Surface towed electromagnetic system for mapping of subsea Arctic permafrost

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Abstract

Sea level has risen globally since the late Pleistocene, resulting in permafrost-bearing coastal zones in the Arctic being submerged and subjected to temperature induced degradation. Knowing the extent of permafrost and how it changes over time is important for climate change predictions and for planning engineering activities in the Arctic environment. We developed a controlled source electromagnetic (CSEM) method to obtain information on the depth, thickness, and lateral extent of marine permafrost. To operate in shallow water we used a surface towed electric dipole-dipole CSEM system suitable for deployment from small boats. This system was used to map permafrost on the Arctic shelf offshore Prudhoe Bay, Alaska. Our results show significant lateral variability in the presence of permafrost, with the thickest layers associated with a large river outflow where freshwater influx seems to have a preserving effect on relict subsea permafrost.

Keywords: Controlled source electromagnetics, Permafrost, Arctic hydrology, Gas hydrates

1. Introduction and Motivation

Temperatures cold enough to form permafrost have been present in the Arctic since the end of the Pliocene (about 1.9 Ma) and as a result permafrost underlays roughly 20% of land in the northern hemisphere (*Collett et al.*, 2011). Permafrost is defined as ground that has been held at or below 0°C for at least two years, “ice-bearing” specifies that the pore spaces contain a mixture of ice crystals and pore fluid while “ice-bonded” describes cementation of the sediment grains in the permafrost (*Osterkamp*, 2001). For the remainder of the manuscript we will use “permafrost” to describe a combination of ice-bearing and ice-bonded permafrost.

Over the last 20,000 years sea level has risen 120 m and submerged portions of Arctic coastline and permafrost. The warm, saline conditions introduced by sea level rise are causing the relict subsea permafrost to thaw (*Osterkamp*, 2001). The maximum extent of permafrost offshore, therefore, is the 120 m isobath (*Collett et al.*, 2011), but current seismic studies suggest that permafrost does not extend beyond the 20 m isobath (*Brothers et al.*, 2012). The reason for discrepancy is uncertain (*Collett et al.*, 2011), and a number of offshore measurements of permafrost with simple and effective ways are required.

Permafrost often has associated gas hydrate, which requires low temperature and high pressures to be stable and typically exists beneath the deep ocean floor. However, consistent cold ground temperature in the presence of ice-bearing permafrost allows gas hydrate to currently exist on land and in shallow water environments. Without the permafrost cap to thermally stabilize them, hydrates could dissociate and release their bound methane

26 gas. Methane is a potent greenhouse gas and its release from beneath sub-
27 sea permafrost would influence current and future warming trends (*Ruppel*,
28 2011).

29 A map of our survey area can be found in Figure 5. Prior studies con-
30 ducted on the Beaufort Shelf, Alaska, have used the seismic refraction method
31 to infer extent of permafrost. Seismic refraction uses the increase in P-wave
32 velocity (V_p) due to ice in pore spaces to determine the presence of per-
33 mafrost. Theory predicts a V_p of 2.5 - 2.8 kms^{-1} for sands with less than
34 40% ice saturation, and above 40% saturation a sharp jump in seismic ve-
35 locity to between 3.4 and 4.35 kms^{-1} due to the onset of cementation of sed-
36 iment grains (*Johansen et al.*, 2003). In the Beaufort Sea velocities varied
37 between 1.7 - 4.6 kms^{-1} with all permafrost layer refractions (PLRs), defined
38 as $V_p > 2.3 \text{ kms}^{-1}$, occurring within 30 km of the coast and shallower than
39 20 m water depth (*Brothers et al.*, 2012).

40 Although PLRs can be used to define the lateral extent of permafrost,
41 the seismic method has difficulty determining depth boundaries and bulk
42 properties of permafrost (*Brothers et al.*, 2012). The velocity contrast at the
43 top boundary is so large that most of the seismic energy does not propagate
44 beneath it. The bottom permafrost boundary is likely diffuse and is a velocity
45 decrease, both of which result in a weak or non-existent refraction signal. As
46 a consequence, the seismic method can image the top of the permafrost layer
47 at best. Additionally, the refraction method cannot differentiate between
48 frozen and unfrozen fine grained sediments, which means the depth obtained
49 from this method may not be the top of the permafrost layer if fine sediments
50 overly coarse ones (*Brothers et al.*, 2012).

51 Electromagnetic methods complement seismic data by resolving the top
 52 and bottom boundaries, as well as estimating the volume fraction of per-
 53 mafrost (e.g., *Hauck*, 2016; *Todd and Dallimore*, 1998). CSEM is sensitive
 54 to the resistivity increase caused by ice forming in pore spaces, as well as to
 55 the resistivity decrease below the permafrost as the sediments transition to
 56 an unfrozen state. Onshore boreholes have measured ice-bonded permafrost
 57 with resistivities from 100 Ωm to over 1000 Ωm to depths of 660 m near
 58 the coast of Prudhoe Bay (*Collett et al.*, 1989; *Jorgenson et al.*, 2008). Be-
 59 low the permafrost layer resistivities drop to between 5 and 15 Ωm (*Collett*
 60 *et al.*, 1989). We expect the resistivity of offshore ice-bearing permafrost to
 61 be lower than the measured onshore values because of intrusive saline pore
 62 fluids (*Daniels et al.*, 1976). Land borehole measurements of hydrate beneath
 63 permafrost in the Eileen accumulation, located near the coastline just west of
 64 Prudhoe Bay, reach 2000 Ωm at a depth of about 600 m (*Collett et al.*, 2011).
 65 Gas hydrate occurs at greater depths and has a higher resistivity than per-
 66 mafrost, and so with careful interpretation one might be able to distinguish
 67 between the two.

68 **2. Methodology**

69 Shallow seafloor resistivity, a good proxy of marine geological settings,
 70 is effectively mapped using a fixed-offset towed array (e.g., *Constable et al.*,
 71 2016). Towed systems have been used extensively in deep water and are
 72 faster to use than deploying seafloor CSEM receivers. However, shallow
 73 water depths on the Arctic shelf make using a deep-towed system unfeasible
 74 and make a surface towed system more appropriate.

75 2.1. Instrument Design

76 We had previously developed a deep-towed fixed-offset CSEM system to
77 map gas hydrates, the “Vulcan” system of *Constable et al.* (2016). This de-
78 sign was adapted for a surface-towed system by repackaging our Vulcan deep-
79 tow receivers in buoyant PVC cases. We call the assembly a “Porpoise” since
80 they occasionally appear to leap from the water when encountering ice. The
81 data logger within the plastic case records inline electric field, acceleration
82 in three directions, and a timing pulse sent from an external GPS receiver.
83 Inline electric field is measured using a 2 m dipole held 0.67 m underwater.
84 As with the Vulcan instruments, we chose rigid antennas instead of cable
85 arrays to avoid motional noise associated with cable motion through Earth’s
86 magnetic field (*Constable et al.*, 2016). The Vulcan system is equipped with
87 the ability to receive real-time data and telemetry from the receivers. These
88 features are necessary for maintaining a steady receiver depth, and to pre-
89 vent unwanted contact with the seafloor, as well as monitoring data quality
90 during tows lasting many days. Neither of these concerns affect Porpoises
91 because their positions can be verified by sight and they are typically only
92 towed for hours at a time, allowing for frequent data quality checks without
93 the real-time capabilities. A schematic of the Porpoise system is shown in
94 Figure 1 and pictures of the instrument can be found in Figure 2.

95 GPS receivers are attached to the top of the Porpoises on short masts
96 and a serial data logger within the logger package records time and location.
97 A second serial data logger records pitch, roll, and heading from a com-
98 pass/tiltmeter every second. Conductivity, temperature, and depth (CTD)
99 were continuously measured in the top meter of sea water by towing a CTD

Figure 1: A: Surface-towed Porpoise array. Multiple receivers are easily clipped into the tow rope during deployment. B: Side view schematic of a Porpoise receiver. C: Plan view of a Porpoise receiver.

100 package alongside the vessel. When towing into deeper water, the CTD sys-
 101 tem was lowered to the bottom once the vessel was stopped and the Porpoises
 102 were on deck to measure a conductivity profile. A total of twelve conductivity
 103 profiles were collected and reached a maximum depth of 10 m.

104 The transmitter operates off 110 - 240 VAC power and can output up to
 105 50 A on a 50 m antenna in the form of an arbitrary GPS stabilized ternary
 106 waveform. Data were collected over two years, in the first year a 50 m
 107 antenna was used with between 30 and 40 amps output. In the second
 108 year, the transmitter was upgraded to produce a current controlled output
 109 of 40 amps on a 45 m antenna.

Figure 2: A: Porpoise floating on the surface after being deployed. B: Porpoise ready to be deployed. C: All four Porpoises stored on deck, electrodes and GPS receivers still need to be attached.

110 The entire Porpoise system is light and small enough to be easily air
111 freighted, transported in one load using a utility vehicle, and operated off
112 a small (16 m) fishing boat. This makes it a relatively easy and inexpen-
113 sive system to operate, and while designed to operate in the shallow Arctic
114 shelf, it has already been used for a number of other applications including
115 groundwater and geothermal exploration in the Atlantic and Gulf of Califor-
116 nia, respectively (e.g., *Martin*, 2015).

117 Shown in Figure 3 are noise spectra of Porpoises at three offsets compared
118 with published noise spectra of a Vulcan deep-tow instrument and an ocean
119 bottom electromagnetic (OBEM) receiver (*Constable et al.*, 2016). At 1 Hz
120 noise levels are on the order of a few microvolts and at 30 Hz the noise
121 contributes tens of nanovolts. The peak centered around 0.3 Hz is due to
122 wave noise or strumming of the instrument. Porpoise receiver noise is higher
123 than that of the Vulcan because of increased wave and environmental noise
124 on the surface. The noise floor of the OBEM is the lowest because it is
125 subject to fewer motional noise sources.

126 2.2. Sensitivity to target

127 We conducted synthetic forward modeling to determine the offsets and
128 frequencies most sensitive to permafrost, modeled as a 200 m thick layer of
129 100 Ωm material buried at a depth of 200 m in 1 Ωm marine sediment. This
130 model was intentionally chosen to be conservative and was constructed based
131 on modelling and well log data (*Osterkamp*, 2001). Forward responses were
132 calculated for the permafrost model as well as a halfspace of pure sediment.
133 In both cases water depth was 5 m and the calculation was done for offsets
134 of 100 to 1100 m and frequencies of 1 to 40 Hz. Figure 4 shows (A) the

Figure 3: Power spectra of various EM receivers. Porpoise data were collected on a test cruise offshore San Diego in May of 2014. Vulcan and OBEM spectra are from *Constable et al.* (2016) and are shown as broken lines.

135 amplitude of the permafrost model response normalized by the halfspace
136 response and (B) the difference in model phases. Largest sensitivity to a
137 permafrost layer occurs at offsets longer than 500 m and at frequencies less
138 than 30 Hz.

139 In our survey design Porpoises were towed at offsets of 250, 500, 750 and
140 1000 m. We find that including a shorter range Porpoise helps constrain the
141 shallow sediment and ultimately helps resolve permafrost in inversions of the
142 data. The array was kept at 1 km in length for logistical reasons, but we
143 experimented with moored seafloor receivers that measured two orthogonal
144 directions of the horizontal electric field. Moored receivers were weighted and
145 marked with a buoy so that they could be deployed and recovered by hand.
146 Source receiver offsets of up to 6 km were obtained using these receivers. We
147 transmitted a 1 Hz fundamental of Waveform-D, a modified square wave that
148 spreads power from the fundamental frequency into higher odd harmonics
149 spanning at least one decade (*Myer et al.*, 2010).

150 3. Survey Area

151 The surface towed Porpoise system was used in the summers of 2014 and
152 2015 on the Beaufort Shelf off the north coast of Alaska, with operations
153 based out of Prudhoe Bay on the North Slope. The Beaufort shelf is 80
154 - 100 km wide with water depths less than 20 m to more than 20 km off-
155 shore. The shelf was not glaciated during the late Pleistocene, rather was
156 exposed to subfreezing temperatures and formed permafrost hundreds of me-
157 ters thick. Holocene sediments are missing over much of the Beaufort shelf
158 because of the recent (7.5ky) 40 m sea level rise (*Hill et al.*, 1985). Instead,

Figure 4: A: Amplitude anomaly or permafrost model amplitude normalized by a 1 Ωm halfspace amplitude. An amplitude anomaly of 1 indicates no difference. B: Phase anomaly or phase difference between the two models. The permafrost model was a 200 m thick permafrost layer with 100 Ωm resistivity buried at a depth of 200 m in 1 Ωm sediments.

159 Tertiary and Quaternary sand, silt, and clay make up the bulk of shallow
 160 shelf sediments (*Houseknecht and Bird, 2011*). Sedimentary rocks in this
 161 region can be divided into 4 sequences: the Franklinian (Cambrian to Devo-
 162 nian), Ellesmerian (Mississippian to Jurassic), Beaufortian (Jurassic-Lower
 163 Cretaceous), and Brookian (Cretaceous-Holocene) (*Hubbard et al., 1987*). All
 164 known and inferred gas hydrate on North Slope is found in the Brookian se-
 165 quence (*Collett et al., 2011*).

166 Our surveys were conducted off the R/V Ukpik during late July/early
 167 August in 2014 and 2015, shortly after the ice pack broke away from shore. In
 168 total we towed 12 days and covered 365 km, on average 30 km per 12-hour day
 169 (compare this to 3 km per 12-hour day when using deployed receivers (e.g.,
 170 *Weitemeyer et al., 2006*)). Initially the plan was to tow shore perpendicular

Figure 5: Ship tracks from both years of data collection. A: borehole near Harrison Bay. B: borehole near Prudhoe Bay. Both years the edge of the ice pack was near the barrier islands. An inversion of Tow 3 is shown in Figure 8.

lines coincident with seismic data collected by the U.S. Geological Survey. However, the annual sea ice pack limited how far offshore we were able to collect data and so instead we focused our efforts on expanding coverage along the shoreline. In 2014 the edge of the sea ice was near the 20 m isobath, where seismic data indicates the edge of the permafrost lies. Although in 2015 we returned later in the season in hopes of collecting data farther offshore, the sea ice was in fact closer to shore and prevented us from towing past the seismically determined edge of permafrost. Harrison Bay was the one location where we successfully collected data across the presumed edge of permafrost. Tow lines for both years are shown in Figure 5.

4. Results and Discussion

The inline electric field time series measured on each Porpoise was Fourier transformed and stacked into 60 second windows (*Myer et al., 2010*). Stacking the data increases its statistical reliability and provides estimates of er-

Figure 6: Fence diagram of offset apparent-resistivity pseudosections offshore Prudhoe Bay and the Sagavanirktok (Sag) river outflow for both years computed for 3 Hz amplitudes. Blue corresponds to higher resistivity values (up to $30 \Omega\text{m}$) while red is more conductive (down to $1 \Omega\text{m}$).

185 rors. Processing was done for eleven harmonics of Waveform-D, from the
 186 1 Hz fundamental to 33 Hz. We created pseudosections by calculating the
 187 forward solution for 1D models, including appropriate water depth from an
 188 echosounder and seawater conductivity from CTD measurements, underlain
 189 by halfspaces ranging from 0.1 to $1000 \Omega\text{m}$ and interpolating to find the
 190 best resistivity for each value of stacked amplitude. Appropriate skin depths
 191 at each frequency can be used to create a frequency pseudosection at one
 192 source-receiver offset (e.g., *Weitemeyer and Constable, 2010*), but with four
 193 different offsets we can create the more conventional offset pseudosection for
 194 each frequency (*Hallof, 1964*). Pseudosections are useful for displaying the
 195 lateral variability of a data set, but have limited ability to determine depths
 196 (*Weitemeyer et al., 2006*). The pseudosections for a frequency of 3 Hz are
 197 shown in Figures 6 and 7, short offsets are plotted as shallow structure and
 198 long offsets as deeper structure.

199 The offset pseudosections (Figures 6 and 7) show significant lateral vari-
 200 ability in permafrost extent. Harrison Bay appears very conductive with

Figure 7: Fence diagram of offset apparent-resistivity pseudosection in Harrison Bay and near the Colville river outflow computed for 3 Hz amplitudes.

201 highest resistivities near $10 \Omega\text{m}$, which is unexpected given that permafrost
 202 is seismically inferred in this area (*Brothers et al.*, 2012). However, it is con-
 203 sistent with well logs that show thinning of the ice bearing permafrost layer
 204 onshore to the west of Prudhoe Bay (*Collett et al.*, 1989). Onshore at Har-
 205 rison Bay a well log (labelled A in Figure 5) measures maximum resistivity
 206 near $100 \Omega\text{m}$ down to 120 m, where resistivity drops to over $10 \Omega\text{m}$ and the
 207 base of the ice-bearing layer is seen at 300 m (*Collett et al.*, 1989). If most of
 208 the 120 m thick resistive layer has thawed, the underlying permafrost would
 209 be consistent with resistivity values in the pseudosection. Along the coast
 210 between Harrison Bay and Prudhoe Bay there is a clear layer of conductive
 211 sediments overlying a buried layer of more resistive material, interpreted as
 212 permafrost. Just east of Prudhoe Bay there is an increase in resistivity at
 213 all depths, implying that permafrost is pervasive in the area near the Saga-
 214 vanirktok (also called the Sag) river outflow. Because of freshwater influx
 215 from rivers and melting of sea ice, sea water conductivity was variable across

the entire survey area. In the top meter ocean conductivity ranged from 0.3 to 0.5 Ωm , but at depth remained more constant near 0.4 Ωm . Seawater conductivity variations and water depth were included in the 1D model layers used for pseudosections and therefore do not cause the apparent resistivity variations.

Inversions were done using the MARE2DEM adaptive refinement finite element code (*Key, 2016*). Two dimensional (2D) inversion of CSEM data is computationally expensive, and many inversions must be carried out to fully understand error structure and the trade off between misfit and model complexity (e.g., *Constable et al., 2015*). We present one preliminary inversion in Figure 8 to illustrate the difference between pseudosections and actual inversion. The inversion shown is for Tow 3, indicated in the map of the survey area in Figure 5, and is limited to amplitude, the three largest harmonics of Waveform-D (3, 7, and 13 Hz) for the first three Porpoises (offsets of 250, 500, and 750 m) and the largest two harmonics (3 and 7 Hz) for the last 1000 m offset Porpoise. The inversion grid size ranged from quadrilaterals 150 m wide by 10 m deep in the shallowest region up to 250 m by 50 m at greater depths. The increasing grid size is meant to mimic the loss of spatial resolution with depth as well as decrease computational resources. The inversion results are shown in Figure 9 as a contour plot with contours at 3, 20, and 100 Ωm to better differentiate the likely ice-bonded permafrost layer. Due to the variation in water conductivity across this region, we allowed a 2-layer ocean model be free parameters in the inversion, bounded by the maximum and minimum values that were measured with our towed CTD. The inversion chose 0.5 Ωm for the top layer and 0.45 Ωm for the bottom,

Figure 8: Inversion of Tow 3, a line spanning the Sag River outflow (bottom). The river outflow begins about 8 km inline distance. Pseudosections for 3, 7 and 13 Hz, the three inversion frequencies, are plotted above in order of increasing depth penetration. Note that data contributing to the conductive artifact present in the pseudosections near 3 km inline distance have been removed in the inversion. The inversion fits to an RMS of 1.5 with an error floor of 3%.

241 which are consistent with the upper range of the CTD measurements.

242 On the northwestern end of Tow 3, in the section before the Sag River
 243 outflow, we see conductive sediments overlying two resistive patches. The
 244 southeastern side of the inversion shows a thin layer of conductive marine sed-
 245 iments at the surface, underlain by a non-uniform layer of resistive material.
 246 Below the resistive layer the inversion returns to higher conductivities consis-
 247 tent with freshwater sediments. The peak resistivity in the tow 3 inversion is
 248 over 300 Ωm , which we consider consistent with ice-bonded permafrost. The
 249 pseudosections for the 3 inverted frequencies of Tow 3 are plotted above the
 250 inversion result, and results are consistent with the inversion. The inversion
 251 fits the data to an RMS of 1.3 with a 3% noise floor and a comparison of
 252 data and model amplitudes is shown in Figure 10.

Figure 9: Same inversion of Tow 3 shown in Figure 8, but plotted with four major contours colored with the same colorscale as Figure 8. The first regime is for marine sediments with resistivities $< 3 \Omega\text{m}$, followed by resistivities $< 10 \Omega\text{m}$ which could be partially thawed ice-bearing permafrost or freshwater pore fluids, and the threshold for ice-bonded permafrost was set at $> 20 \Omega\text{m}$ (Collett *et al.*, 1989) with over $100 \Omega\text{m}$ indicating regions of exceptionally high ice saturation. The dark bar at the bottom of the figure indicates the region of the Sag River outflow.

Figure 10: A: Amplitude at each frequency and offset for data inverted in Figure 8, normalized by the median amplitude of each offset. B: Inverted amplitude for each frequency and offset, normalized by the same values as the data amplitudes. Gaps show data that has been filtered out due to poor signal to noise ratio or from amplifier clipping.

253 Because CSEM and seismic refraction methods are sensitive to different
 254 physical phenomena, we do not expect them to yield identical information,
 255 and indeed if they did it would be redundant to conduct both surveys. Strong
 256 seismic refractions occur when there is a sharp velocity increase while the
 257 CSEM method is more sensitive to the total ice fraction. In the region of
 258 the Sag River outflow, a seismic refraction survey (*Brothers et al.*, 2012)
 259 measured an increased p-wave velocity ($V_p > 2.8\text{km/s}$) where the CSEM
 260 pseudosections show resistive anomalies over $30\ \Omega\text{m}$. Together, these obser-
 261 vations indicate an extensive amount of permafrost correlated with the river
 262 outflow. Seismic refraction data continues to see elevated p-wave velocities
 263 ($V_p > 2.8\ \text{km/s}$) from northwest of the Sag River through Harrison Bay, while
 264 the resistivity values decrease. Resistivity still provides a positive indication
 265 of permafrost, but lower resistivity implies that there is less total ice volume,
 266 and is again in agreement with the thinning trend from Prudhoe Bay to Har-
 267 rison Bay seen in onshore permafrost. The pseudosections would not resolve
 268 a thin layer of ice-bonded permafrost and are likely producing an average
 269 resistivity of ice-bonded permafrost and surrounding marine sediments.

270 In contrast, to the immediate southeast of the Sag River, seismic refrac-
 271 tion data detected lower velocity increases ($V_p > 2.3\ \text{km/s}$) consistent with
 272 ice-bearing, but not ice-bonded permafrost. CSEM pseudosections show con-
 273 tinued elevated resistivity ($>10\ \Omega\text{m}$) before transitioning to higher conduc-
 274 tivities at short offsets (shallow) and resistivities of $8\text{-}10\ \Omega\text{m}$ at long offsets
 275 (deeper). Either the ice-bearing permafrost has a stronger resistivity signa-
 276 ture than velocity increase or the resistivity anomaly in this area is from a
 277 different source, i.e. freshening of porewater fluids. Ice-bonded permafrost

278 is impermeable to water, so any ground water needs to flow under or around
 279 the ice-bonded permafrost. The southeast side of the Sag River is the side
 280 without ice-bonded permafrost (*Brothers et al.*, 2012) and therefore where
 281 fresh groundwater can flow and freshen the pore fluid. Beneath the per-
 282 mafrost layer should be water saturated sediment, offshore this may be brine
 283 or fresh, but onshore it will likely be fresh. Borehole measurements taken
 284 beneath the permafrost layer onshore show resistivities up to 20 Ωm (*Col-*
 285 *lett et al.*, 2011), which implies that freshening of pore fluids offshore cannot
 286 increase resistivity beyond about 20 Ωm . Therefore, we infer that the very
 287 high resistivity values offshore Sag River are from increased ice content. Fresh
 288 ground water flowing beneath this permafrost layer, however, may affect our
 289 ability to detect the base of permafrost in this area.

290 Modeling suggests that groundwater flow preserves permafrost by fresh-
 291 ening the porewater fluid and raising the freezing temperature above the
 292 ambient ocean temperature (*Frederick and Buffet*, 2015). By definition, per-
 293 mafrost is ground that is below the freezing point of water. Therefore, when
 294 groundwater is flowing into areas where permafrost exists it is flowing into
 295 a temperature regime where it will freeze, turning to ice and preserving the
 296 thickness of the permafrost layer. Using a synthetic model of the continental
 297 shelf, Frederick and Buffet [2015] calculated the offshore extent of subsea
 298 permafrost in the presence of a ground water flow which ranged from 0 to
 299 over 140 $\text{m}^3\text{yr}^{-1}\text{m}^{-1}$ over many glacial cycles. They concluded that local
 300 hydrology is as important as paleoclimate conditions and sea level rise in
 301 determining current permafrost extent. While this model is consistent with
 302 the presence of permafrost offshore the Sag River, the same results are not

303 seen offshore the Colville River. Brothers et al. (2012) also note that larger
304 rivers can act as a heat source and prevent permafrost from forming when
305 exposed to sub-freezing temperatures (*Frederick and Buffet*, 2014). Taliks,
306 or unfrozen regions in otherwise continuous permafrost, occur in paleo river
307 channels, such as the Colville River which has a talik extending hundreds of
308 feet beneath its bed (*Williams*, 1970). The Colville is larger than the Sag
309 and may be an example of the latter effect of local hydrology and be acting
310 as a heat source rather than a preserving force.

311 5. Conclusions

312 Our surface towed CSEM system was successful at characterizing resis-
313 tivity to a depth of about 800 m in 0-8 m water depths offshore Prudhoe
314 Bay. Apparent resistivity in the region ranged from 1 Ωm , a value typi-
315 cal of marine sediments, to over 100 Ωm , which we interpret as ice-bonded
316 permafrost. In areas with seismically interpreted permafrost, apparent re-
317 sistivity varied from 10 Ωm to over 100 Ωm , indicating that there is both
318 ice-bearing and ice-bonded permafrost offshore Prudhoe Bay and that EM
319 methods may be capable of distinguishing between the two. Inverted base of
320 offshore permafrost is near 600 m, consistent with well logs measuring an on-
321 shore permafrost layer of 660 m at Prudhoe Bay. Pseudosections and initial
322 inversions both show increased resistivity at the Sag River outflow, offering
323 support to the idea that groundwater preserves permafrost. Future work will
324 include carrying out inversions on all lines, integrating seismic horizons into
325 the inversions, and further exploring the role of local hydrology in permafrost
326 preservation.

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