# Stratigraphic profiling with ground-penetrating radar in permafrost: A review of possible analogs for Mars

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[1] We review our past and ongoing use of ground-penetrating radar to investigate permafrost in Alaska and in the Dry Valleys of Antarctica. The results may be relevant to radar efforts on Mars because of arid conditions and the presence of ice. The pulses were centered at 50, 100, and 400 MHz. We interpret profiles from two sites in the eastern Taylor Valley to show glaciolacustrine and glaciofluvial stratigraphy. The maximum depth of stratigraphy profiled there was about 33 m. Near Fairbanks, Alaska, the depth of penetration at 50 MHz was near 80 m in marginally frozen and stratified alluvial sands. At the Fairbanks sites, supplementary drilling was required to differentiate between reflections from conductive bedrock, a graphitic schist, and those from the water table at depths of 20–25 m. At a site on the North Slope of Alaska, we profiled present and remnant freezing fronts in an alluvial floodplain. The relative permittivity at most sites ranged between about 4 and 5.5, which is consistent with dry conditions, the mineralogy, and low ice content. Weak interface reflectivity or the lack of further interfaces may have limited the interpretation of maximum penetration where no water table was present because signal absorption should have been low and scarce diffractions imply that scattering was weak. The interface reflectivities beneath Taylor Valley may be a function of only density contrasts, since free water, and possibly ice, is absent. *INDEX TERMS:* 0925 Exploration Geophysics: Magnetic and electrical methods; 1823 Hydrology: Frozen ground; 5109 Physical Properties of Rocks: Magnetic and electrical properties; 9310 Information Related to Geographic Region: Antarctica; 9315 Information Related to Geographic Region: Arctic region; *KEYWORDS:* Ground-penetrating radar, permafrost

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## 1. Introduction

[2] The prospect of success in using ground-penetrating radar (GPR) for the detection of ice, water and related features on Mars must rely on experience with Earth GPR exploration in permafrost terrain. Since 1992 ERDC/CRREL has conducted extensive GPR studies of permafrost in the Fairbanks and North Slope areas of Alaska, and in the Antarctic Dry Valleys with the University of New Hampshire. These areas may serve as Martian analogs because of the lack of liquid water caused either by very cold temperatures (Dry Valleys) or by the coarseness of the sediments. Here we demonstrate the ability of GPR to penetrate certain types of permafrost and discuss some examples of reflecting strata that indicate the presence of ice and former presence of water.

[3] We used a commercial transient-type GPR system that radiated pulses centered near 50, 100, and 400 MHz. We traveled to the Antarctic sites by helicopter; the sites in Alaska were easily accessible by an all-terrain vehicle. We fixed our position and elevation with 10-cm accuracy using differential GPS in Antarctica. We correlated photographic, borehole, and test pit information with the GPR results. As with Mars, we expected that density or mineralogy differences in Antarctic sediments might provide the only contrasts in relative permittivity, ε, with depth because of the lack of liquid water (the mean, near-surface ground temperature is −18°C) and likely low ice content, if existing borehole information is representative. In Alaska it is more


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plausible that differences in ice content, in concert with differences in grain size, aided the stratification. Of equal concern in the Dry Valleys were the abundant 20- to 50-cm surface boulders that typify proximal polar glacial drift and made traversing difficult. At depth these boulders could cause scattering loss and clutter. An unknown but potentially important factor in the Antarctic sediments is the elevated salt content of buried soils, the salt being derived primarily from marine aerosols and also reworked salts from glacially transported evaporites and marine sediments. There is some evidence that the glacial drift that our GPR penetrated in eastern Taylor Valley in Antarctica occurs as thin, stacked drift sheets that range in age back to 140,000 years old.

[4] We have previously reported some of these Alaska GPR studies [*Arcone et al.*, 1998a, 1998b]. We are not aware of any other GPR permafrost work that has reached the depths that we have. Similarly it appears that GPR results for the Dry Valleys have only been reported by us [*Arcone and Delaney*, 2000; *Prentice et al.*, 2002, M. L. Prentice et al., Subsurface glacial stratigraphy in lower Taylor Valley, Antarctica based on GPR results: Implications for West Antarctica Ice Sheet history, submitted to *Journal of Quaternary Science*, 2002 (hereinafter referred to as Prentice et al., submitted manuscript, 2002)]. The sediments we have profiled are both stratified silts, sands, and gravels, as well as massive till with low or little ice content and virtually no water. The stratigraphy and 30 m of penetration previously achieved in deltaic, well-drained sands and gravels using 50- and 100-MHz GPR [e.g., *Johl and Smith*, 1991] encouraged our work in Antarctica; profiles of sand deposits in more temperate or tropical environments generally show less than 20 m of penetration [*Bristow et al.*, 2000]. We have also yet to find any reported GPR profiles of lacustrine layering of silts and clays, but this may simply reflect the presumed high rates of attenuation associated with these sediments when wet. The depth of penetration we interpret in lacustrine sediments strongly favors a very dry state.

## 2. Equipment and Methods

[5] We used the portable, battery-powered GSSI 16-bit System 2 control unit in Antarctica in 1998–99, a 16-bit System 10A in Fairbanks in 1994–95, and an 8-bit System 8 on the North Slope in 1992–93. The antennas were a “400-MHz” model 5103 (rated at 1.2 W peak power), a “100-MHz” model 3207 (800 W peak power), and our own 50-MHz resistively loaded dipoles (Figure 1) polarized transverse to the transect directions. The estimated gain of each antenna is less than 5 dB because of the electrically short dipole lengths needed to produce the high-resolution pulses. We hand-dragged the antennas along the rough Dry Valleys terrain and used a vehicle over snow-laden trails in Alaska (Figure 1) while recording continuously at data acquisition rates of 24–48 traces per second. The 1- and 2-m width of the 100- and 50-MHz antennas forced us to choose our paths carefully in Antarctica to avoid antenna–ground decoupling noise. We generally recorded with the range gain at 120 nanoseconds (ns) at 400 MHz, 500–700 ns at 100 MHz, and 1000–1500 ns at 50 MHz.

[6] Processing has included high- and low-pass noise filtering, positional and elevational corrections, and, for some records, automatic gain control and deconvolution. We migrated some records using a two-dimensional Kirchhoff scheme. Diffractions in these records, assuming point sources, give dielectric constants, $\varepsilon$, of about 4–5.5. These are realistic values for primarily felsic frozen sands and gravels. Given the limited penetration and the nearly flat-lying stratigraphy, a three-dimensional migration approach is not warranted.

## 3. Results: Dry Valleys Sites

[7] The Dry Valleys of Antarctica cover about 4000 km$^2$. The polar desert climate and mountain blockage of outflow of the East Antarctic Ice Sheet keep the valleys relatively free of snow and ice compared to other valleys within the Transantarctic Mountains. However, in the past, large outlet glaciers from both the East and also the West Antarctic Ice Sheet advanced into and receded from the valleys, leaving a sedimentary record of their fluctuations. Ice-bonded or dry permafrost may be up to 600 m deep, and the area is often compared with possible conditions on Mars. As with Mars, the origin of many features of the valley geomorphology is controversial. Some argue that the valleys formed largely through fluvial erosion [*Denton et al.*, 1993; *Sugden et al.*, 1995] prior to 20 million years ago and were preserved since at least 14 million years ago in a polar-desert climate or beneath continuous, frozen-bottom glacial ice advances. This implies minor glacial erosion because of the lack of bottom sliding. Dynamicists argue for significantly fluctuating climate and repeated ice sheet glaciation over much of the last 40 million years. They would argue for wet-based ice advancements into the Dry Valleys, characterized by episodes of melting [*Prentice and Matthews*, 1991; *Webb et al.*, 1996; *Prentice et al.*, 1998], such as during the early Pliocene (3–5 mya). The latter scenario implies glacial sliding and repeated episodes of major glacial erosion and fluvial deposition. If the latter case is true, then diagnostic stratigraphy of fluvial origin should be commonly uncovered with GPR and would make this possible Martian analogy all the more relevant.

[8] Our Antarctic sites are located in the eastern Taylor Valley (Figure 2). The sites include Hjorth Hill, which is on the northern valley wall at the eastern (Ross Sea) end, and the Sloth Lake area, which is farther west, near Lake Fryxell. These sites contain hummocky glacial deposits associated with incursions of the marine-based West Antarctic Ice Sheet into Taylor Valley. They are, in fact, considered very important for resolving the late Quaternary history of the ice sheet but are controversial because of the lack of exposure. Our GPR profiles provide the first internal images of these features. The GPR data from these sites have been integrated with the surficial geology and drill core data (Prentice et al., submitted manuscript, 2002). The elevated strandlines throughout Taylor Valley and the widespread occurrence of silty deposits along the lower valley walls leave little doubt that a lake (Glacial Lake Washburn) up to several hundred meters deep at the eastern end existed during the Holocene [*Green and Friedmann*, 1993]. Although hundreds of 1-m-deep pits have been obtained throughout the Dry Valleys, only about 12 deep holes (80–300 m within Taylor Valley [*McKelvey*, 1981]) have been

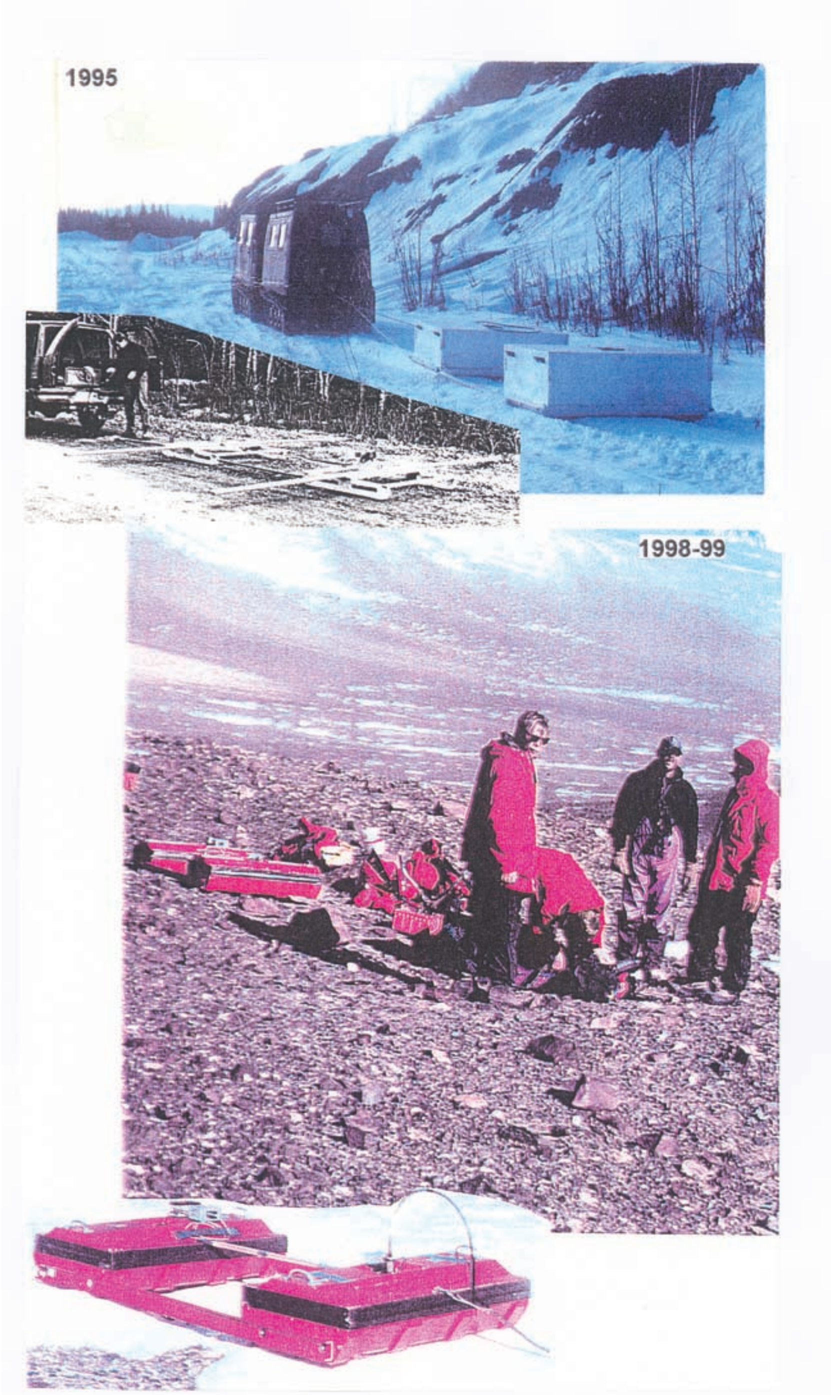


**Figure 1.** Top: Shielded 50-MHz antennas under tow in the Fairbanks area of Alaska; the antenna width is 2 m. Top inset: Unshielded 50-MHz antennas used in the Dry Valleys (photo taken in interior Alaska). Bottom: Equipment and crew in the Dry Valleys. The largest surface rocks were usually 20- to 50-cm boulders of sandstone, granite, and dolerite. Bottom inset: 100-MHz antennas; the width is 1 m.

drilled, and there is no prospect for drilling in the near future because of environmental concerns.

[9] Of all the sites discussed in this review, those of the Dry Valleys may be most similar to possible Martian conditions. The penetration into lacustrine sediments implies that subsurface conditions are extremely dry, the sediments are ice-bonded but not ice-saturated, and the medium to coarse sands show 30–40% black grains derived from the McMurdo area volcanics [*Kyle*, 1981]. Some aspects of the mineralogy are discussed later.

### 3.1. Sloth Lake

[10] This area (Figures 3 and 4) contains numerous sinuous ridges that are a few meters in height, 10–20 m in width, up to about 100 m in length, and primarily oriented transverse to the valley's long axis in a NE–SW

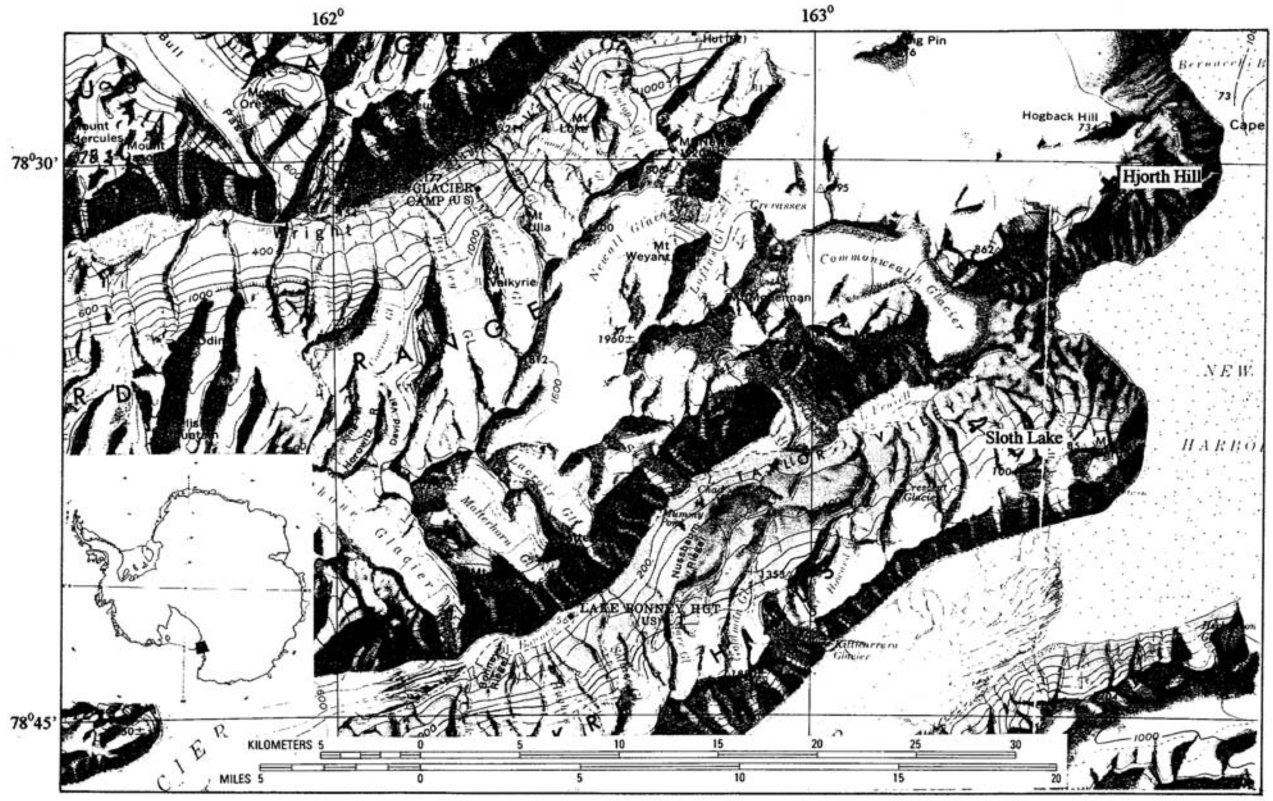


**Figure 2.** Location of Sloth Lake and Hjorth Hill in the eastern Taylor Valley. The white areas are glaciated, the gray areas are valley bottoms, and the dark areas are valley walls. Section taken from USGS Control Number 76190-W1-RR-250, Ross Island and Vicinity (Antarctica), 1:250,000 scale.

direction. The ridge morphology and the surface veneer of volcanic debris derived from across McMurdo Sound suggest that they might be minor recessional moraines deposited at the grounding line of a lobe of the West Antarctic Ice Sheet that advanced into the valley from the east. Several other origins have been considered [*Stuiver et al.*, 1981; *Denton et al.*, 1989, 1991, 1993]. Numerous excavations have revealed very little ice [*Prentice*, 1990].

[11] Figures 5 and 6 show profiles that crossed two of these mounds in the northeast-southwest direction. The 100-MHz profile in Figure 5 shows mainly horizontal reflection horizons, parts of which are discernible in the lower part of the profile to about a 30-m depth between distances of 0 and 95 m. These horizons are interpreted to represent near-horizontal stratification of fine-grained glacial drift deposited in Glacial Lake Washburn. Near the top, especially from 0–80 and 150–190 m, is faint evidence of dipping beds.

**Figure 3.** Aerial photograph (taken Dec 1999) of the Sloth Lake area that contains the mound we profiled (arrow). Northeast is across the valley, to the upper right. The scene is approximately 1 km wide.

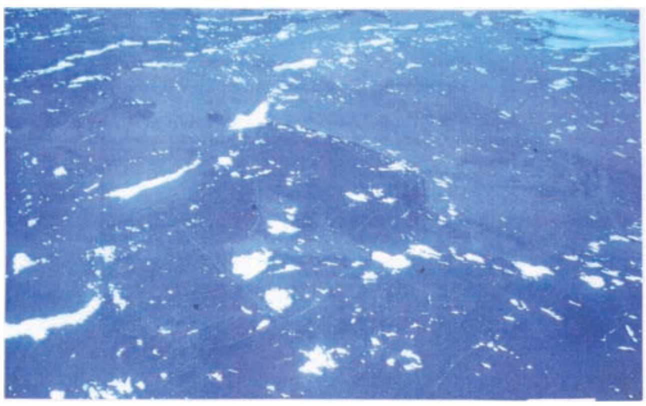

**Figure 4.** Aerial photograph (taken Dec 1999) of the Sloth Lake mound area showing our profile transects (faint lineations). The profile discussed in Figure 6 ran from the lower left (northeast) to upper right (southwest). The scene is approximately 0.5 km wide.

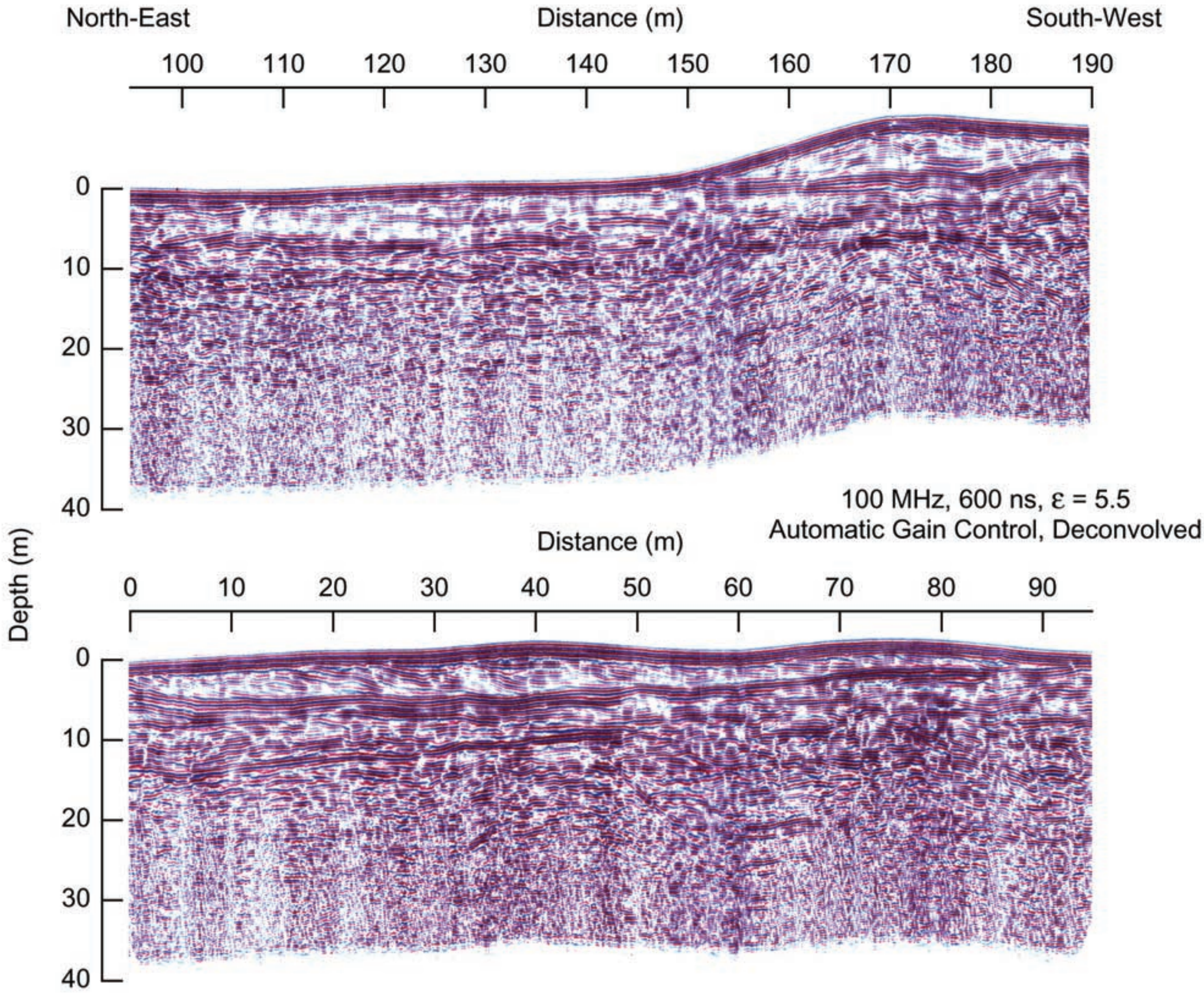


**Figure 5.** 100-MHz profile of the stratigraphy at Sloth Lake.

[12] A 400-MHz profile (Figure 6) shows these near-surface features to be a series of dipping reflectors that extend to a depth of about 6 m. They are crossed by shallow-dipping scalloped-shaped reflectors, which are most prominent between 0 and 80 m. Beneath these beds are the more continuous glaciolacustrine horizons, which also appear between 200 and 245 m. The few diffractions within the dipping beds and in other records indicate an $\varepsilon$ value of 5.5, which is consistent with ice-bonded (but not saturated), dense sands and gravels [*Arcone et al.*, 1998a, 1998b]. Many of these reflections are characterized by simple wavelets, which indicate isolated interfaces. In some cases there are faint, extended cycles that indicate the bottom reflection of a thin layer. Each wavelet has a characteristic phase, which we define as the polarity (plus or minus) sequence of the amplitude dominant central three half-cycles [*Arcone et al.*, 1995]. Red bands indicate positive phase polarity and blue indicates negative. The profile has not been deconvolved, so the phases of the reflected wavelets are retained through the processing. For this model antenna, a red-blue-red sequence indicates a reflection from an interface between a material of higher $\varepsilon$ (and density) under one of lower and vice versa. Higher values of $\varepsilon$ could be caused by interstitial filling between sand grains by silt or ice, so that a higher $\varepsilon$ corresponds with higher bulk density.

[13] We interpret the scallop-shaped reflections to be the surfaces of migrated ripple deposits [*Prentice et al.*, 2002, submitted manuscript, 2002]. They cannot be aeolian dune features because small boulders are present. This ripple deposit surface interpretation is reinforced by the series of dipping reflectors between them, which we interpret to be foreset beds. The steepness of the foreset slopes is an artifact of the horizontal compression used for display. In fact, the slopes are on the order of 10–20°, far less than the asymptotic slope of a model hyperbolic diffraction (shown in the figure). For this reason the migration shown in Figure 6 does not improve the image.

[14] The scalloped horizons are characterized by a blue-red-blue wavelet phase sequence, which indicates an interface between lower-density material below higher-density material, implying that density increases with depth between ripple horizons (e.g., less-dense silt and sand at the top grading into denser silt, sand, and gravel at the

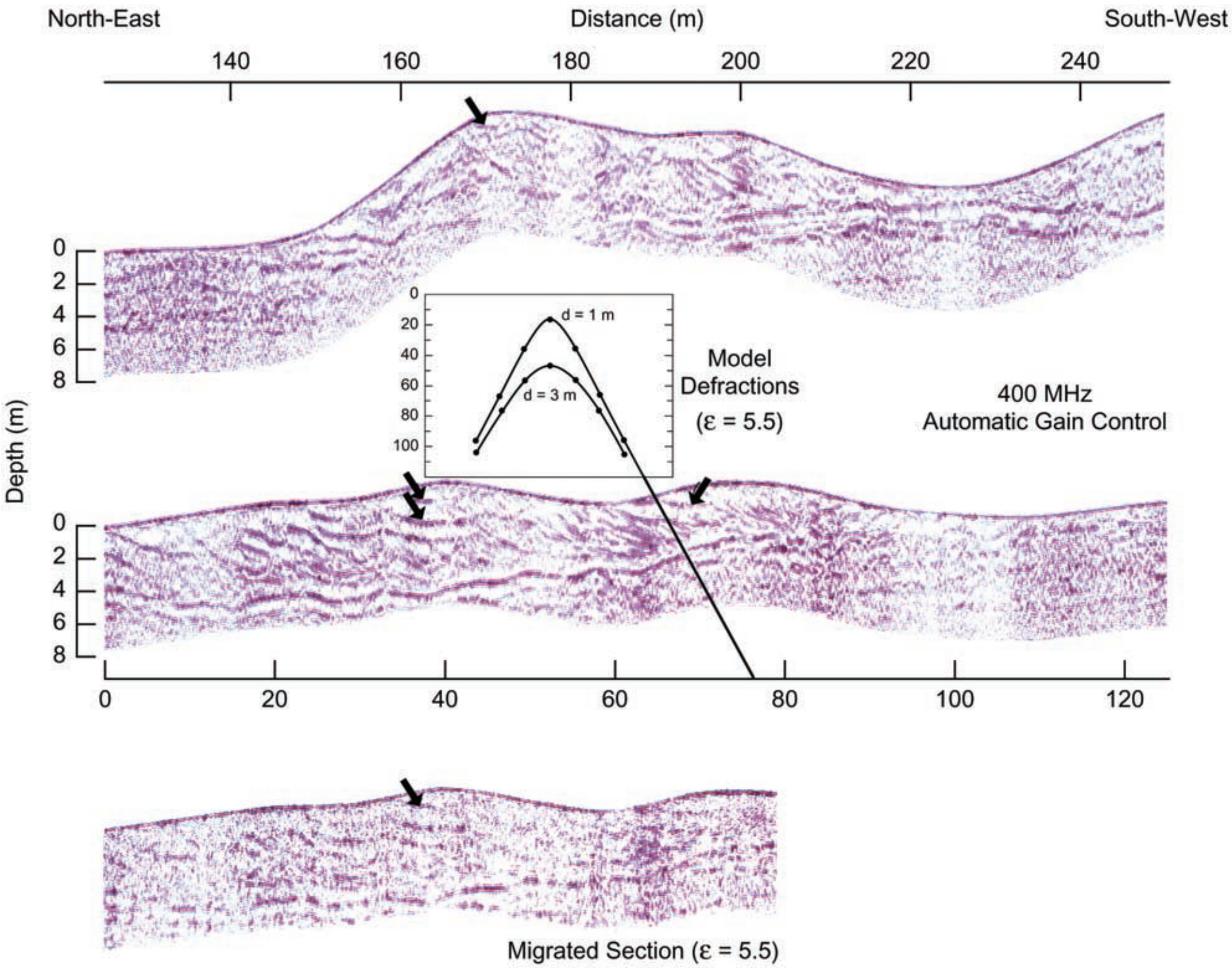


**Figure 6.** 400-MHz profile of the stratigraphy at Sloth Lake. The events in the profile are true reflection horizons because the model diffraction has a much steeper slope than any of them. The bottom profile is a two-dimensional migration of one section. The small arrows indicate a few of the reflection horizons that we interpret to be interfaces between ripple deposits.

bottom). They are also characterized by faint, trailing extra cycles that indicate that the interfaces are actually thin layers of fine-grained material.

### 3.2. Hjorth Hill

[15] This site (Figure 7) is located at the eastern end of the Taylor Valley (Figure 2) on the top of a set of prominent lateral moraines overlooking the Ross Sea. The top of the hill is bare meta-igneous rock. There is currently a debate as to whether the lateral moraines were deposited during the Last Glacial Maximum (18,000 to 21,000 years ago) by the West Antarctic Ice Sheet and possibly local piedmont glaciers or whether older advances were involved as well. The mound of sediment we profiled was about 500 m across and contains a small basin.

[16] Figure 8 compares 50- and 100-MHz profiles of the same transect, which went east-west across the moraine. The deepest horizon has a maximum depth of about 33 m. We think that it represents bedrock because it peaks under the surface topographic high at 400 m. Cross profiles show

**Figure 7.** Aerial photograph (taken Dec 1999) of the Hjorth Hill area that includes the bifurcated mound structure (arrows) we profiled. North is toward the top of the photo. The scene is approximately 1 km wide.

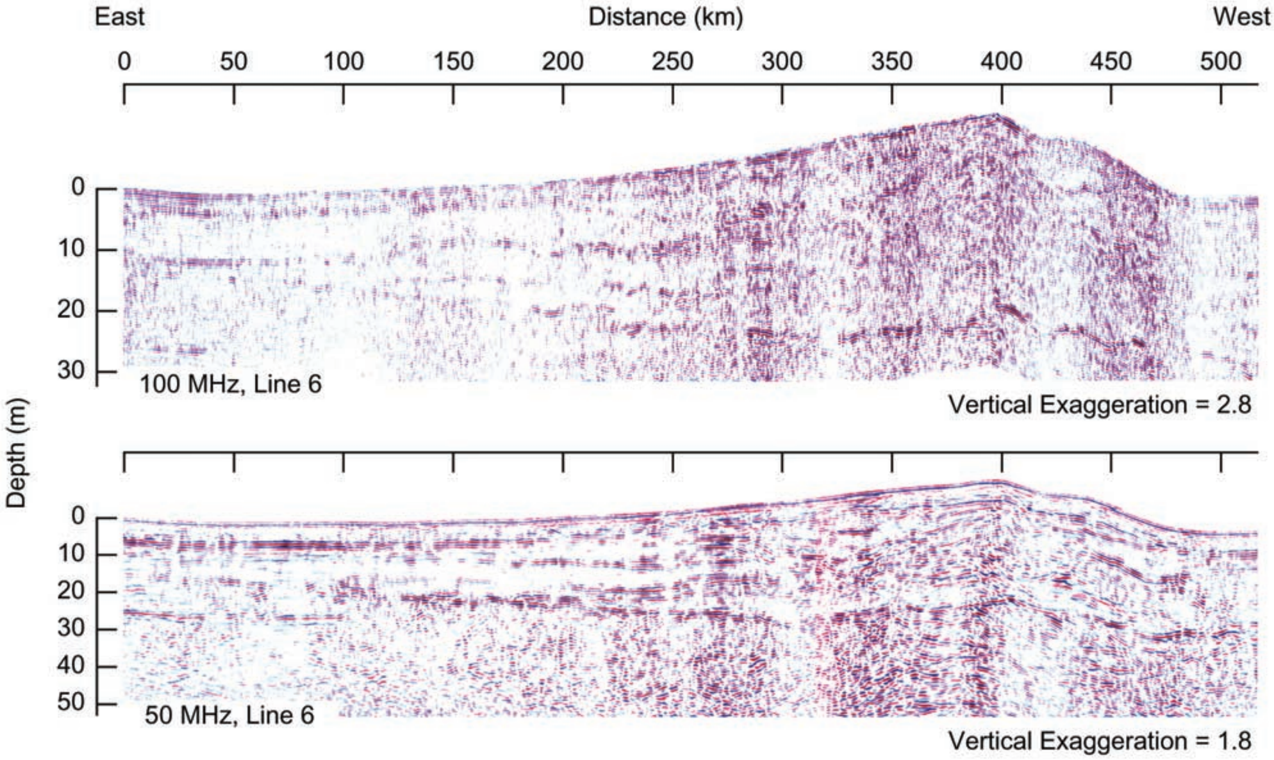


**Figure 8.** Comparison between 50- and 100-MHz profiles across the same transect at Hjorth Hill. Both profiles show about 30- to 35-m penetration. Uphill is into the page.

that it rises near the surface uphill from this transect (Prentice et al., submitted manuscript, 2002), and it is doubtful that this type of bedrock would be stratified. The upper horizons, which are clearer at 100 MHz, may be stratified glaciolacustrine drift deposited by the Ross Sea ice advance, which is known to have reached up this hill [*Denton and Marchant*, 2000]. The 50-MHz profile gives a more continuous response to the lower horizon, but it is also seen at 100 MHz. Therefore, it is unlikely that there are significant loss processes at work because loss mechanisms (e.g., scattering or water and magnetic relaxations) are frequency dependent. Consequently the penetration of about 33 m in this permafrost should only be apparent because there may be no deeper horizons present.

## 4. Results: Alaska Sites

[17] Interior Alaska is characterized by discontinuous permafrost, whereas the North Slope area contains continuous permafrost. Interior permafrost may reach 100 m deep and temperatures of a few degrees Celsius below zero in the Fairbanks area. Permafrost on the North Slope reaches about 600 m deep, and temperatures are much lower. The presence of unfrozen water is primarily controlled by grain size and salinity. Silts and clays retain unfrozen capillary and adsorbed water, while salinity depresses the freezing point. As opposed to the ice-poor examples of the Dry Valleys, our Alaskan profiles demonstrate the GPR penetration and performance possible in ice-rich deposits, which may not be common in the near-surface Martian environment. Where silt is minimal, penetration is greatest. The examples from the North Slope show responses that may be unique to the presence of water and ice.

### 4.1. Fairbanks

[18] Alluvial floodplain deposits are extensive between the present Tanana River and the foothills of the Alaska Range; Fort Wainwright alone covers 665,000 acres and is only a small part of the interior. Attenuation rates at 100 MHz for this icy silt, sand, and gravel are about 0.5–1.0 dB/m [*Arcone and Delaney*, 1989]. Our Fairbanks sites are located on Fort Wainwright, just north of the Chena River (Figure 9). GPR investigations throughout this area are discussed by *Lawson et al.* [1996] and *Arcone et al.* [1998b]. Typical summer and winter permafrost conditions are diagrammed in Figure 10. The discontinuous permafrost here is usually within one degree Celsius of zero. The active layer (the surface layer of seasonal freezing) is at least 0.6 m thick beneath heavy vegetative mats and is usually saturated silts, organics, and fine sands. Bedrock, a highly reflective, quartz-mica graphitic schist, is generally 20–260 m deep. Subpermafrost water is transported by aquifers because in many places permafrost is frozen into bedrock. Consequently we have used GPR as a method to find candidate aquifers. However, the highly reflective bedrock makes for difficult GPR interpretation of aquifers, so we often verified our interpretations by drilling.

#### 4.1.1. Subpermafrost Water

[19] Figure 11 shows two perpendicular profiles recorded at site 1 (Figure 9). The interpretation in the figure is for profile 6 and is aided by eight well logs. This profile appears to cross several aquifers, one of which we profiled

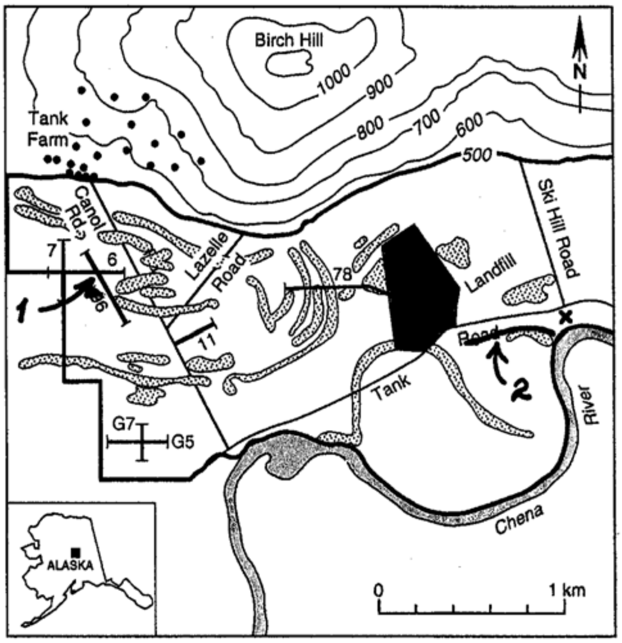


**Figure 9.** Location of profiles (sites 1 and 2) on Fort Wainwright, in the Fairbanks area.

along transect 96. Profile 6 is filled with reflections and diffractions; without drilling here we could not distinguish bedrock reflections from groundwater reflections. We determined an ε of 5.4 for these sediments, both from several well logs and from hyperbolic fittings of the diffractions [*Arcone et al.*, 1998b]. This value puts the deepest water at about 23 m deep. Despite the uniform value of 5.4, the diffractions indicate a large degree of inhomogeneity. Their presence, however, also indicates that single scattering occurred.

#### 4.1.2. Deepest Penetration

[20] We recorded this profile (Figure 12) at 50 MHz at site 2 (Figure 9) because the proximity of the Chena River suggested that permafrost might be shallow and underlain by the water table. The profile shows the deepest events that we have recorded in permafrost. A water table would provide strong reflections, but none are apparent in the profile. Instead, the profile shows stratification and no diffractions. Subsequent drilling in 1995 found bedrock below 70 m of ice-rich sand at the 90-m distance but no water table or the bottom of permafrost. Using a depth calibration based on an estimated ε of 4 for ice-rich, felsic sands, the faint events indicated by arrows occur at about 80 m deep and may indicate features near or beneath the bedrock surface.

### 4.2. North Slope

[21] The Alaskan North Slope area is the piedmont and outwash zones located between the Brooks Range and the Beaufort Sea. Permafrost is about 600 m deep here, and the active layer is about 30 cm thick. Nearer the coast about 40–50% of the land surface is covered by lakes of a few meters depth, and the ground itself is ice-rich. The floodplains of the many rivers that reach from the mountains to the sea may be up to 6–10 km wide. The braided streams of these floodplains provide enough seasonal thawing to force permafrost to start at a few meters deep.

[22] In January and April 1993 we performed a GPR study of seasonal migration of freeze fronts associated with

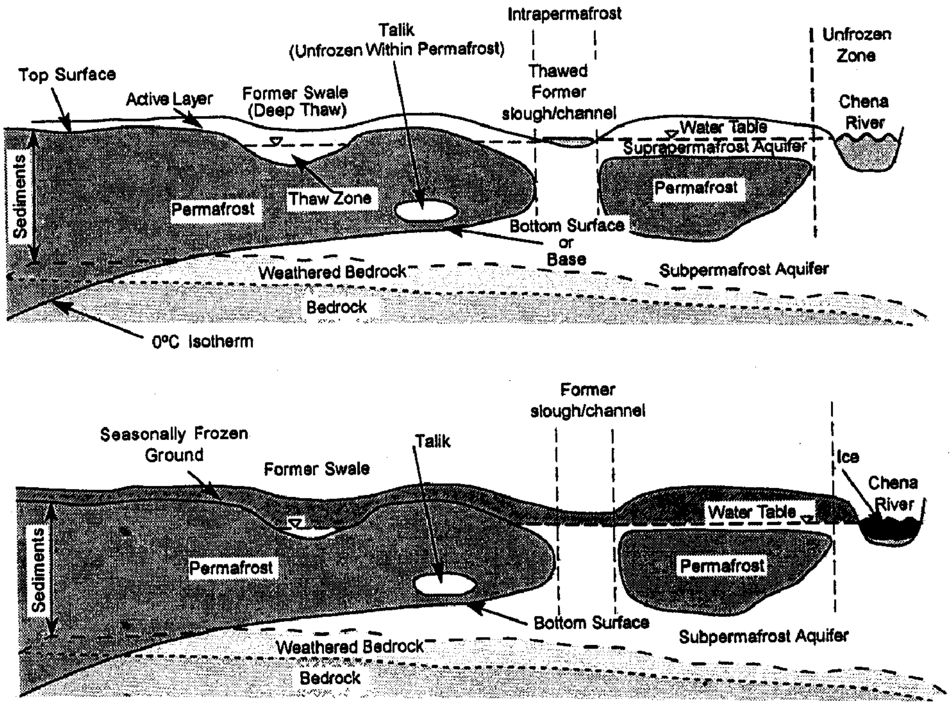


**Figure 10.** Idealized sketch of the configuration and nomenclature for discontinuous permafrost and groundwater aquifers associated with it; (top) summer conditions and (bottom) winter conditions.

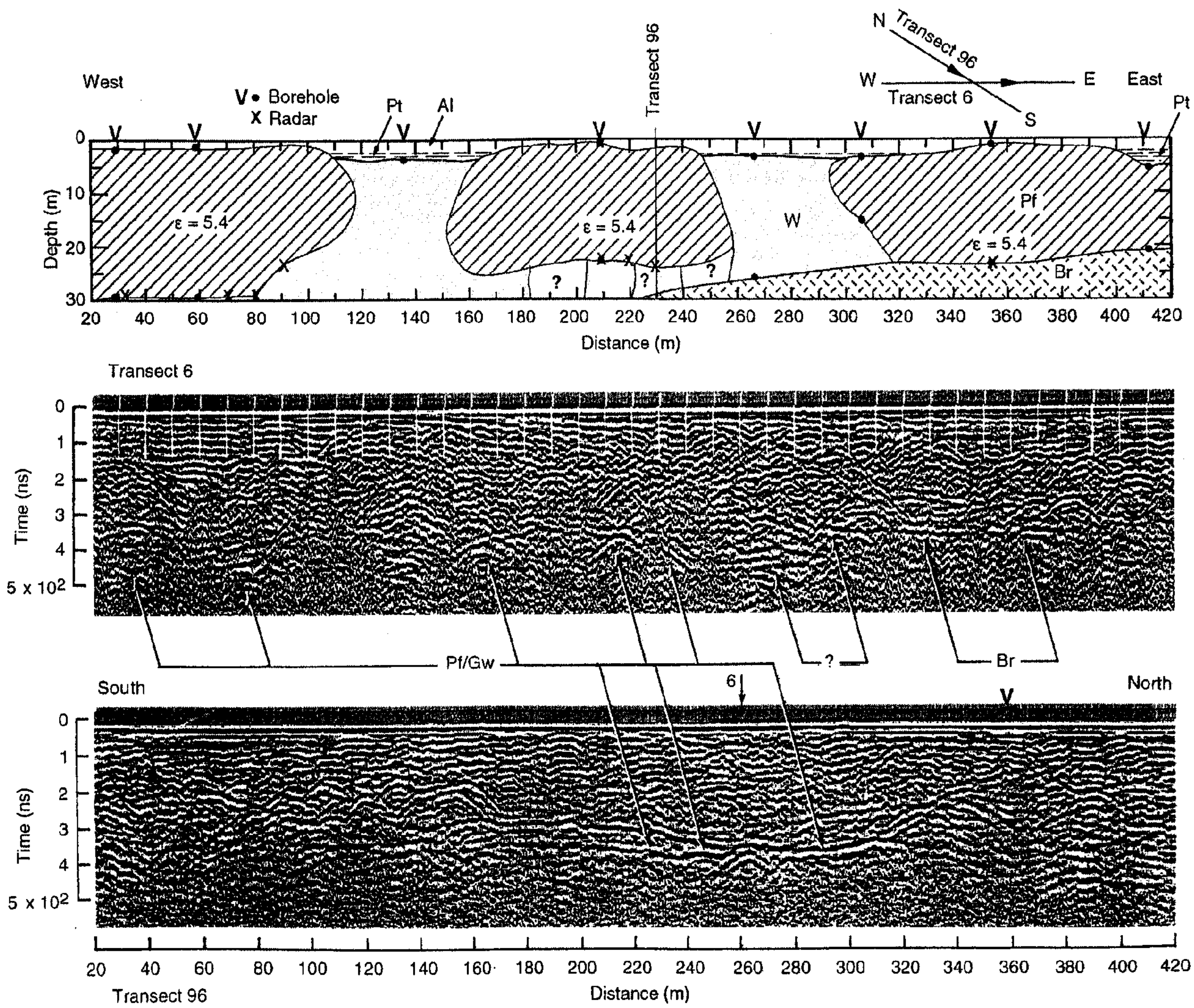


**Figure 11.** 50-MHz profiles of two transects over discontinuous permafrost (site 1, Figure 9), and an interpretation of transect 6 [from *Arcone et al.*, 1998b]. Subpermafrost groundwater appears discontinuous along transect 6 but continuous within 96. The symbols used are: Al, active layer; Pt, perennial thaw; Pf, permafrost; Br, bedrock; and Gw, subpermafrost groundwater. The question marks within the interpretation indicate probable unfrozen and unsaturated sediments; within the radar profile they indicate unexplained horizons in one of the interpermafrost thaw zones (W).

thaw zones beneath the braided streams of the Saganavirktok River, which runs through Prudhoe Bay [*Arcone et al.*, 1998a]. If such thaw zones (or "bulbs") persist for more than one year, they are known as taliks, which generally refers to thaw zones above or below 0°C. If they are below 0°C, they may be referred to as cryopegs. Here we loosely refer to them as taliks. The study area is between the Dalton Highway and Franklin Bluffs about 20 km from Prudhoe Bay (Figure 13). The snow thickness generally ranged between 5 and 25 cm. The profile transects usually crossed the stream directions. One cannot identify any particular stream from the outdated topographic quadrangle maps.

#### 4.2.1. Control Lines

[23] The profiles in Figure 14 were obtained, along with drilling, to verify the presence of free water and to obtain talik depths in order to determine ε for the cyclically frozen ground above them. The reflections from the bottom of the channel ice and from the top surface of the talik have the same phase (tone sequence in their reflection bands) and are, therefore, from interfaces between materials of lower ε above and of higher ε below. Intermittent reflections from the bottom of the talik are evident only in profile B. The hyperbolically shaped diffractions from within the taliks indicate an inhomogeneous permittivity structure. Weak reflections and diffractions throughout the frozen material indicate slight variations in its permittivity.

[24] The drilling depths and reflection times give an ε between 3.5 and 5.5 for the frozen alluvium above the taliks. The higher values are more typical for this material [*Arcone and Delaney*, 1989; *Arcone et al.*, 1992] and represent an ice content of probably less than 20%. The lower values indicate an ice content in excess of 70% [*Delaney and Arcone*, 1984; *Arcone and Delaney*, 1989]. We interpret a decrease in drilling resistance encountered

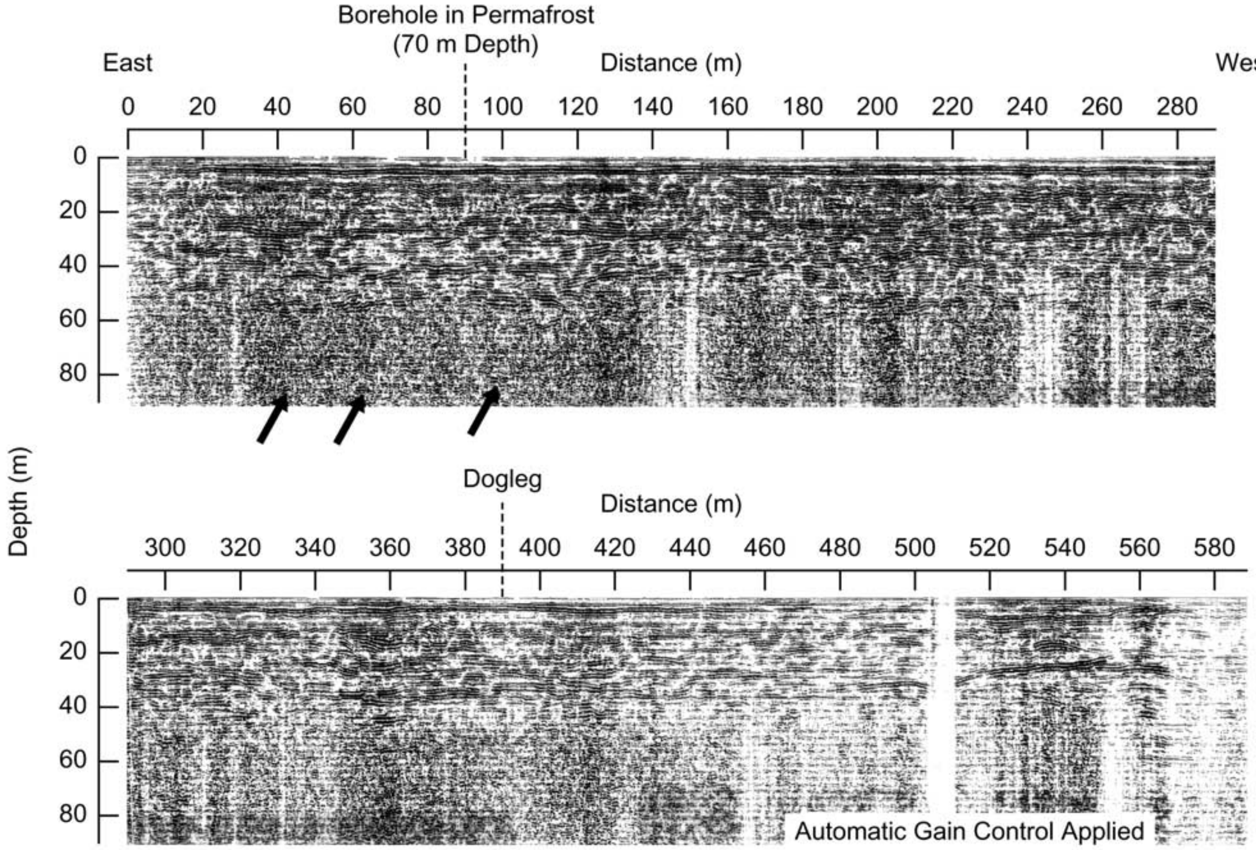


**Figure 12.** 50-MHz profile from Fort Wainwright, near the Chena River. The arrows indicate deep reflections that may originate beneath the permafrost. We applied automatic gain control (AGC) and additional low-pass filtering to alleviate the noise that accompanied the AGC process.

just before free water was reached at 33-m distance in profile A to indicate an ice layer, whose reflection is marked in the profile. The free water from the talik rose approximately 60 cm above the ground surface and then settled to a slow trickle that lasted over 30 minutes.

#### 4.2.2. Partial Freezing and Ice Layers in a Talik

[25] Cross-sectional profiles of a talik profiled in January and April 1992 are compared in Figure 15. In January the talik was thickest beneath the deepest part of the channel, where it had just lost contact with the ice bottom. Only a small talik remained in April, and a lower surface is barely discernible. The difference in phase between the talik top and bottom reflections (trace 385 in Figure 16) verifies the expected contrasts in $\varepsilon$ across the interfaces. The horizon between distance points A and C within the January talik appears to be from the bottom of a partially frozen zone that extends to the talik surface. The relative strength of this reflection is seen in trace 311 of Figure 16. Partial freezing is also consistent with the generally weaker reflection from the talik surface near point B, where only 1.5 cycles are visible in the profile. The greater strength of the signal reflected from the talik surface outside of points A and C makes all 2.5 cycles visible and is consistent with a completely unfrozen state. As seen in the previous example, there is evidence of an accretionary ice layer (event c in the profile) on the surface of the talik.

[26] An interpretation of the talik structures is given in Figure 17. From the control line measurements we assume that $\varepsilon$ linearly decreased from 4.7 at less than 2-m and greater than 40-m distances, to 3.7 near the center at 22 m for the frozen sediments surrounding the upper surface of the talik in January. The 3.7 value corresponds with a volumetric ice content of about 73% [*Delaney and Arcone*, 1984]. If this percentage held throughout the talik, only about 1% of the remaining volume within the upper portion of the talik between points A and C needs to have been unfrozen water to have provided an $\varepsilon$ value of 4.7 (using the complex refractive index mixing model [*Annan et al.*, 1994]) and to have generated the observed weak reflections with the proper phase. Most of the talik appears to have been less than 1 m thick and with a bottom surface of fairly uniform depth. The bottom surface appears to have moved about 1.3 m deeper by April. It is also possible that the April talik was partially frozen.

## 5. Discussion

[27] The penetration of 70–80 m in the Fairbanks area was possible because of the lack of silts or clays, which precludes the presence of any significant liquid water. The lack of water in the sediments precludes the usual dc conductivity and Debye relaxation mechanisms as significant causes of

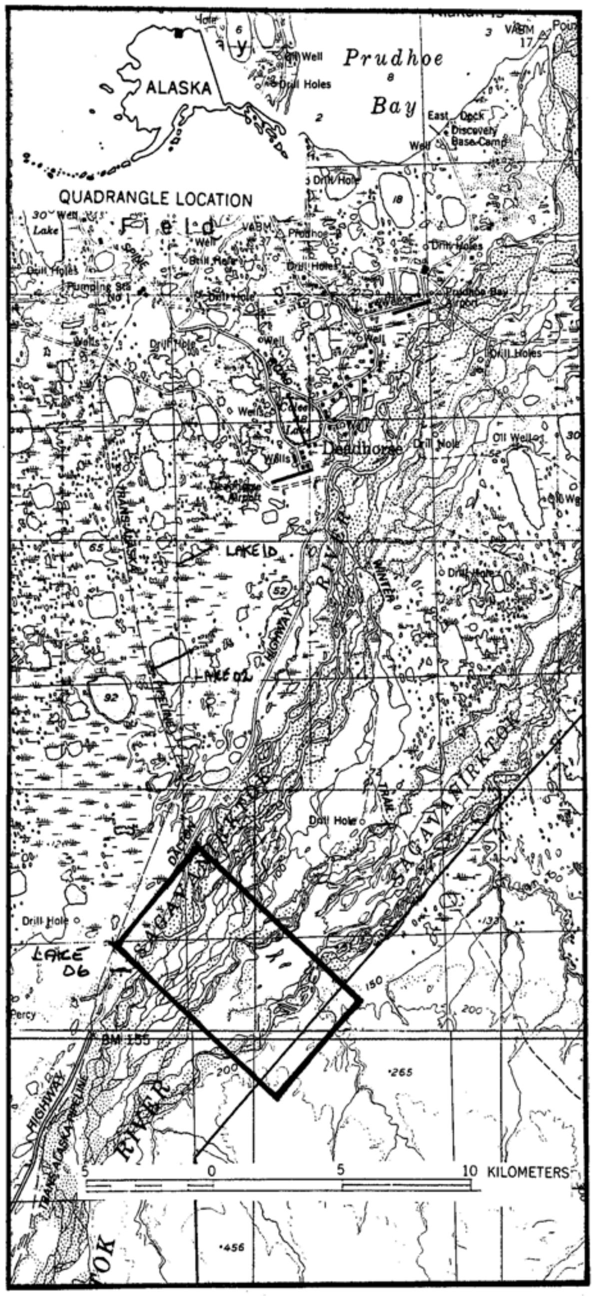


**Figure 13.** Survey area on the Sagavanirktok River floodplain, about 20 km south of Prudhoe Bay.

signal attenuation. In this case of frozen sands the temperature may be of little consequence to penetration so long as the ground is below freezing. In addition, the lack of gravel or cobbles should preclude any losses caused by scattering.

[28] The Fairbanks water table examples are complicated, required drilling for complete interpretation, and represent cases where unfrozen water was present, probably even within the alluvial permafrost and associated with silt. The contrast in ε between frozen and unfrozen water most likely caused the stronger diffractions and reflections. In all these cases the events profiled are local phenomena that could only have been detected with a detailed ground survey.

[29] The deepest penetration in the Dry Valleys of about 33 m represents only the deepest interface that occurred. The lack of diffractions within the 100-MHz profiles suggests weak ε contrasts between sediments and rocks and a sparse volumetric distribution of larger rocks at depth. This is consistent with single scattering theory [*Smith and Evans*, 1972], which predicts only an 8-dB loss over a round-trip distance of 30 m for an estimated 10% volumetric content of boulders (as on the surface) with an ε value of 8 (Ferrar dolerite; mainly calcium-rich feldspars at ε =7 and augite at ε = 7–10) and radii < 1/6 in situ wavelength (< 25 cm) within the sedimentary matrix (minimum ε = 4). Therefore, weak contrasts in ε and interface roughness may have limited the depth of our deepest detectable interface at the Sloth Lake site, while at Hjorth Hill there may have been no deeper interfaces. Consequently, we think that actual penetration may have been far greater.

[30] About 15–20% of the medium to coarse sands from the eastern Taylor Valley that we have observed in cores (archived at the Antarctic Marine Geology Research Facility, Florida State University) exhibit some magnetism. However, the depth of penetration suggests that magnetic [*Olhoeft and Capron*, 1994] and Maxwell-Wagner [*Matzler*, 1998] relaxation losses were very low in the Dry Valleys. Hematite, magnetite, and any other conductive or magnetic minerals are a small fraction of the soil [*Claridge*, 1965; *Bockheim*, 1997]. Low magnetic relaxation losses may contrast with probable Martian areas containing strong concentrations of magnetic minerals in sediments, as might be derived from basaltic rocks.

## 6. Conclusions and Recommendations

[31] Our interpretation of the scalloped and dipping reflections at Sloth Lake indicate a history of strong water flow associated with glacier incursions to deposit these beds of silt, sand, and gravel in ice-proximal positions. In contrast, the North Slope profiles record a history of freeze-front migration. Both examples could serve as references for interpretation of future Martian GPR profile stratigraphy. Within all these profiles the detailed stratigraphic or thermal responses are too localized for an airborne or spaceborne radar to detect.

[32] The penetration we have achieved in permafrost can also serve as a guideline for possible GPR performance on Mars, given similar surficial geology. Our transient-type, short-pulse GPR system was able to penetrate at least 30 m in deeply frozen and probably dry silts and about 80 m in ice-rich sands. The Dry Valleys results are particularly encouraging for Mars because of the penetration in the presence of volcanic material. In all cases maximum penetration may have been deeper because there appears to have been sufficient signal strength to reach reflections from deeper interfaces.

[33] Penetration in dry, felsic sediments at 100 MHz could be greater than 100 m with more-advanced, lower-noise, FMCW-type systems moving at very slow speeds, which allow high rates of signal stacking to improve signal-to-noise ratios and clutter suppression. Penetration would be greatly improved with higher gain antennas, but this would be mechanically awkward and would preclude a high-resolution pulse waveform. In addition, colinear antennas

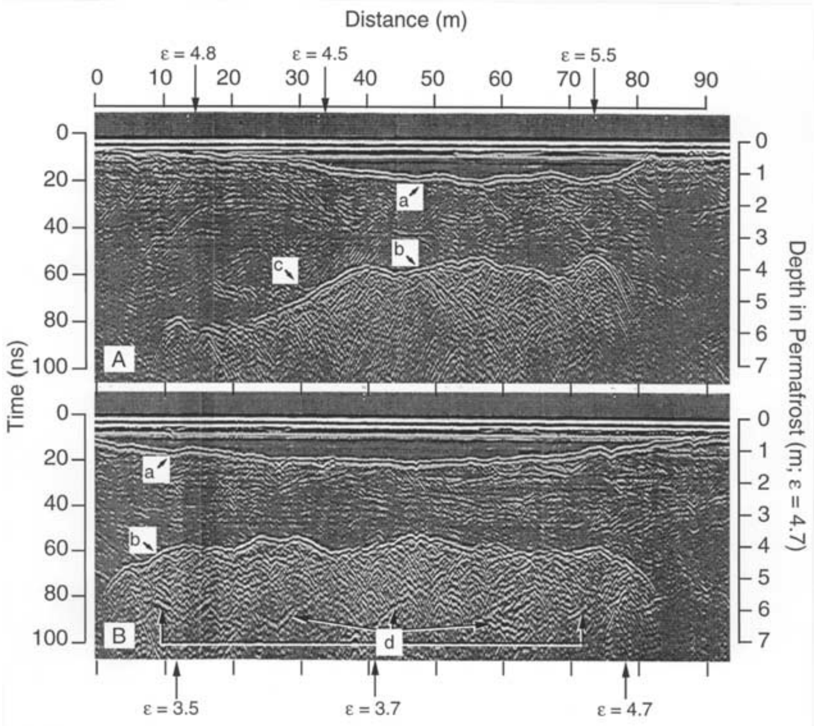


**Figure 14.** Profiles along two sections of a control line. The vertical arrows locate boreholes where we calculated ε for the frozen alluvium above the talik. The labeled events are the bottom of the channel ice (a), the talik surface (b), the ice layer (c), and the bottom of the talik (d). The direct coupling between antennas at 0 ns is the surface reference. The maximum ice thickness is 1.5 m in the top profile and 1.7 m in the bottom. We used an older version of the Model 5103 antennas that transmitted the inverse of the present 5103 waveform.

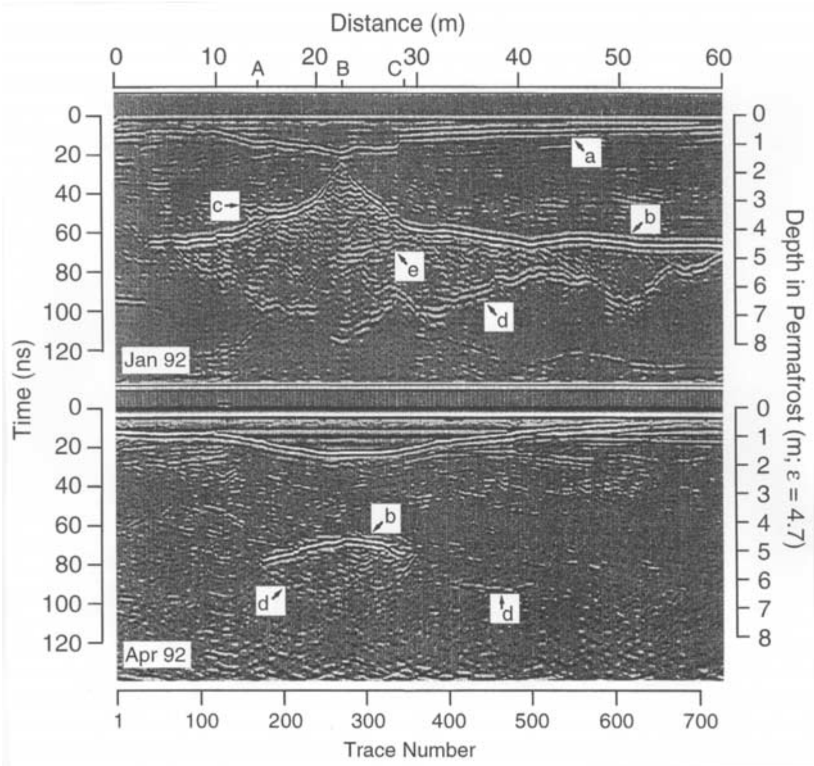


**Figure 15.** Time-migrated seasonal profiles of a talik. The profile events are labeled as in Figure 14. We interpret the area within the talik above event (e) to be partially frozen. The maximum channel ice thickness (the layer along the top) is about 1.6 m.

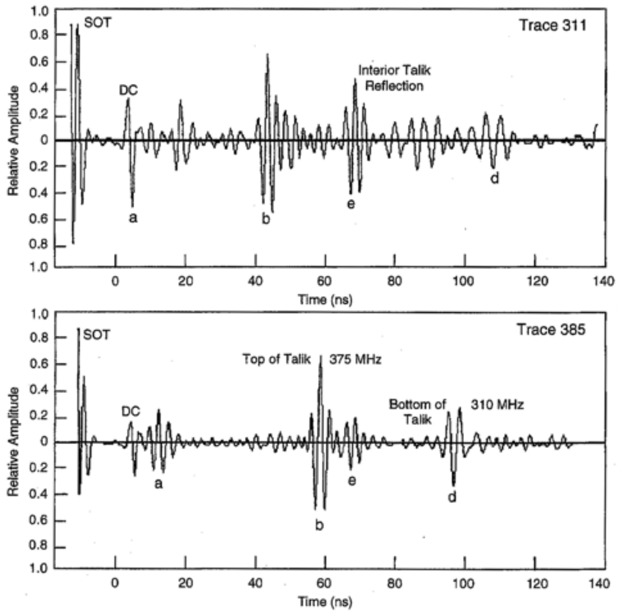


**Figure 16.** Sample traces, with the same letter labels, from the January profile of Figure 15. Added gain makes the waveforms more visible. SOT (start-of-trace) is an artificial signal that triggers recording. DC, the direct coupling (partially filtered) between antennas, represents the channel ice surface. The reflections from the top and bottom of the talik have different phases. Trace 311, at 25-m distance, shows the strong reflection from within the talik. The phase agreement for the ice bottom, the talik surface, and the internal talik reflections indicates increasing values of ε across each interface. The shift in local wavelet frequency to about 310 MHz (trace 385 at 30-m distance) is characteristic of propagation in a wet medium.

[e.g., *Barbin et al.*, 1996] for Mars surveying would avoid antenna decoupling, such as might occur when part of an antenna is lifted by a rock.

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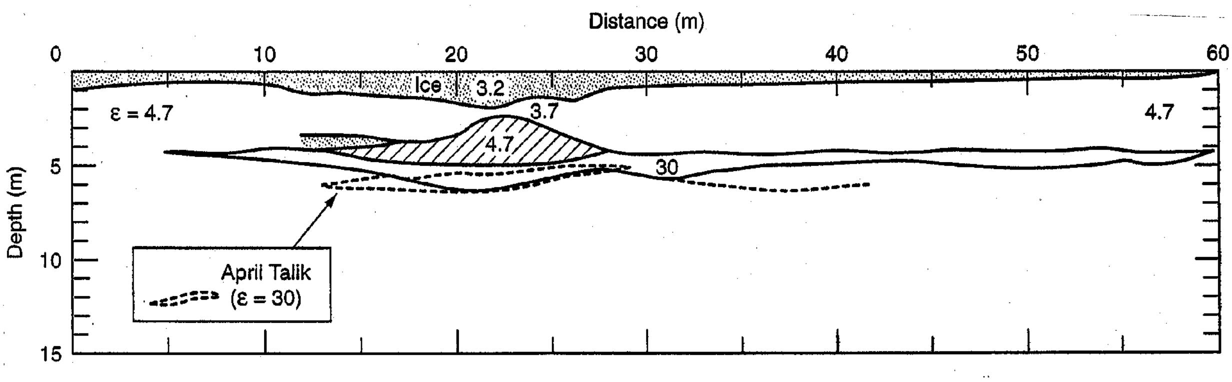


**Figure 17.** Interpretations of the profiles in Figure 15. Permittivity values for each section are labeled. The cross-hatched area is a partially frozen section of the January talik. The talik appears to have migrated deeper by April.

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