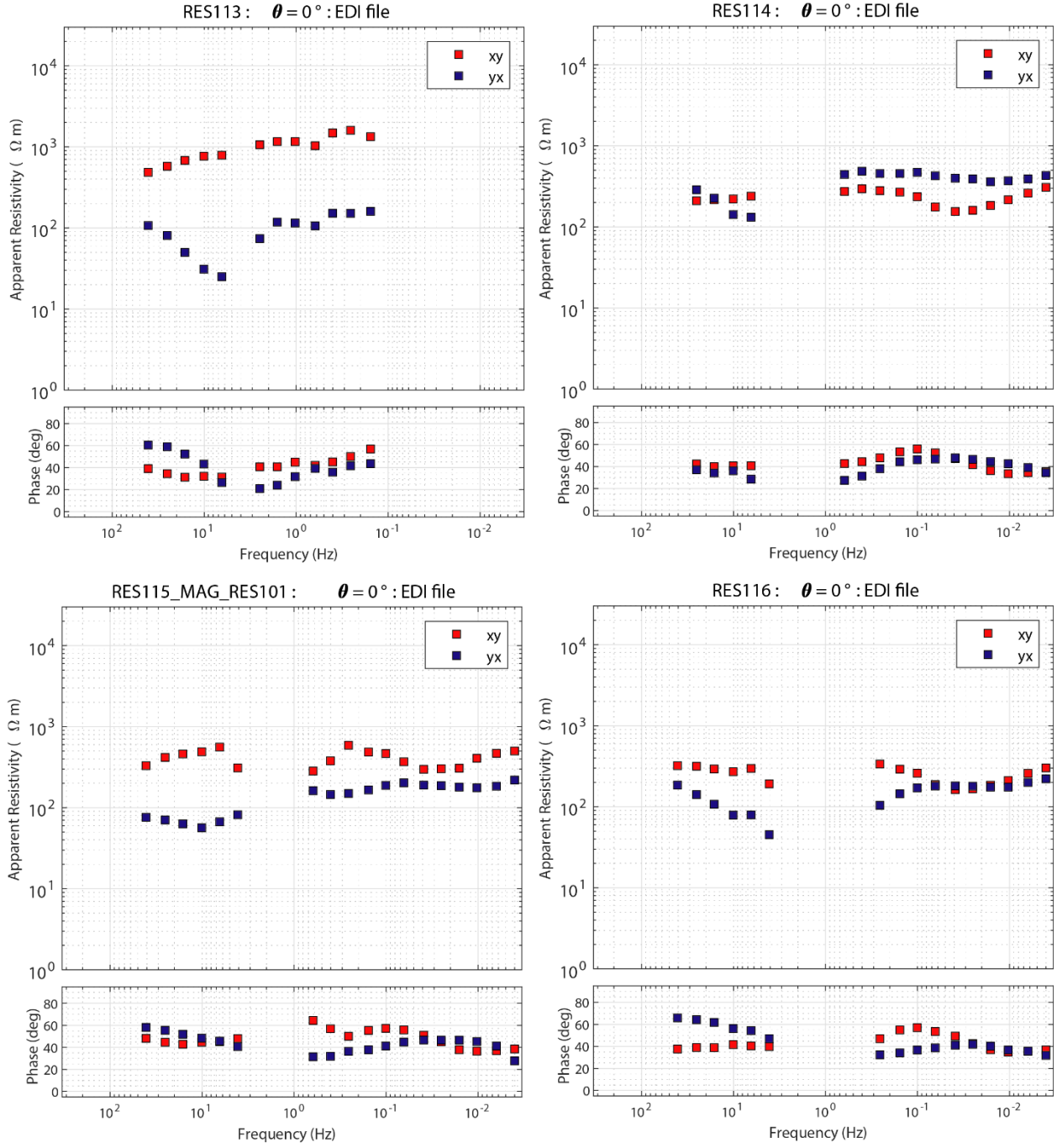


Figure B7: MT data at stations RES113, RES114, RES115, and RES116.

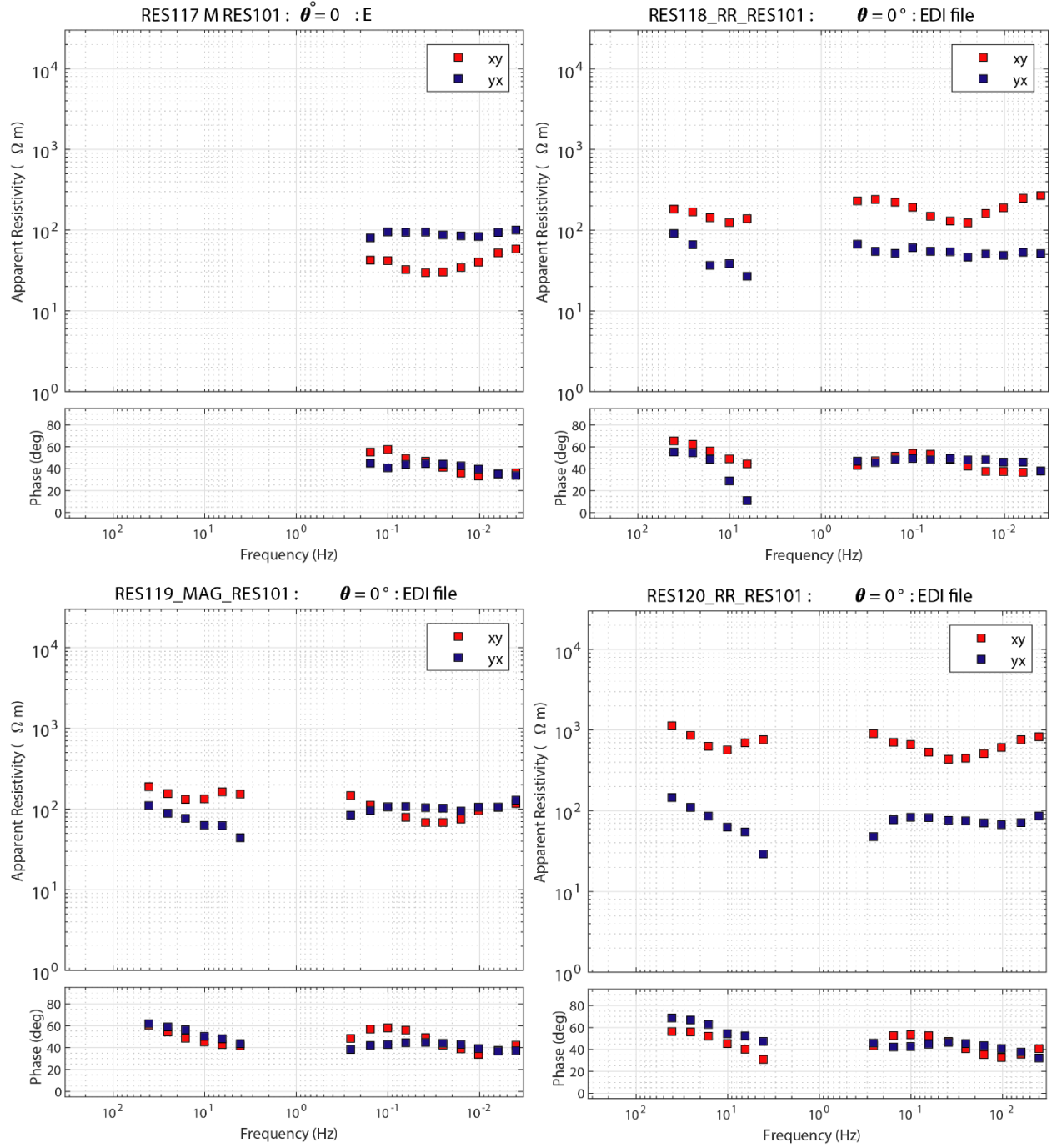


Figure B8: MT data at stations RES117, RES118, RES119, and RES120.

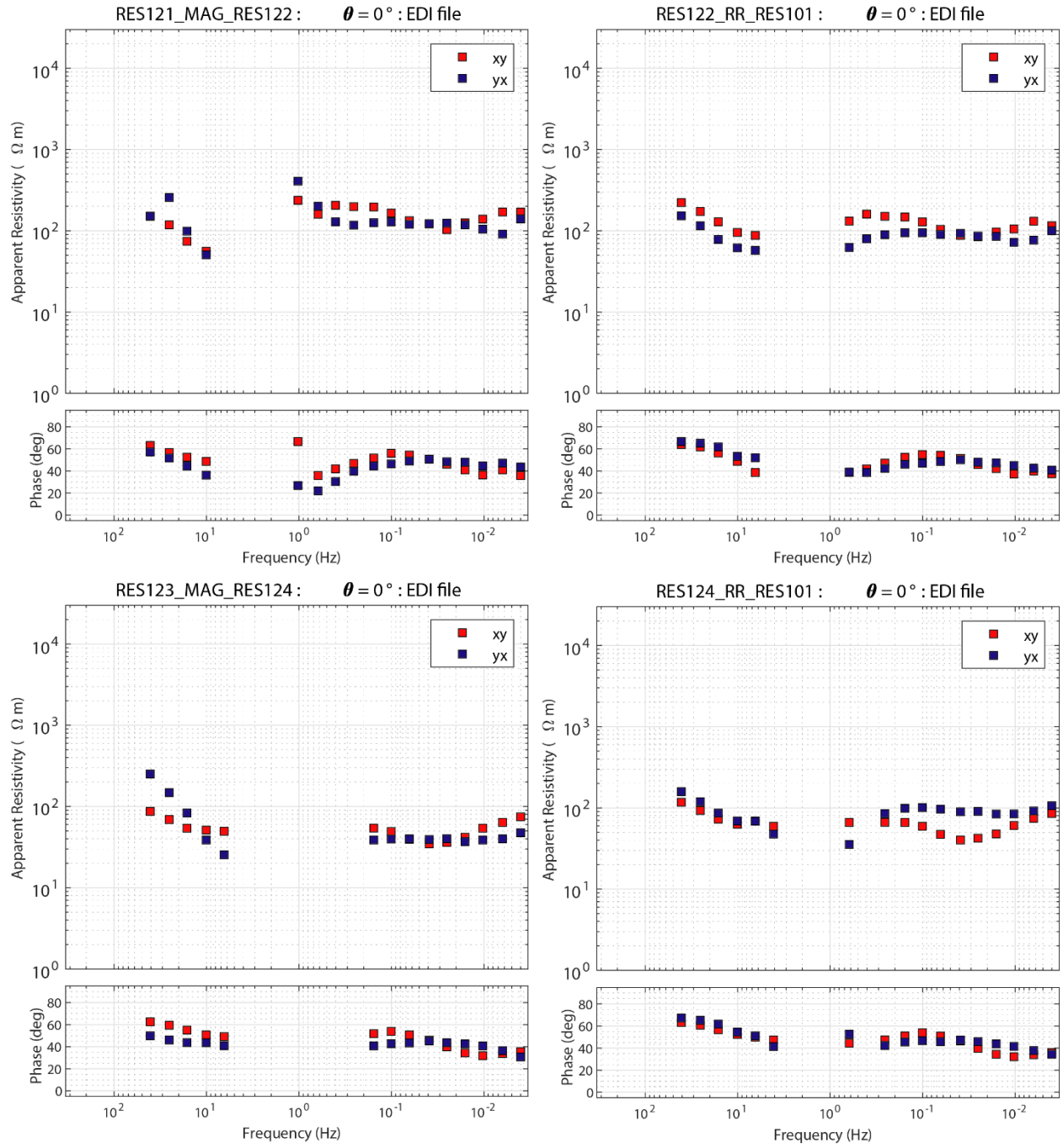


Figure B9: MT data at stations RES121, RES122, RES123, and RES124.

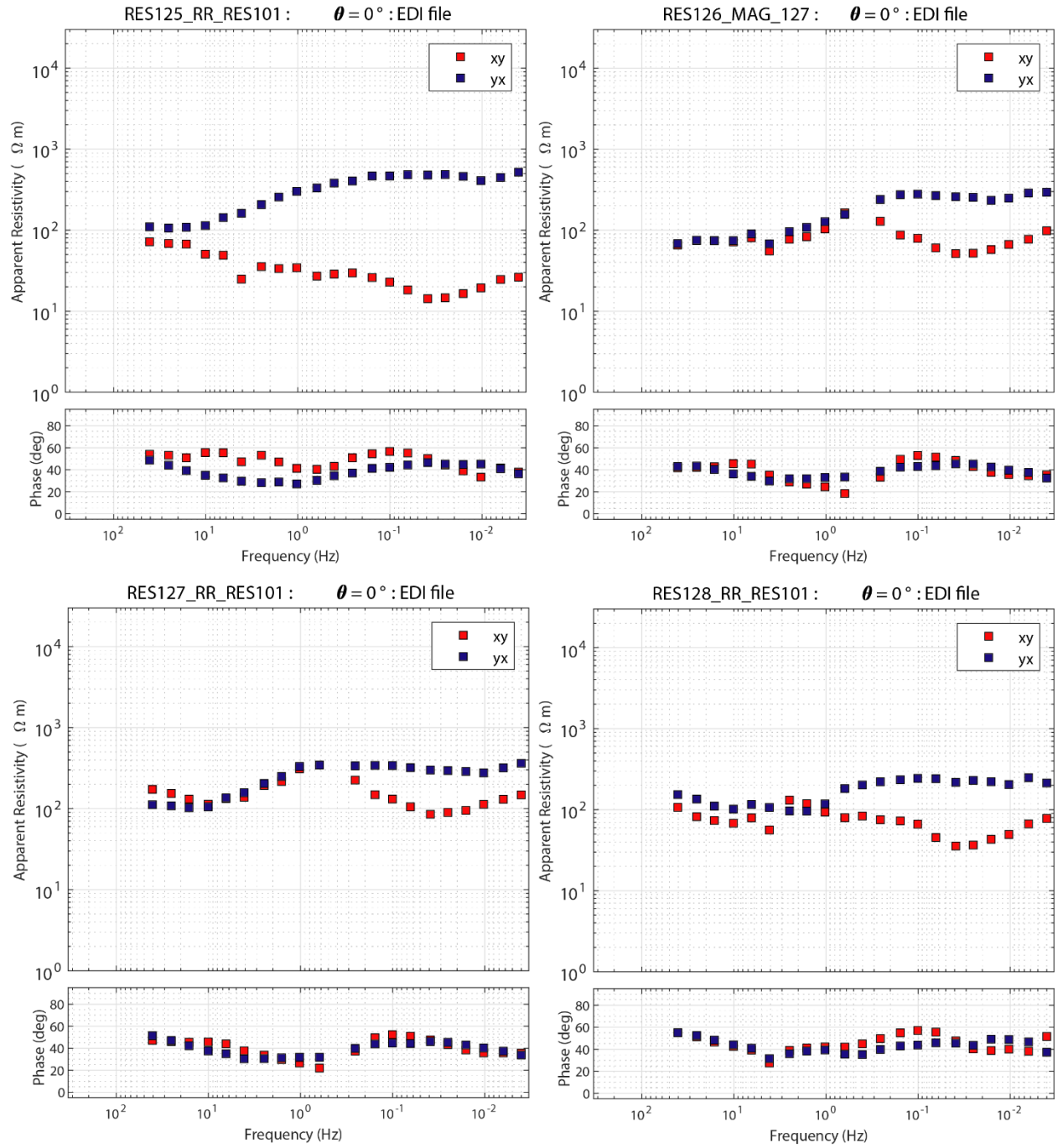


Figure B10: MT data at stations RES125, RES126, RES127, and RES128.

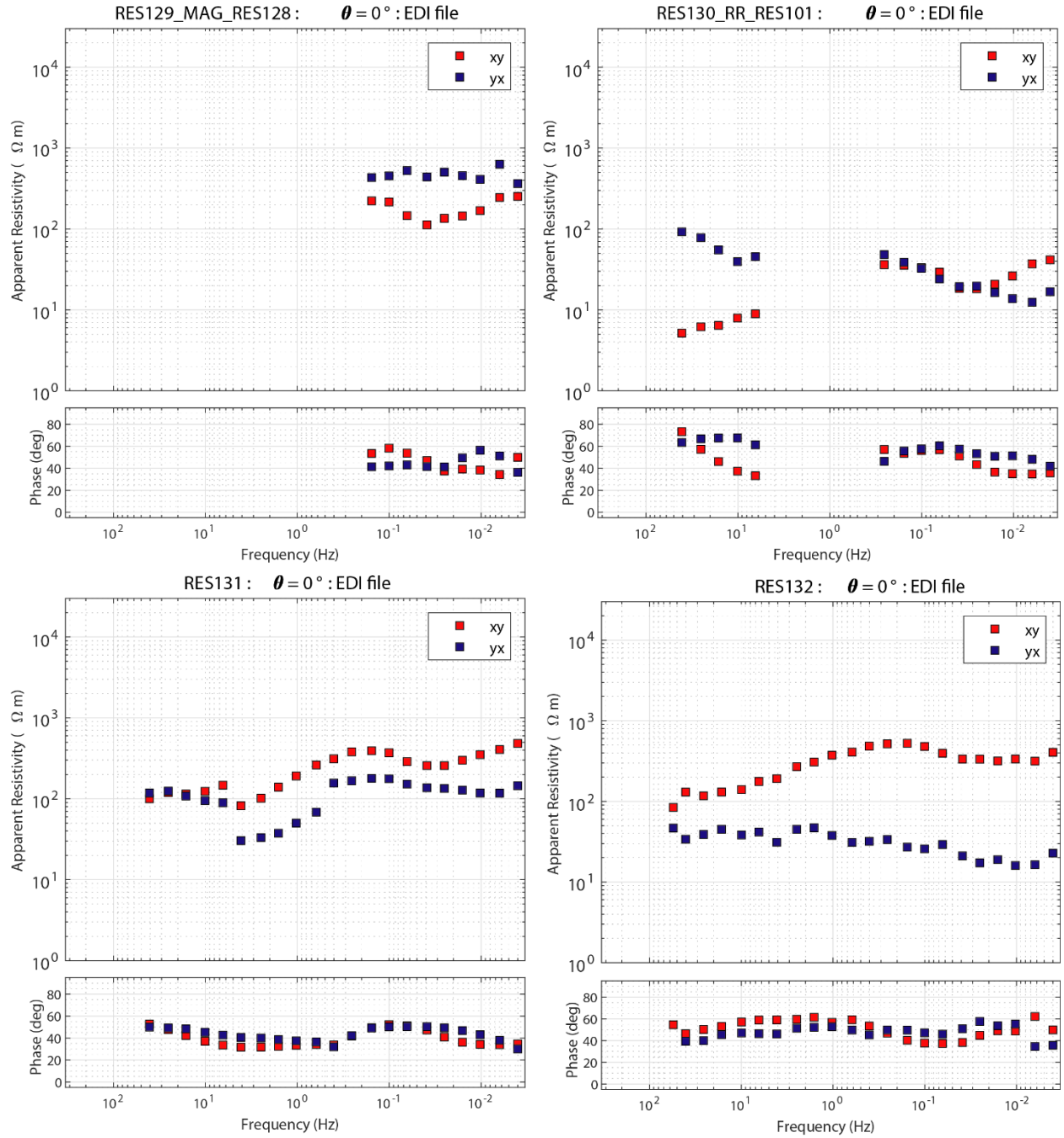


Figure B11: MT data at stations RES129, RES130, RES131, and RES132.

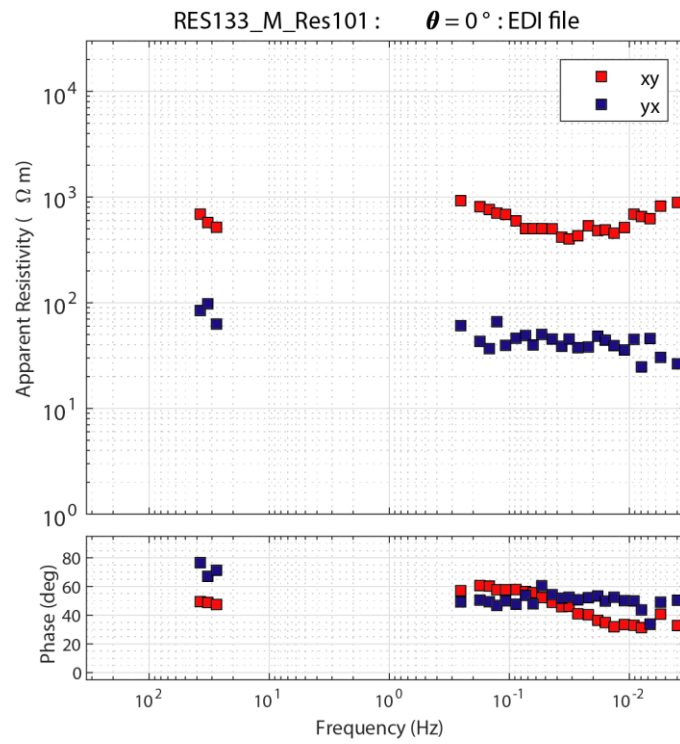


Figure B12: MT data at station RES131.

Station	Error Floor %	# layers	# iterations	Final RMS Misfit
RES001	5	150	7	0.99596
RES007	5	150	15	1.0206
RES008	10	150	200	1.8817
RES010	5	150	6	0.99946
RES011	5	150	7	0.9974
RES101	5	150	6	0.996
RES102	5	150	6	0.99849
RES103	5	150	5	0.99928
RES104	10	150	7	1.146
RES105	5	150	7	0.99972
RES106	5	150	7	0.99884
RES107	10	150	5	0.99974
RES108	5	150	9	1.2718
RES110	5	150	6	0.99941
RES111	5	150	7	0.99744
RES112	10	150	4	0.99978
RES113	5	150	9	0.99923
RES114	5	150	5	0.9986
RES115	10	150	4	0.9995
RES116	5	150	5	0.99982
RES118	5	150	5	0.99916
RES119	5	150	5	0.9984
RES120	5	150	14	1.1534
RES121	10	150	7	1.1109
RES122	5	150	5	0.99929
RES123	10	150	4	0.99973
RES124	5	150	5	0.99893
RES125	5	150	4	0.99914
RES126	5	150	5	0.99841
RES127	10	150	5	0.99981
RES128	5	150	5	0.99993
RES129	5	150	8	1.0467
RES130	5	150	6	0.99875
RES131	10	150	5	0.99997

Table B3: Summary of 1-D inversions for Resolute Bay using the OCCAM inversion algorithm of Constable et al., (1987). Several stations could not be reasonably fit using the algorithm and were excluded from 1-D analysis.

Name	# Stations	NX	NY	NZ	Topo/ Bath	Start model	Regularization Parameter, λ	covariance (default 0.5)	Start RMS	End RMS	Iterations
RES_inv01	42	48	50	65	n	100 Ω m h.s.	1	0.5	3.707	1.38	80
RES_inv02	42	98	104	65	n	(500 m , 1000 Ω m); (1000 m, 100 Ω m); (1000 Ω m h.s.)	1	0.5	7.283	1.62	94
RES_inv03	42	98	104	65	n	1000 Ω m h.s.	1	0.5	10.78	1.71	103
RES_inv04	29	44	64	65	n	100 Ω m h.s.	1	0.5	3.701	1.51	80
RES_inv05	29	44	64	65	n	100 Ω m h.s.	1	x,y 0.6	3.701	1.50	77
RES_inv06	29	66	104	65	n	100 Ω m h.s.	1	0.5	3.701	1.53	78
RES_inv07	29	66	104	65	y	0.3 Ω m ocean, 100 Ω m h.s.	1	x,y 0.4	3.56	1.62	35
RES_inv08	29	66	104	65	y	0.3 Ω m ocean, 100 Ω m h.s.	1	x,y 0.6	3.52	1.48	72
RES_inv09	29	110	186	75	y	0.3 Ω m ocean, 100 Ω m h.s.	1	0.5	3.51	1.83	69
RES_inv10	29	66	104	75	y	0.3 Ω m ocean, 100 Ω m h.s.	1	0.5	3.56	1.51	73

Table B4: Summary of 3-D inversions RES_inv01 – RES_inv10 for Resolute Bay using the ModEM inversion algorithm of Kelbert et al., (2014).

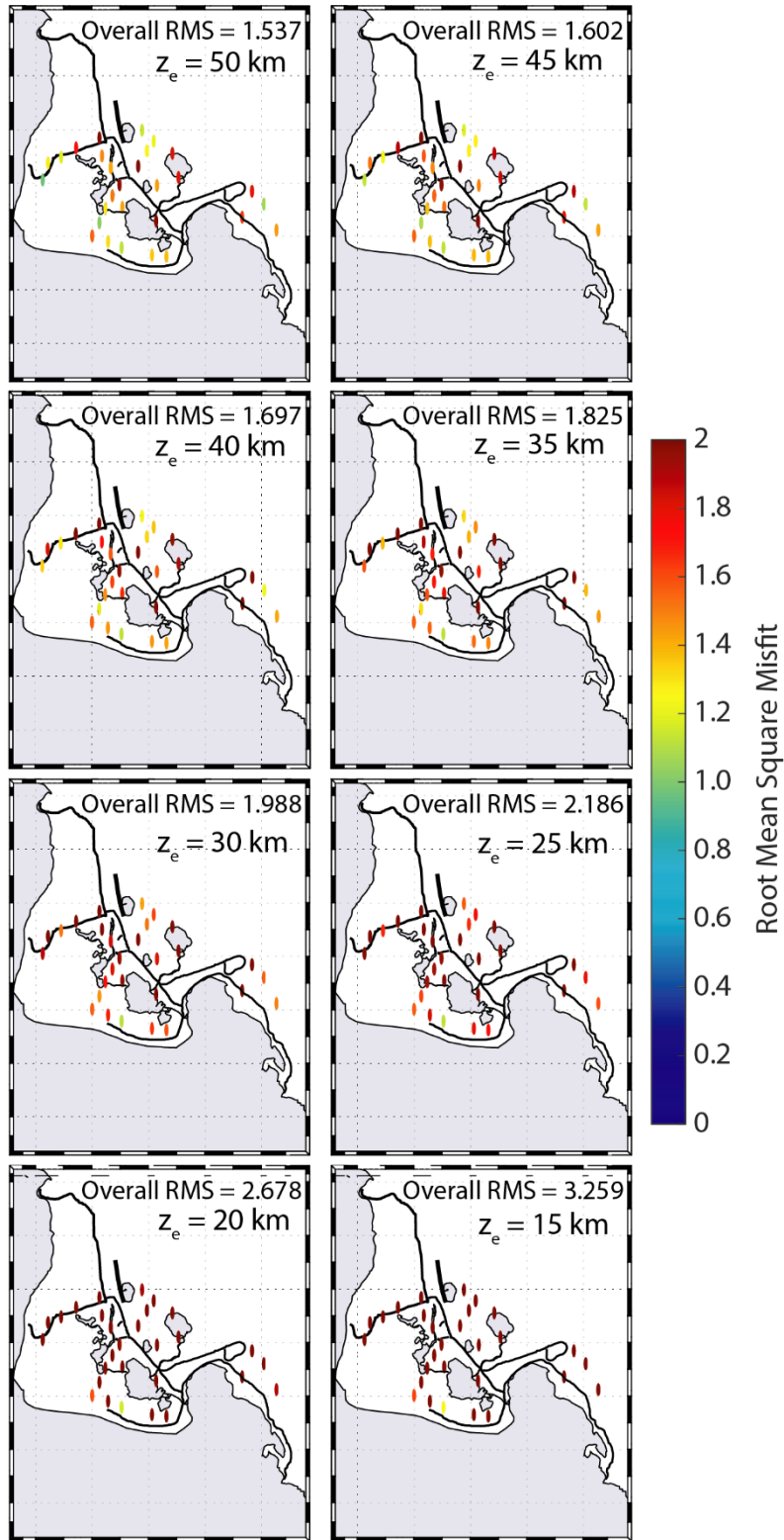


Figure B13: RMS misfit maps for edited models. The model *RES_inv08* was edited to a 1000 Ωm half-space at depths between 50 km and 15 km.

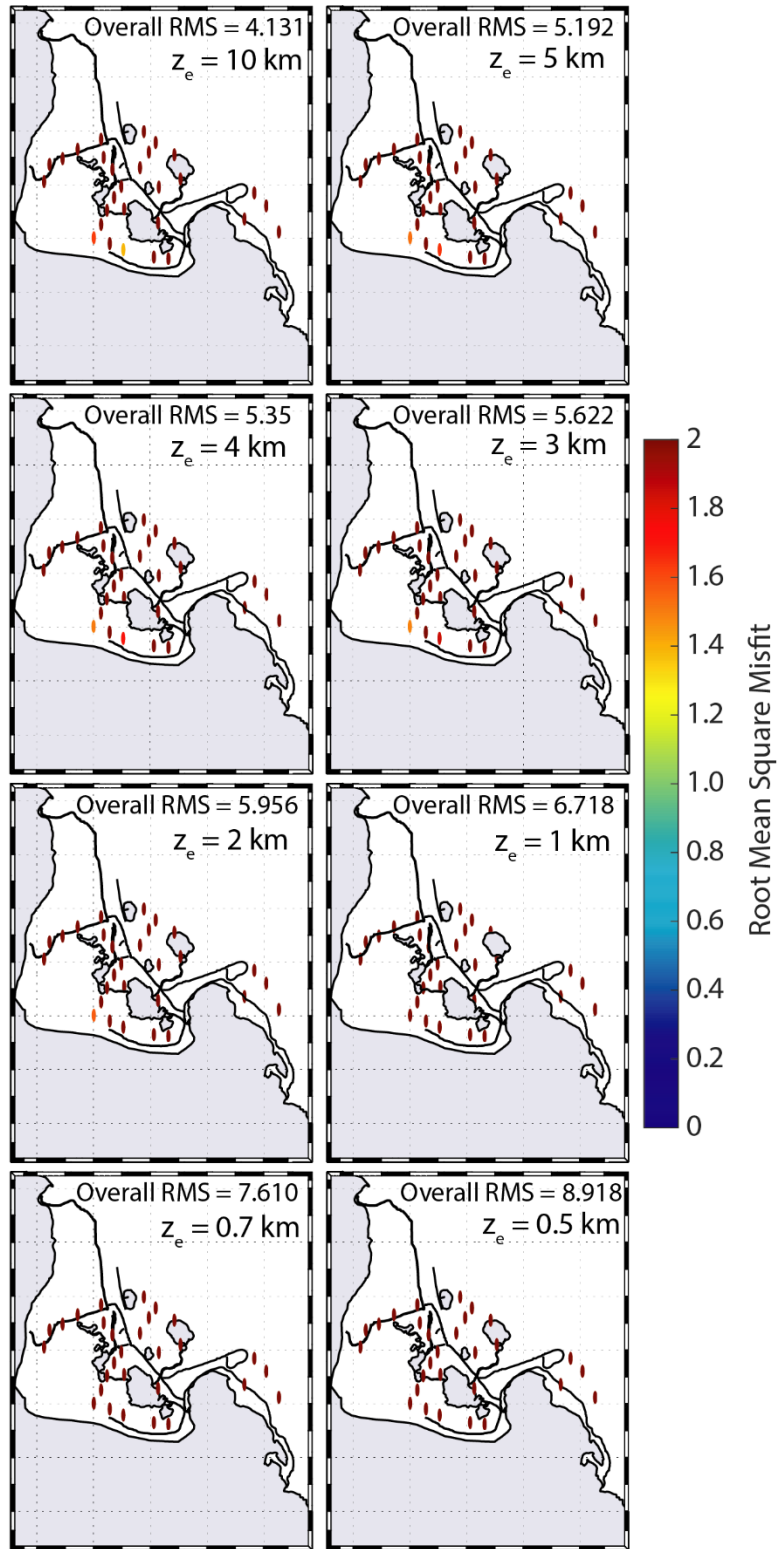


Figure B14: RMS misfit maps for edited models. The model *RES_inv08* was edited to a 1000 Ωm half-space at depths between 10 km and 0.5 km.

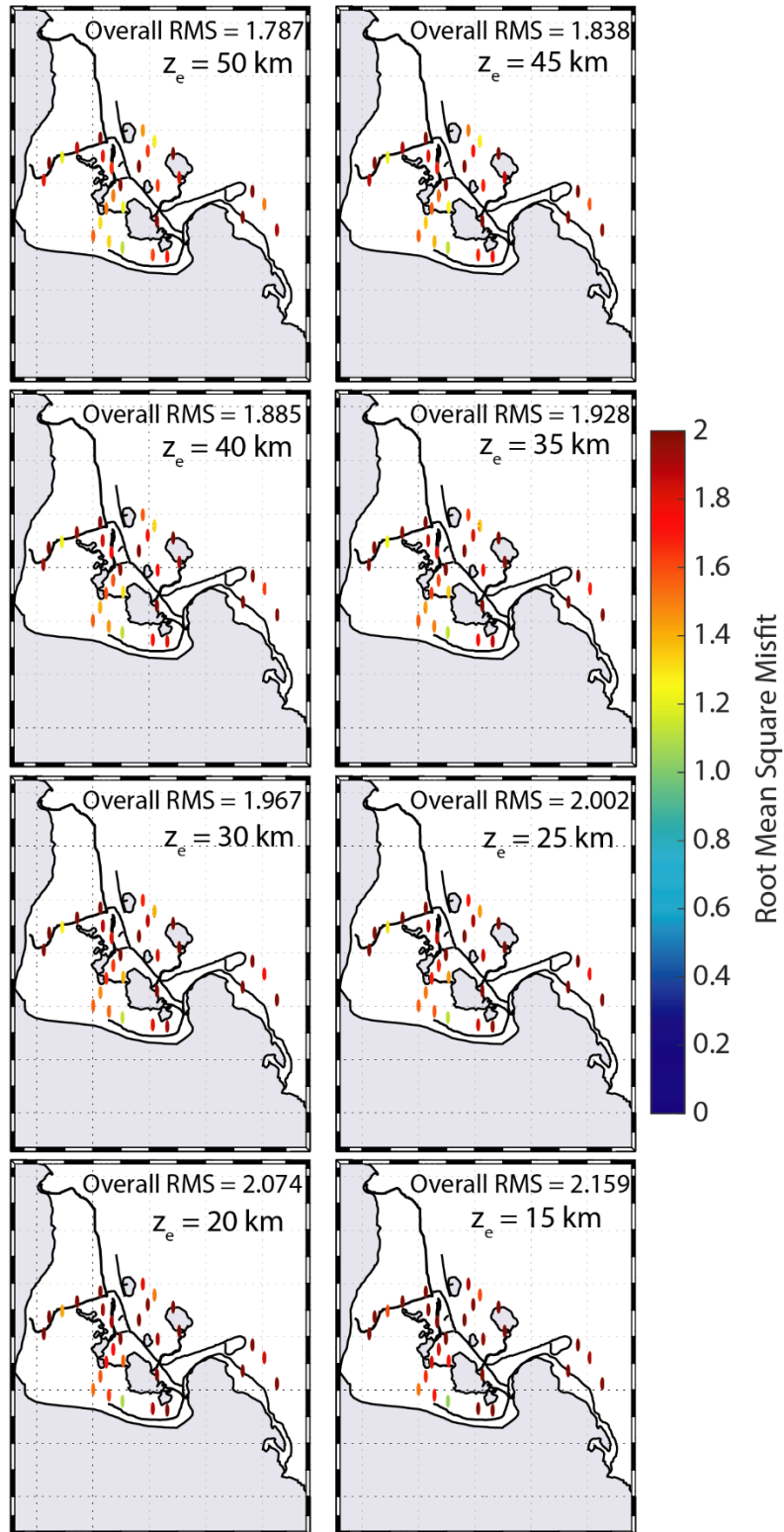


Figure B15: RMS misfit maps for edited models. The model *RES_inv08* was edited to a $10 \Omega m$ half-space at depths between 50 km and 15 km.

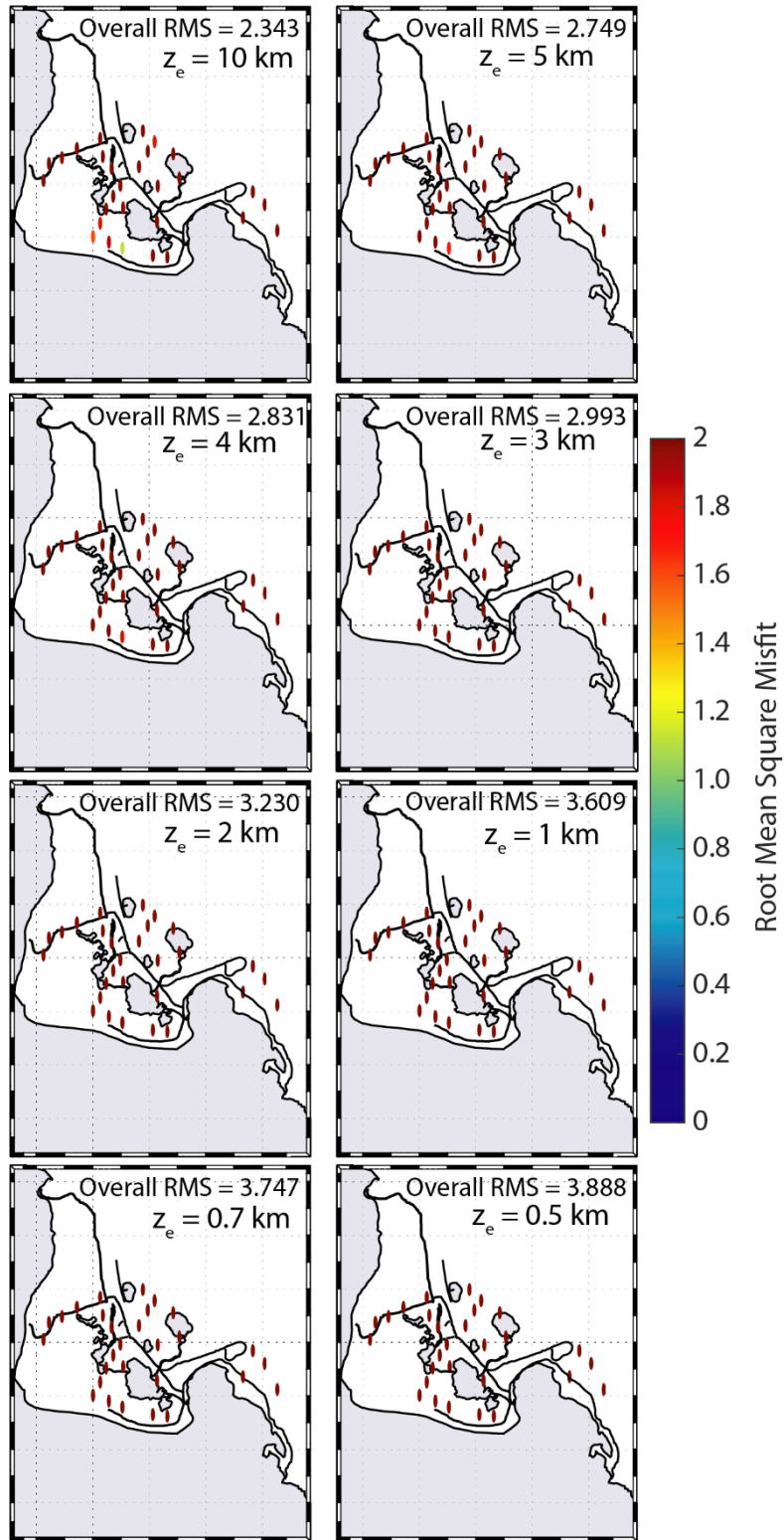


Figure B16: RMS misfit maps for edited models. The model *RES_inv08* was edited to a $10 \Omega m$ half-space at depths between 10 km and 0.5 km.