Mapping geothermal heat flux using permafrost thickness constrained by airborne electromagnetic surveys on the western coast of Ross Island, Antarctica

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ABSTRACT

Permafrost is ubiquitous at high latitudes, and its thickness is controlled by important local factors like geothermal flux, ground surface temperature and thermal properties of the subsurface. We use airborne transient electromagnetic resistivity measurements to determine permafrost thickness on the coast of Ross Island, Antarctica, which contains the active volcano Mt Erebus. Here, resistivity data clearly distinguish resistive permafrost from the electrically conductive fluidsaturated materials underlying it. For our study, we define permafrost as frozen material with a resistivity $> 100 \Omega \cdot m$; more conductive material contains a significant fraction of water or (more likely) brine. We observe that permafrost is very thin near the coast and thickens within several hundred metres inland to reach depths that are typically within the range of 300-400 m. We attribute the sharp near-shore increase in permafrost thickness to lateral heat conduction from the relatively warm ocean, possibly combined with seawater infiltration into the near-shore permafrost. We validate this result with a two-dimensional heat flow model and conclude that away from the thermal influence of the ocean, the local geothermal gradient and heat flux are about 45 ± 5 °C/km and 90 \pm 13 mW/m², respectively. These values are in line with published estimates in the vicinity of Mt Erebus and within the actively extending Terror Rift, but do not reflect a strong heat flow anomaly from volcanic activity of Mt Erebus. Measurements made previously in the McMurdo Dry Valleys, on the other side of McMurdo Sound, tend to be a few dozens of mW/m² lower, likely reflecting its different tectonic setting on the uplifted rift shoulder of Transantarctic Mountains. Our study demonstrates a new approach towards constraining geothermal flux in polar regions using airborne electromagnetic (AEM) data that can be relatively efficiently collected on regional scales where ice coverage does not exceed the penetration limits of the AEM device, which for the device used is ~ 500 m under the favourable conditions in the study area. ARTICLE HISTORY Received 1 October 2018 Accepted 30 July 2019

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Introduction

Geothermal heat flux (GHF) is a key constraint on the Earth's crustal and near-surface evolution and history (Davies and Davies 2010). It plays a particularly significant role in high polar environments where it helps to control the thickness of permafrost, distribution of groundwater, and motion of glaciers and ice sheets (Lachenbruch et al. 1982; McKenzie and Voss 2013; Fisher et al. 2015; Begeman, Tulaczyk, and Fisher 2017). Geothermal energy resources in high polar environments may play an increasing role in meeting future human energy needs (Ragnarsson 2003), including its increasing utilisation in support of polar research facilities (e.g. Kyle 1990), although significant barriers to practical implementation at high latitude remain.

Despite the scientific and practical importance of constraining GHF in polar regions, obtaining measurements and estimates of this quantity is challenging due to the technical difficulty and logistical barriers associated with drilling through permafrost and glacier ice in remote polar environments (e.g. Melles et al. 2011; Tulaczyk et al. 2014). Furthermore, it is increasingly recognised that polar environments need to be protected because they may contain uniquely adapted organisms and habitats subject to limited human disturbance (e.g. Doran and Vincent 2011). Such environmental protection measurements increase the difficulty of borehole drilling and temperature gradient measurements in sensitive polar regions (Tulaczyk et al. 2014).

Here, we present a geophysical approach towards mapping geothermal heat flux in a high polar environment using an airborne electromagnetic (AEM) survey of permafrost thickness at the western flank of the Mt Erebus volcano, Antarctica (Figure 1). This survey, which covered $\sim 20 \, \text{km}^2$, was part of a larger 2011 campaign in the McMurdo Dry Valleys region which collected

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Figure 1. Map of the field area. Ross Island is depicted in the main figure frame and the inset shows the location of survey flightlines in green. The northern flightlines cover the southern edge of Cape Royds and Cape Barne, and the southern flightlines cover Cape Evans. In between, a few short lines cover the terminus of the Barne Glacier. The red outline marks the extent of data displayed in Figure 2. Measurements outside the red outline were either over water (limited DOI) or high residual, low signal measurements that were too far from the bulk of the data to use in our calculations.

time-domain electromagnetic data over 300 km² in total. Time-domain electromagnetic methods are ideal for interrogating the first several hundred metres of the subsurface while being small and light enough for an airborne package. The first such survey to use them in Antarctica (2011) detected liquid brines beneath glaciers and permafrost (Mikucki et al. 2015; Foley et al. 2015) and various lakes (Dugan et al. 2015; Mikucki et al. 2015) in the Dry Valleys. A 2018 follow-up survey using an improved system found evidence of sub-permafrost submarine groundwater discharge (Foley et al. 2019) across a similar area.

Various other electrical techniques have been used in Antarctica across a variety of scales and regions. Direct current methods, which require physical and electrical contact with the surface, have been used to study the Antarctic ice sheets (e.g. Hochstein 1967), ice shelves (e.g. Bentley 1977) and on frozen ground in the Dry Valleys (McGinnis et al. 1973). Frequency-domain methods, an electromagnetic technique closely related to time-domain electromagnetics, have been used before in Antarctica on both ground-based and airborne platforms, including to detect brine beneath permafrost (Ruotoistenmäki and Lehtimäki 1997). Frequency-domain electromagnetics is often used to measure Antarctic sea ice thickness (e.g. Worby et al. 1999), including on helicopter-borne packages (Rack, Haas, and Langhorne 2013). In the Arctic, helicopter-borne systems are commonly used to measure sea ice with great accuracy (Pfaffling, Haas, and Reid 2007). Northern hemisphere permafrost is also measured with a variety of electromagnetic techniques (Todd and Dallimore 1998; Hauck et al. 2001; Korhonen et al. 2009). Magnetotellurics, with their ability to image to great depth, have been used to investigate the magmatic system of Mt Erebus (Hill et al. 2017), the state of the bottom of the ice sheet through the lithosphere at the South Pole (Wannamaker et al. 2004) and the Transantarctic Mountains (Wannamaker et al. 2017).

This survey, a subset of the 2011 AEM campaign, covered the western coastal zone of Ross Island stretching from Cape Evans in the south to near (but not into) the edges of the Antarctic Specially Protected Areas ("ASPA") on Cape Royds in the north, and reached from a few hundred metres to a few kilometres inland. Permafrost is readily detectable in the AEM data because it represents a high-resistivity frozen layer underlain by a lower resistivity, fluid-saturated medium. For clarity, we use the term "permafrost layer" to refer to the upper part of the subsurface in which solid ice exists in some or all interstitial space. By assuming that the permafrost is in steady-state and estimating its surface and bottom temperatures, we calculate the conductive temperature gradient through the permafrost layer and estimate GHF. Our GHF values fall around 100 mW/m² which is close to the middle of the GHF range reported previously for this region (Morin et al. 2010) and is consistent with our study area being on the flank of an active volcano (Parmelee et al. 2015) where surface volcanic rocks date to 70,000–90,000 years ago (Harpel et al. 2004).

Methods

AEM resistivity

In late 2011 we measured subsurface electrical resistivity across the McMurdo Sound region (Mikucki et al. 2015) using a helicopter-borne time-domain electromagnetic (TEM) sensor, commercially known as SkyTEM (Sørensen and Auken 2004). The SkyTEM system has a rigid frame \sim 20 \times 30 m in dimension on which a transmitter coil, TEM instrumentation and receiver coil are mounted. The frame is suspended beneath a helicopter and measurements are conducted as the helicopter flies. The system used in this survey has a maximum transmitter moment of $\sim 160,000 \, \text{Am}^2$ and uses a dual moment transmitter that is capable of recording near-surface (tens of metres) resistivity with the low moment, and deep (hundreds of metres) resistivity with the high moment. Dual-moment acquisition balances high-resolution in the near-surface with large penetration depths; low- and high-moments in this case refer to signals collected following a current pulse of 9.5 and 95 A, respectively. The TEM signal is produced by pulsing a current in the transmitter coil, which is shut off as rapidly as possible. During the current turn-off, a timevarying magnetic field is induced in the subsurface. This time-varying magnetic field generates eddy currents in the subsurface, which in turn generate a secondary time-varying magnetic field that can be detected inductively using a receiver coil mounted on the suspended frame. The amplitude and time characteristics of the observed time-varying secondary field contain information about the electrical resistivity structure of the subsurface. Electrical properties of the Earth can be used to gain insight into subsurface geology; for example, permafrost regions have large resistivities, whereas regions containing unfrozen brines have very low resistivities. Readers are directed to chapter 6 of Kirsch (2009) for a more comprehensive discussion of the TEM method. An alternative AEM method that has been employed previously at high latitudes is frequency-domain electromagnetics (FEM) (Kovacs and Holladay 1990; Multala et al. 1996; Haas et al. 2009; Ball et al. 2011; Minsley et al. 2012). TEM and FEM differ in data acquisition protocol, where FEM measurements oscillate currents continuously and measure signals while currents are still present in the transmitter coil. Both schemes are able to image the electrical properties of the subsurface. The relative performance of FEM and TEM systems is system dependent; however, the SkyTEM system employed in this work is able to provide both shallow and deep information and has, in general, a much larger depth of investigation compared with the FEM system. The FEM system's penetration depths are limited by the lowest available frequency. Operation at low frequencies reduces signal-to-noise ratios and can increase instrumentation weight (Steuer, Siemon, and Auken 2009). Furthermore, at great depth, the secondary magnetic field can be a small fraction of the primary magnetic field. Because FEM systems transmit and receive simultaneously, distinguishing the two fields can be difficult when the secondary field is weak; TEM systems sidestep this problem by measuring the secondary field when the primary field is off (zero) (Zhdanov 2017).

TEM systems are particularly sensitive to conductive materials and can achieve maximum penetration (up to 400-600 m in our survey) when conductive material is below resistive material. In this scenario, the deep conductive material generates a strong secondary magnetic field, and signal is not lost to attenuation within the resistive overburden. The collected TEM data were inverted using a spatially constrained inversion scheme (Auken and Christiansen 2004; Viezzoli et al. 2008) in AarhusInv (Auken et al. 2015; Kirkegaard and Auken 2015), which creates a quasi-3D resistivity model of the subsurface using a forward model that is based on 1D physics. Despite the 1D forward model, the spatially constrained inversion scheme imposes similarity among neighbouring 1D models, allowing reproduction of 3D structures. In contrast to past results from the Dry Valleys that were collected using this SkyTEM system (Dugan et al. 2015; Foley et al. 2015; Mikucki et al. 2015), here we use a "sharp" inversion, which prioritises the production of models containing few layers of relatively homogenous intralayer properties (Vignoli et al. 2015). This results in sharp transitions between regions of contrasting resistivity, helping pinpoint horizons between neighbouring layers. This inversion scheme is appropriate for the subsurface structure in our study region, which consists of high resistivity materials (glaciers, permafrost, sea ice) on top of conductive materials (saturated rock or sediments, saltwater lakes, seawater), where the strong geological stratification should coincide with similarly strong contrasts in electrical properties.

The presented inversion results are masked (i.e. not shown) beneath the depth of investigation (DOI) (Vest Christiansen and Auken 2012). The DOI represents the depth to which the estimated resistivity profile has an appreciable impact on the data fit. Below the DOI, the estimated models cannot be discussed with great certainty as they are not well-constrained by the data. The DOI is reached within a few tens of metres in seawater or highly conductive substrate but is hundreds of metres deep when conductors lie beneath non-conductors, as in the case when permafrost or glaciers overlie saturated sediments. The reason for variable DOI is that the conductive seawater attenuates the TEM signal more rapidly (as a function of depth) than the non-conductive permafrost, thus allowing the system to image deeper when thick permafrost layers are present. More details on the system and processing for our Antarctic data are available in Foley et al (2015).

Electromagnetic methods are useful in the cryosphere because changes in the phase of water (from liquid to solid) can correspond to orders of magnitude changes in electrical resistivity. TEM is well-suited to the targeted system in this work (permafrost overlying a conductive brine), in that the depth to the conductor can be well-constrained and the resistive overburden allows the system to penetrate deeper. Note that in regions of very thick permafrost or ice it can be difficult to generate a measurable signal, because the eddy currents produced in the resistive ice correspond to a very weak secondary field. As such, if the permafrost layer exceeds a certain thickness one can lose the ability to detect the presence of an underlying conductor. To enhance depth penetration in these scenarios larger transmitter moments or lower internal noise receivers would be necessary.

Measuring permafrost thickness

Permafrost freezes from the top down which should be represented in electromagnetic-derived resistivities as a highly resistive (frozen) layer overlying a conductive (water-saturated) layer. The measured bulk resistivity of this conductive layer depends on the resistivity of the fluid and the porosity of the material that contains the fluid. This can be estimated with an empirical relationship called Archie's Law:

$$\phi = \left(\frac{\rho_0}{\rho_b}\right)^{-1/m} \tag{1}$$

where ϕ is the porosity, ρ_0 is the bulk resistivity, ρ_b is the resistivity of the fluid and *m* is a "formation factor" that represents the geometries of the pore network (Archie 1950). Assuming that the fluids we measure have a resistivity similar to seawater, and assuming a moderate value for the formation factor, measured resistivities of 100 and 10 Ω ·m correspond to porosities of 6% and 19%, respectively. Despite the uncertainties in determining the porosity of a saturated sediment from resistivity alone, 100 Ω ·m is a reasonable threshold to distinguish between permafrost and thawed material: it is too conductive to be permafrost or unweathered rock (Palacky 1987), and has been shown to correspond to material transitions elsewhere in the region (Foley et al. 2015).

To estimate permafrost thickness, we determined the depth from the surface to the 100 Ω ·m layer using Aarhus Workbench software. Note that the depth to the permafrost/brine interface is well-resolved because of the strong resistivity contrast between these layers (and the thickness of the brine layer); significant changes to the estimated interface depth result in poorer data fits. We exclude any measurements acquired over the ocean, or, in limited cases, where the 100 Ω ·m layer was not detected above the DOI.

Estimating geothermal flux

The simplest approach to estimating geothermal flux from permafrost thickness is to use (Osterkamp and Burn 2003, based on their equation 1):

$$G \simeq K \frac{\Delta T}{Z} \tag{2}$$

where *G* is the geothermal heat flux in SI units (W/m²), K is the thermal conductivity of permafrost (W/m C), and ΔT is the temperature difference between the mean surface temperature and the freezing point at the bottom of the permafrost layer. Clearly, this simple equation is underlain by a number of assumptions. First, this is a steady-state solution to a 1D heat diffusion equation, which implies that all of the quantities involved have not varied significantly over the time required to reach the steady state. A rough estimate of this equilibration timescale can be obtained by calculating the e-folding timescale, τ , for a 1D heat diffusion problem for a layer of thickness *Z*:

$$\tau = \frac{Z^2}{D} \tag{3}$$

where D is the thermal diffusivity of the permafrost layer. In the absence of published values for thermal properties of phonolite, the primary rock in the study area, we use the diffusivity of 25.2 m²/year $(8 \times 10^{-7} \text{ m}^2/\text{s})$ for pure basalt (Hartlieb et al. 2016), the mineralogy of which has similar thermal properties (Horai and Simmons 1969). We expect some interstitial ice in the permafrost layer; ice has a higher thermal diffusivity which would result in a shorter, hence less conservative, diffusive timescale. With the diffusivity of pure basalts, Equation (3) yields an e-folding timescale of \sim 3500 to \sim 6500 years for Z of 300–400 m, as observed by us away from the coast (Figure 2). This period falls within the mid-to-late Holocene climate period during which long-term regional climate has been relatively stable, with a slight cooling of up to a few degrees towards the present time (Cunningham et al. 1999; Baggenstos et al. 2018). Hence, below we assume that the permafrost thickness determined is in a steady state.

In the absence of surface temperature measurements from our study area, we use the standard (2 m above ground) mean annual temperature of -20 °C from this region (Doran et al. 2002; Monaghan et al. 2005) plus an offset of +2 °C to account for ground temperatures being a few degrees warmer in high polar regions than the corresponding mean annual temperature (e.g. Brown 1966). This is consistent with surface temperatures projected from Dry Valleys Drilling Project (DVDP) borehole #3 on nearby Hut Point Peninsula



Figure 2. Results and calculations derived from electromagnetic measurements. In each, a cross-section is shown along the same red line marked on the map. (a) Permafrost thickness, as calculated by depth from the surface to the 100 Ω ·m layer. Contours are in 50 m increments. Permafrost is very thin near the coast and increases in thickness more gradually inland. The cross-section shows a profile through the resistivity data that were used to calculate permafrost thickness. The resistivity results are faded below the DOI. (b) The vertical temperature gradient from the bottom of the permafrost to the top. Contours are in 10 °C/km increments. Values are capped at 200 °C/km (near the ocean, where the permafrost is thin). (c) Vertical heat flux (mW/m²) through the permafrost. Contours are in 10 mW/m² increments. Capped at 300 mW/m² near the coast where permafrost is thin.

(Decker and Bucher 1982), a variety of temperature proxies from the McMurdo Sound Regions (Nichols and Ball 1964) and observed mean annual air temperatures at McMurdo Station and Scott Base. For the temperature at the base of the permafrost, we use the freezing point of seawater of approximately -1.9 °C. We choose the freezing point of seawater rather than that of freshwater because we commonly observe very low electrical resistivities beneath the permafrost layer, e.g. $\sim 10 \,\Omega \cdot m$. Such bulk resistivity is difficult to obtain for freshwatersaturated mafic rock, given that $\sim 10 \,\Omega$ ·m is by itself a reasonable resistivity for groundwater. Using Archie's law (Equation 1), this would imply that the material underneath the observed permafrost layer consists of unreasonably high fraction of groundwater. However, seawater has resistivity of $0.36 \Omega \cdot m$ and would yield bulk resistivity of saturated rock at sensible porosities of \sim 11–27% (Foley et al. 2015). The fact that our study area is in a coastal setting yields additional support to our assumption of seawater as the interstitial fluid beneath permafrost. With the mean annual surface temperature of -18 °C and basal temperature of approximately -2 °C, the temperature jump across the

permafrost layer, ΔT , is ~ 16 °C. This, together with permafrost thickness derived from our resistivity data allows us to calculate the geothermal gradient, which is then multiplied by thermal conductivity to obtain GHF. At low porosities thermal conductivity of basalts is slightly below 2 W/m C, whereas the thermal conductivity of ice itself is slightly higher than that (Horai 1991). Here we use 2 W/m C in our calculations of GHF, which is nearly the same as the thermal conductivity observed at the middle of DVDP borehole #3 at Hut Point Peninsula (Decker and Bucher 1982), which has a similar geology to our study area.

The greatest limitation of the 1D heat flow model (Equation 2) is that it cannot account for the lateral heat flow from the ocean towards land (Risk and Hochstein 1974). We use a 2D finite difference heat conduction model to account for the impact of steady-state horizontal heat transfer from the seafloor towards land. The bespoke MATLAB code solves the standard 2D heat diffusion equation (Recktenwald 2004):

$$\frac{\partial T}{\partial t} = D\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2}\right) \tag{4}$$

where *t* is time and *x*, *z* are the two spatial coordinates, and *D* has been assumed to be uniform throughout the subsurface domain. The seafloor is prescribed seawater temperature of -2 °C, whereas land surface temperatures are assigned using sea-level temperature of -18 °C and a regional lapse rate of 0.0098 °C/m from Doran et al. (2002). The lower boundary condition is constant heat flux, equal to the prescribed geothermal flux, which is assumed to be constant across the lower boundary of the model domain. Both lateral boundaries are set to have zero horizontal heat flow. Initial conditions are given by temperatures following a linear vertical temperature gradient, equal to the assumed geothermal gradient, from the bottom of the domain to the land surface or seafloor.

The numerical solver uses the explicit forward time, centred space technique (Recktenwald 2004). The code was verified by solving numerically a Dirichlet 2D problem on a rectangle and then comparing the model output to an analytical harmonic solution. Although the code solves a transient partial differential Equation (4) given above, we use it to approximate a steady-state solution by running it for long enough so that no element within the numerical domain changes its temperature by more than 10^{-6} °C per year. This typically happens after more than 20,000 model years of holding all boundary conditions constant.

Results and discussion

Analytical results of permafrost thickness and heat flux

Using the $100 \Omega \cdot m$ layer as an indicator of unfrozen ground, we found the thickness of permafrost (or glacier ice, at Barne Glacier) across $\sim 20 \text{ km}^2$ of the coastal western flank of Mt Erebus (Figure 2a). Permafrost is very thin at or near the coast and increases in thickness rapidly over the first $\sim 500 \text{ m}$ distance from the coast. Permafrost must necessarily be thin at the coast because the ocean temperature is steady at about $-1.9 \,^\circ\text{C}$, whereas mean annual air temperature at sea level is about $-18 \,^\circ\text{C}$ (Doran et al. 2002; Monaghan et al. 2005). This results in very high horizontal heat flow from the ocean to the nearby land (including permafrost and glaciers).

The thin coastal permafrost is reflected in our derived analytical models. Given the low resistivity, we assume the base of the permafrost is at -1.9 °C (the freezing point of seawater) and the mean annual air temperature is -18 °C at sea level, with an environmental lapse rate of -6.5 °C/km (Wallace and Hobbs 2006). Using these values and the permafrost thickness, we calculate a vertical temperature gradient from bottom to top of the permafrost (Figure 2b). The temperature gradient is high near the coast due to the influence of the ocean, but quickly settles to ~ 45 °C/km inland. To estimate the uncertainty on this geothermal gradient, we use the square-root of the sum-of-squares assuming $\sim 10\%$ standard deviation on ΔT (approximately \pm 1.6 °C) and 5% standard deviation on Z (approximately \pm 20 m for a 400-m thick layer). This estimate yields one standard deviation of 11%, which for the geothermal gradient of \sim 45 °C/km is \pm 5 °C/km.

We estimate vertical heat flux (Figure 2c) by assuming a thermal conductivity for basaltic material with ice filling its pores of 2.0 W/m C (based on Horai 1991). The vertical heat flux is very high ($> 300 \text{ mW/m}^2$) near the coast, but settles to $\sim 90 \text{ mW/m}^2$ inland; this is slightly high for the global average but reasonable for the region, which has high geothermal flux (Decker and Bucher 1982), especially given the location of the site on the flank of Mt Erebus, an active volcano, and in an area of regional extension. We add an estimated uncertainty in thermal conductivity of basaltic material of 10% (i.e. one standard deviation of $\pm 2 \text{ mW/m}^2$) to the uncertainty in geothermal gradient calculated above to arrive at one standard deviation for GHF of 15%, i.e. approximately $\pm 13 \text{ mW/m}^2$ for $\sim 90 \text{ mW/m}^2$.

Past measurements of temperature gradient and heat flux from boreholes on Ross Island have noted the importance of heat flow from the ocean (Risk and Hochstein 1974). Even accounting for the ocean's influence, heat flow is variable in this region. Risk and Hochstein (1974) measured fluxes of $164 \pm 60 \text{ mW/m}^2$ at Arrival Heights on Hut Point Peninsula (near McMurdo Station, see Figure 1), which they attributed to recent volcanic activity on the peninsula. Our data come from an old part of the volcano with a complicated history of subsequent eruptive and intrusive events (Esser et al. 2004). Because our technique does not measure heat flux directly, but rather derives it from its effect on permafrost thickness, it may integrate heat sourced from older and more recent volcanic events, and thus be less biased by recent events, than a single borehole measurement could be.

Our 2D numerical heat flow model includes the effect of lateral heat transfer from the ocean on subsurface temperature distribution and calculated permafrost thickness (Figure 3a). Comparison of the permafrost thickness distribution derived from resistivity observations with the permafrost thickness predicted by the model shows, however, a significant mismatch (Figure 3b). The model predicts that the excess vertical heat flux caused by the heat exchange between land and the ocean should decay relatively rapidly inland, within a distance equal to about one to two permafrost thicknesses (Figure 3c). Beyond this distance, the impact of conductive heat transfer between the land and ocean domains should be quite small. In our model, this leads to a very rapid rise in the permafrost thickness near the coastline so that the simulated permafrost thickness reaches half of its "inland" value within just a few hundred metres from the coast. By contrast, the permafrost



Figure 3. Temperature field from a 2D heat flow model (a). The scale is chosen to show only the temperature range within the permafrost layer. Surface topography and the position of the coastline are taken from the same profile line shown above in Figure 2. This example is for assumed geothermal gradient of 60 °C/km. (b) Comparison between the permafrost thickness distribution based on resistivity data (circles) and simulated permafrost thickness for two different values of geothermal gradient: 60 °C/km (dashed line) and 50 °C/km (solid line). (c) The high magnitude of excess vertical heat flux near the coastline. The excess vertical heat flux (%) is calculated by subtracting the prescribed geothermal heat flux from the total vertical heat flow and then dividing the difference by the prescribed geothermal heat flux. The blue line shows the distribution at 100 m below sea level, the red line at 200 m below sea level and the black line at 300 m below sea level.

thickness based on the $100 \Omega \cdot m$ resistivity threshold rises only slowly over the first several hundred metres and then more rapidly further from the coast. We concur with the suggestion of Risk and Hochstein (1974) that near-shore permafrost on Ross Island, and perhaps in general, may be subject to seawater intrusion. As seawater intrudes into colder permafrost it may become cryo-concentrated into hypersaline brine, leading to development of a "slushy" permafrost zone whose bulk resistivity drops below the $100 \Omega \cdot m$ threshold used here to define frozen ground.

Conclusions

Although direct measurements of such fundamental geologic and geophysical properties like geothermal

heat flux, groundwater salinity and permafrost are desirable, practical or environmental considerations often render them unattainable in many polar locations. In a span of one day, we collected electromagnetic resistivity data that allowed us to infer some of these properties and achieve results consistent with past measurements within the region. We estimate that the permafrost layer (here defined as near-surface material with bulk resistivity > 100 Ω ·m) is 300–400 m thick in the region, except in the near-shore environment where it rapidly decays to nil at the coastline. This decay is partly due to lateral conductive heat flux from the warm ocean to the cold adjacent land. We also speculate that intrusion of seawater into coastal permafrost further impacts the properties of the nearshore permafrost, likely making it $< 100 \Omega \cdot m$ in resistivity. Our resistivity data do

not suggest any positive evidence for local hydrothermal circulation (e.g. localised thinning of the permafrost layer) and do not necessitate invoking geologic control on subsurface resistivity distribution on the flank of this active volcano. Based on relatively low resistivities of the sub-permafrost material, we speculate that basaltic rocks are saturated by seawater in this coastal zone. The estimated geothermal gradient away from the coast is $\sim 45 \pm 5$ °C/km and geothermal flux $\sim 90 \pm 13$ mW/m². These values are close to the middle of the range for other measurements of geothermal gradient and geothermal flux in this region (e.g. Morin et al. 2010).

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Disclosure statement

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