Investigations of Freshwater and Ice Surveying Using Short-Pulse Radar

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COVER: GSSI 101C antenna mounted on struts off a helicopter; a short-pulse waveform; and the reflection coefficient of a water layer on ice.
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PREFACE

This report was prepared by Dr. Kevin O’Neill, Research Civil Engineer, Civil and Geotechnical Engineering Research Branch, Experimental Engineering Division, and Dr. Steven A. Arcone, Geophysicist, Snow and Ice Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. Funding for this research was provided by DA Project 4A161102AT24, Research in Snow, Ice and Frozen Ground; Task SS, Properties of Cold Regions Materials; Work Unit 014, Electromagnetic Characteristics of Snow, Ice and Frozen Ground, and DA Project 4A762784AT42, Design, Construction and Operations Technology; Work Unit BS-011, Development of Electromagnetic Subsurface Exploration Systems in Cold Regions. CRREL’s ILIR program also provided some funding for this research.

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NOMENCLATURE

$\alpha$ angle with horizontal
$\sigma$ dc conductivity; as subscript means time index relative to any discrete time point $n$ at which $\chi_\sigma$ begins

$d_{(d)}$ maximum correlation values as a function of layer thickness
$\tau$ time required for wavelet to make round trip through layer depth

$A_X(A_s)$ amplitude spectrum of signal segment $x$ (wavelet)
$S(f)$ Fourier transform of wavelet

$C_t$ correlation function
$t_1$ radar wavelet as a function of discrete time $t$ as a subscript means discrete time index

$d$ depth of layer
$\tau_t$ dielectric relaxation time

$\Phi_{x}(\phi_s)$ phase spectrum of signal segment $x$ (wavelet)
$T_{rel}$ (complex) frequency dependent two-way transmission coefficient for layer

$\epsilon_j$ dielectric constant of medium $j$
$T_{im}$ (complex) interface transmission coefficient for waves propagating from medium $j$ to medium $m$

$\epsilon_0$ permittivity of free space
$\omega$ angular frequency

$g(t(d))$ synthetic echo pattern for layer thickness $d$
$\chi_{\omega}$ section of $X_t$ same length as wavelet

$k$ (complex) wave number
$X_t$ measured radar response as a function of discrete time $t$

$\eta_j$ impedance of $j^{th}$ medium
$R_i$ $i^{th}$ reflection factor

$\tau_{ij}$ (complex) interface transmission coefficient for waves propagating from medium $j$ to medium $m$
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INTRODUCTION

This report describes some of our recent activities in pursuing the long-term objective of remote surveying of ice conditions on fresh-surface-water bodies. Detection of water depth, bottom profile, and related hydrological and geotechnical features is also of interest. Specifically, this investigation used short-pulse UHF radars to evaluate the possibility of airborne surveying of winter hydrological features over large (multi-kilometers) sections of terrain. Some experiences in unfrozen environments will also be recounted. Overall, some field and laboratory experience is presented and theoretical and signal processing items are pursued relating to the particular problems and possibilities in this sort of surveying.

Airborne radar surveying offers the possibility of exploring large sections of terrain in relatively short times. Because the antennas are separated from the ground surface, signal processing strategies based on the search for a reliably well-defined wavelet are facilitated by the lack of variable impedance loading. In recent exploratory work, we have sought to profile ice conditions as well as depth and bottom features of rivers and lakes using commercially available short-pulse radar systems. Antennas were typically mounted on a cantilever off helicopter skids; alternative arrangements were tried on occasion, such as suspension of the antenna in a cargo net below the helicopter, or the very non-airborne practice of submersion of the antenna from a boat.

In general, past attempts at airborne surveying have used radars with pulse center frequencies and bandwidths no greater than about 250 MHz (Dean 1977, Kovacs and Morey 1978, Rossiter et al. 1980). In this frequency range, it is sometimes difficult to distinguish between echoes from top and bottom ice surfaces. This is because one typically determines thickness by pulse timing, clocking the delay between echoes from distinct, well-separated interfaces. When the delay is less than the pulse duration, successive echoes will overlap, resulting in a loss of resolution. In the above mentioned systems, pulse duration was such that resolution would be lost for an ice layer less than about 60 cm thick, hence these systems would be unsuitable for many practical situations. In recent years the possibility of increased resolution has come through the development of smaller and lighter UHF antennas. Successful use in surveying at ground level (Arcone 1985, Arcone et al. 1986) has stimulated interest in testing these systems when the antennas are mounted on a helicopter. While airborne surveying can cover large portions of terrain rapidly, one must contend with coherent clutter from aircraft and ground surface, and also with very large amounts of data that result. Questions of lateral resolution, effects of air speed and altitude, scattering and signal strength losses arise.

Short-pulse radar antennas located on the ground will experience ground loading. That is, unavoidable response by the ground surface will influence currents in the antenna, lowering the overall frequency content and changing the pulse shape. The specifics of this alteration will depend upon the particular characteristics of the ground at hand (under foot). The ground also causes odd lobing in the radiation pattern (Engheta et al. 1982). When the antenna is airborne it radiates a single lobe of minimum pulse duration, although it does so over a large beamwidth. Relatively greater reliability of pulse shapes from airborne antennas greatly facilitates deconvolution and interpretation of records, especially when thin layers are encountered. Pulse distortion in ground based surveying makes pulse timing more difficult, but not altogether impossible. If a recognizable enough pulse is present in the record so that one can at least identify a lump of energy, then it is possible to determine the arrival time of the return at least within a time span roughly equal to the duration of the wavelet. A well-defined wavelet allows one to process the return so that a much more precise arrival time can be identified.

Beyond this, well-defined wavelets will produce well-defined interference patterns with one another.
This is the basis for successful processing of some of our returns from thin ice. In this context, the definition of “thin” depends on the radar system as well as on the geophysical system. The delay between top and bottom echoes from a layer will be equal to twice the layer thickness divided by the wave velocity within the layer. For separation of interface echoes, this delay must be greater than the wavelet duration, which is about 4 to 6 ns for our two antennas with center frequencies between about 900 and 500 MHz respectively. Assuming a dielectric constant for freshwater ice of 3.2, this means that top and bottom returns will begin to overlap when the ice is about 30 or 50 cm thick. For thicknesses on this order or greater, we employ digital processing techniques capable of locating the pulse, and accommodating non-minimum delay wavelets, some wavelet uncertainty and noise. For thinner ice, theoretical thin ice reflections were used whose waveforms contain characteristic interference patterns between successive interface returns. These were used in a matched filter approach with good results. In principle the same thin layer approach can find application in other, contrasting systems. For example, one might use lower frequency radars with much longer wavelet durations to achieve greater penetration in ground-probing measurements. In this case a layer of rock with a dielectric constant on the order of that of ice would appear “thin,” even if it were much thicker than the limits given above.

Beyond the many qualitative inferences one can extract from returns by “eyeball” and beyond the various quantitative treatments by signal processing lie many challenges: rough surfaces and rubbly ice, volume scattering and other loss or pulse distortion mechanisms, and the incoherence of returns from heterogeneous ice-water mixtures. This last item is especially important, because such formations are characteristic of ice jams, which are a major cause of river flooding. After noting the manifestations of these problems below, we analyze briefly the expected effects of critical angle phenomena in ice-free water depth surveying and the effects of wet ice surfaces on our short-pulse returns.

FIELD SURVEYS

Methods and paraphernalia

When resolution of surface ice layers was the main concern, relatively high frequency antennas were used. These consisted of the GSSI model 3102 and 101C antennas with center frequencies between about 500–600 MHz and 900–1000 MHz, respectively, depending upon the loading. Manufacturer’s peak radiated power ratings are 8 and 2.1 W respectively. Both antennas have bandwidths approximately equal numerically to their center frequencies. Each of these could be mounted on struts extending approximately 1.5 m from the body of a helicopter, as shown in the cover illustration. The radar was controlled by a Xadar Electromagnetic Reflection Profiling System (Model 1316) carried inside the helicopter. The resulting pulses resemble that shown on the left in Figure 1, being similar in shape for both antennas though differing in frequency. Duration of the pulse from the 3102 antenna is about 6 ns, with a length in ice of about 1 m, while the 101C pulse is about half that long. The waveform on the right in Figure 1 is from a Sensors and Software Pulse Ekko system, operating at about 50-MHz center frequency. This antenna was suspended beneath the helicopter as shown in Figure 2 but was not used as much in our surveying because of its lack of shielding.

Operating at a pulse repetition frequency of about 50 kHz, the Xadar unit compiles the signals it receives into eight scans per second. Ultimately, the scans are recorded in analog form on cassette magnetic tape, having been sampled and converted to an audio-frequency facsimile before compilation. See the paper by Arcone and Delaney (1987) and its references for a more complete description of the system’s operation. Under ordinary airborne surveying conditions in previous endeavors, the records had to be digitized later. Handling and digitization of data represented a substantial task, given that a helicopter speed of 2 m/s and a digitization rate of 25,000 samples/s produces about 12.5 MB of

Figure 1. Examples of short-pulse radar waveforms for the GSSI model 101C resistively loaded dipole, vintage 1979, and the A³ Pulse Ekko low-frequency folded dipole, vintage 1985. The waveforms have been modified slightly by the filters of a control unit.
data per kilometer of survey. Digitally recording tape decks are now available commercially.

The GSSI antennas are resistively loaded dipoles that must sacrifice the advantage of high gain to produce a short pulse. The GSSI 3102 and 101C are geometrically similar and both have a 3-dB beamwidth of 70° in both planes (Arcone et al. 1986). Arcone and Delaney (1987) present formulae and a table showing the area of ice bottom illuminated as a function of elevation, ice thickness and air speed. For 1 m of ice, the area can range from about 8.4 m² at a helicopter altitude of 1.5 m and air speed of 1.5 m/s, to about 192 m² at 9.0 m and 9.0 m/s. Thus, spatial resolution and waveform coherence can vary owing to changes in altitude and direction.

Results

In this section we provide a representative sampling of previously reported results, to provide an impression of the character, successes and problems in this sort of surveying. Figure 3 (Arcone 1988) shows a profile of the ice on Birch Lake, southeast of Fairbanks, Alaska, with indications of principal features. The horizontal bands of coherent clutter represent helicopter and internal system noise; fortunately, the altitude, signal strength and regularity of the clutter are all such that one can easily distinguish meaningful returns. This ice layer was quite level and uniform in character, thus the separation of lighter (weaker) top and darker (stronger) bottom echoes is approximately constant, while changes in altitude cause the appearance of a wavy surface. A ghostly multiple reflection within the ice is visible below the dark return from the ice bottom. Overall, this survey is marked by the coherence of the echoes. The presence of thin white lines within the main echoes, indicating amplitude zero crossings, means that successive traces show about the same pulse character and do

Figure 2. A 3 Pulse Ekko antenna suspended below a Bell Jet Ranger 206.
Figure 3. Profile of ice thickness over Birch Lake, Alaska, made with the model 3102 antenna. The vertical ice thickness scale applies only within the first two reflections. The decibel values given in parentheses refer to the round trip propagation loss for the water and ice multiple events at the points indicated. Portions of some events may not be visible in the reproduction because the figure was made light enough to show the signal zero crossings (thin white lines) in the stronger events.

Figure 4. Surveys of a cross section of the Tanana River in Alaska.
not suffer from a chaotic overlay of echoes from across the entire beamwidth. Returns are being received from a rather small area directly beneath the antenna.

Figure 4 (Arcone and Delaney 1987) shows two profiles of a cross section of the Tanana River in Alaska, both obtained using the 3102 antenna. In the ground survey in Figure 4a, the dark band at the top is antenna direct coupling superposed on the ice surface echo. These two features are widely separated in results from the airborne survey shown in Figure 4b. Figure 5 (Arcone and Delaney 1987) shows a comparison of ice thickness measurements from ground, airborne and drilling surveys of two river cross sections, including the one depicted in Figure 4 (top of Figure 5). Agreement is good, illustrating the fact that, at least under reasonably benign conditions, airborne surveying can replace or supplement more laborious measurement on the surface.

Some impression of the character of returns under other more general river ice conditions is provided by Figure 6, which shows part of a 3102 profile of the Yukon River in Alaska (Arcone and Delaney 1987). We note the clear delineation of the top and bottom of the ice on the right and left sides of the figure. In the center is a rubble channel, containing angular chunks of ice frozen in place, protruding on the order of 1 m with typical separations of about 3 m. Presumably, conditions on the bottom surface are similar. This produces a degeneration in coherence in both the upper and lower returns. A loss of coherence because of the rough upper surface is necessarily incorporated in the return from the bottom surface, even if the latter is smooth. In this case the character of the lower portion of the record in the rubble channel is typical of what we see in many surveys involving heterogeneous subsurface accretions. These may consist of watery agglomerations of small discoid or spicule “frazil” particles, as well as mixtures of other varied ice fragments (Ashton 1986). Such accumulations can represent a substantially greater presence of ice mass than one would infer from surface features. Heterogeneity of a jumbled ice deposit under the solid ice cover compounds the degeneration of coherence initiated at the rough upper surface. Subsurface ice accumulations like this have practical importance in that excess ice mass can contribute ultimately to ice jams, which are a major cause of flooding and damage to man-made structures.

From the point of view of radar surveying, frazil ice accumulations are problematical in that they may be high in water content, thereby causing severe attenuation. In addition, heterogeneities on the order of a significant fraction of the incident wavelength in water will cause substantial scattering. This means that, at the frequencies where we are operating, the scale of heterogeneities is frequently too large for us to swallow them.
up by treating the medium as an equivalent homogeneous mixture, tantamount to a continuous dielectric medium. They are also too small for us to resolve distinctly.

A higher frequency system is better suited for outlining the bottom of a solid ice layer, de-emphasizing a frazil accumulation below. Higher frequency response to the submerged watery material below is diminished because higher frequencies are scattered more effectively by the distinct pieces of ice; the destructive interference from randomly arrayed scatterers is greater for higher frequencies, and attenuation from dielectric dispersion in the water is greater. A lower frequency system will produce returns that survive scattering better and that suffer less attenuation in the water, thereby probing the submerged frazil more visibly. Overall, the character of the reflected signals from these and other submerged media is determined by the combination of radar system, flight speed, altitude, specific medium composition and temperature. Other things being equal, lower altitudes and slower air speeds illuminate a smaller area per scan, and thereby increase resolution and decrease incoherence. A narrower sampling of a rough domain allows fewer irregularities away from the area directly beneath the aircraft to contribute to the ultimate mix that the antenna receives. Arcone and Delaney (1987) illustrate the different quality of records obtained over irregular ice conditions, depending on the frequency range and manner in which the survey is conducted.

The attenuation of radio waves in water because of dielectric dispersion is a very significant factor in the sort of surveying of interest here. Figure 7 (Arcone 1988) shows the attenuation in decibels per meter in water at 0 and 20°C, as a function of frequency, assuming that the resistivity of the water is 100 Ω·m (a typical value for an only moderately polluted river). The loss
increases drastically with frequency above 100 MHz, and the greater loss at 0°C is due to the temperature sensitivity of the dielectric dispersion mechanisms. Dielectric dispersion is responsible for by far the major part of the loss as one approaches 1000 MHz in any case. This has important implications for the surveying of submerged ice accumulations as well as river and lake bottoms. Higher power, lower frequency antennas (less than 100 MHz) will maximize penetration, but will also mean an overall loss in resolution of ice, bottom features and submerged objects.

Figure 8 (Arcone 1988) shows the pleasingly sharp results of a river bottom survey near the Buckley Bridge in Hartford, Connecticut. A GSSI model 3307 “V” antenna was used with a nominal center frequency in air of about 250 MHz. These results were achieved by submerging the antenna in the water and dragging it with a small boat. Ignoring any frequency shift caused by interaction with the water, we estimate the center wavelength to be about 13 cm. This small wavelength contributes to the keen resolution of the pattern of scour holes as well as the detail in subsurface sedimentation. Unfortunately, such appealing quality in results is not reproduced in some recent airborne attempts using higher power, lower frequency antennas. An airborne attempt to profile the bottom of a shallow reservoir using a GSSI 3107 antenna (center frequency about 250 MHz) produced results dominated by unwanted reflections and multiples between helicopter, water and antenna (Arcone 1988). This occurred despite shielding around the transmit and receive antennas, which were suspended in a cargo net beneath the aircraft. Figure 9 illustrates schematically the various modes in which unwanted reflections can find their way into the results (Wills 1987) of this sort of survey.

**Sloping bottom effects and critical angle phenomena**

Another difficulty in airborne bathymetry lies in combined critical angle and sloping bottom effects. If one considers a planar sloping bottom (Fig. 10), a number of points emerge. For simplicity we assume that
Figure 8. Surface radar profile of scour holes on the upstream side of the Buckley Bridge (Hartford, Connecticut). Numbered vertical demarcations indicate the positions of the bridge piers. Note the sharp pulse defining the bottom and the amount of subsurface information.

Figure 9. Various modes by which clutter can be generated during helicopter surveying.
the bottom has a constant slope in the region surveyed and that the signal enters the water at point A. The bottom forms an angle $\alpha$ with the horizontal, and the "true" depth, i.e., along $AB$, is equal to some quantity $d$.

The response from point B will be deflected so that it returns to point $F$, striking the surface a distance $d \cdot \tan(2\alpha)$ away from the presumed observation point A. For a slope of 22.5° the displacement $AF$ of the return will be equal to $d$. The difficulty of catching this return will be greatly compounded by the very small critical angle for electromagnetic waves passing from water to air. For water between 0 and 20°C, the critical angle is about 6°. Because the angle of incidence at $F$ along the ray path $BF$ is $2\alpha$, a bottom slope of only 3° or greater will cause the return to vanish as a surface wave at $F$ and not reach an airborne antenna. For slopes of 6° or more, even the direct return along the path $CA$ will not be detectable from the air. To the extent that it is detectable (if the antenna at A is submerged, for example), it will correspond to an apparent depth of $d \cdot \cos(\alpha)$, which still provides a reasonably accurate estimate of $d$ for small angles. At point E another return will emerge perpendicular to the surface. It will be displaced $AE = d \cdot \sin(2\alpha)$, but it will be unaffected by critical angle problems. By measuring the distance $AE$ and timing the delay of the return, one could gain an estimate of both the true depth and the slope of the bottom. If the time required for the signal to travel the complete path $ADE$ is $\delta t$ and the electromagnetic wave speed in water is $v$, then

$$AD + DE = d + DE = \delta t \cdot v.$$  

At the same time,

$$AE^2 + DE^2 = d^2.$$  

Thus, knowing $AE$ and combining these two equations, one can determine $d$; from this one obtains $\alpha$ from the expression above for $AE$. While not out of the question, bistatic measurements of this sort are likely to be impractical.

Figure 11 shows an example in which river bottom topography was successfully recorded (Arcone and Calkins 1990). The section is approximately 120 m long.
and the maximum water depth is 2 m. The maximum slope in the record is approximately 7° near the 40-m mark, and here the bottom returns indeed fade out. (The ice cover focuses the beam slightly, thus slightly relieving the 6° criterion stated above.) Generally, the bottom slope is less than 6°. The hyperbolic return at 95 m is a near-surface sub-ice disturbance.

DECONVOLUTION AND THIN LAYERS

Interpretation of data from geophysical exploration systems is frequently based on a presumed pattern of response by a smoothly layered earth model to some impinging energy. We are relatively fortunate in our focus on cold regions in that laterally extensive layers of ice, water, frozen and unfrozen ground often provide a reasonable approximation of the models postulated. In what follows we sketch briefly some common knowledge on the response of an ideally smooth and piecewise homogeneous layered system to an incident wavelet. This illuminates the methods we have chosen to distinguish echoes from well-separated interfaces as well as those that we apply to more problematic cases of thin layers. Whether a layer is to be considered thin or not depends upon the combination of its thickness, the wave velocity within it and the duration of the incident pulse. If these factors come together such that reflection off the bottom of the layer and subsequent multiples interfere with the reflection from the top, then for discrimination purposes we consider the layer to be thin.

As long as we remain with impulse radar of the sort discussed here, it is difficult to make the thin layer problem go away. In some cases one can achieve improved resolution of a relatively thin surface ice layer by resorting to a higher frequency antenna, which produces a shorter pulse. However, many of the layers we wish to detect lie beneath or consist of media that attenuate higher frequencies severely, as discussed above. If we resort to lower frequencies to enhance penetration, we will end up with longer pulses, relative to which the layer of interest may be quite thin. In what follows we describe preliminary results of our attempts to develop signal processing techniques that enhance resolution by approximately an order of magnitude. They do this by distinguishing the different delay structures in different waveforms consisting of superposed wavelets.

Figure 12. Impulse response by the ice layer over a perfect reflector.
Figure 12 shows schematically the impulse response of an ice layer over a perfect reflector, assuming normal incidence. Any layer will produce a similar response pattern, with the signs and magnitudes of the sequential reflections \( R_i \) depending on the dielectric constant \( \varepsilon \) contrasts. In the notation below, we refer to the material above the subject layer as medium 1, that in the layer as medium 2, and that below as medium 3. In the example shown, we assume that the system is piecewise homogeneous, consisting of air \((\varepsilon_1 = 1)\) underlain by ice \((\varepsilon_2 = 3.2)\) over a perfect reflector. The dielectric constant of water at 0°C is 88, and the impulse response sequence for an air–ice–water system consists of a set of \( R_i \) with the same signs as shown but with less contrast between the magnitudes of the surface reflection \( R_0 \) and the succeeding \( R_1, R_2, \ldots \)

The complete echo for the system shown will consist of the convolution of an incident wavelet with the impulse response series illustrated. That is, the response \( x_t \) to the wavelet \( s_t \) will be

\[
x_t = R_0 s_t + R_1 s_{t-\tau} + R_2 s_{t-2\tau} + \ldots
\]

where the subscript \( t \) is a discrete time index and \( \tau \) is the length of time required (in the same units as \( t \)) for the wavelet to traverse the layer and return. For the most part, it is \( \tau \) that we are trying to determine. Knowing it, we can calculate the layer's thickness from \( \varepsilon_2 \), or vice versa. The response coefficients \( R_i \) can be calculated from reflection and transmission coefficients for the various interfaces assuming normal incidence

\[
G_{jm} = \frac{\eta_j - \eta_m}{\eta_j + \eta_m}
\]

\[
T_{jm} = \frac{2\eta_m}{\eta_j + \eta_m}
\]

where \( G_{jm} (T_{jm}) \) is the reflection (transmission) coefficient for a signal propagating from the material of layer \( j \) to layer \( m \), and \( \eta_j \) is the impedance of the \( j \) material. In these terms, the response coefficients are

\[
R_0 = G_{12} \quad R_1 = T_{12} G_{23} T_{21}
\]

\[
R_i = R_1 (G_{21} G_{23})^{i-1} \quad \text{for } i > 1.
\]

Under conditions common in our frequency range and survey environments, the reflection and transmission coefficients may be regarded as essentially independent of frequency. Thus, we may evaluate the coefficients \( R_i \) as far out into the sequence as we like, and then use eq 1 to construct the overall response \( x_t \) for our wavelet \( s_t \). An alternative and generally more informative approach is to construct a complex, frequency dependent reflection coefficient \( G_q \) for the entire layer system by using an impedance transformation (Cheng 1983).

\[
G_q(f,d) = \frac{Z - \eta_1}{Z + \eta_1}
\]

where we have replaced \( \eta_2 \) in eq 2 with a complex impedance, \( Z \), for the combined medium 2–medium 3 portion of the system. \( Z \) is given by

\[
Z(f,d) = \eta_2 \cos (kd) + i \eta_2 \sin (kd)
\]

where \( t \) denotes the square root of minus one, \( k \) is the wave number in medium 2, and \( d \) is the thickness of the layer (medium 2). The expression \( G_q \) may be used as a transfer function for the layer to construct \( x_t \)

\[
x_t = \int G_q(f,d) S(f) e^{2\pi ifd} df
\]

where \( S(f) \) is the Fourier transform of the wavelet. Thus, in theory we could take the transform of any return \( x_t \) and divide it by \( S(f) \) to obtain the transfer function, which in turn could be inverted to produce an impulse train such as appears in Figure 12.

Figure 13 displays \( G_q \) for two different systems, consisting of air (medium 1), ice (medium 2), and underlayers with \( \varepsilon_3 = 22 \) in one case (dashed line) and \( \varepsilon_3 = 352 \) in the other (solid line). The dielectric constant of water at 0°C is 88; thus, these plots show the effect of doubling or halving the square root of \( (\varepsilon_2/\varepsilon_3) \) relative to an air–ice–water system. Increasing the value of \( \varepsilon_3 \) raises the magnitude of reflections over all frequencies but has little effect on the overall shape of the magnitude and phase curves of the transfer function. In other words, the structure of the delays between different components of the total return is unchanged, and that is reflected in the similarity in pattern in the two transfer functions.

**Well-separated echoes**

To avoid problems in dividing the transform of a signal by \( S(f) \), which may have zeroes or areas of low magnitude, it is common practice to use digital inverse filters in the time domain. When the \( R_i \) series contains well-separated echoes, one can apply a filter designed to respond to the wavelet itself. The by now classical approach to design such filters is the Wiener least
To proceed, we excise a segment $\chi_\sigma$ from the original signal, where

$$\chi_\sigma = x_t, \ n \leq t \leq n + M, \ \sigma = t - n$$

where $n$ is any particular point in (discrete, denumerative) time and $M$ is the wavelet length. Thus, $\chi_\sigma$ contains a segment of $x_t$ beginning at $t = n$ and encompassing the next $M$ points. As the method proceeds, the point $n$ is located at all time points in succession. The transform of $\chi_\sigma$ is calculated, and we test it for similarity to the wavelet spectrum by examining the inverse of the function $U(f)$.

$$U(f) = X(f)/S(f)$$

where $X(f)$ is the transform of $\chi_\sigma$. This corresponds to a filtering operation in the time domain producing $u_\sigma$

$$u_\sigma = \int_{f_1}^{f_2} U(f)e^{2\pi i f \sigma} \, df$$

where $(f_1, f_2)$ is the frequency band chosen on the basis of $S(f)$. Because all the transforms treated here ultimately result from real sequences, their values for negative frequencies are simply complex conjugates of values for corresponding positive frequencies. Thus, in eq 8 we need only integrate between positive frequency limits and retain the real part of the result. If $\chi_\sigma$ consists of a wavelet, then $U(f)$ should approximate a constant with zero phase, and $u_\sigma$ should contain a spike at $\sigma = 0$.

To construct the filtered signal $y_n$ in the time domain, we only determine the value of $u_\sigma$ at $\sigma = 0$, to see the maximum response where a spike should be, and insert that value in $y$ at time point $n$.

$$y_n = u_0 = \text{Re} \sum_{f_1}^{f_2} U(f)$$

This means that the filtered signal is constructed simply by summing the transform values of small segments of $x_t$ as they are calculated at each frequency.

In practice we introduce one further modification. To compensate for the loss of optimality relative to the Wiener filter in our band-limited calculations, and to further suppress unwanted responses of the filtering operations, we also apply a time-dependent weighting factor. This factor can have a variety of forms, the primary requirement being that it equal one when (similarly scaled) $S(f)$ and $X(f)$ agree and that it decline rapidly when they do not. One successful weight used is

![Figure 13. Transfer function for a combination of media 2 and 3 as a function of wave number times layer thickness.](image)
where "avg" denotes an average over the frequency band \((f_1, f_2)\) of the cubed quantity in brackets, and the variables are defined by the expressions for the scaled spectra

\[ X(f) = A_x(f)e^{i\phi_x(f)} \]
\[ S(f) = A_s(f)e^{i\phi_s(f)} \]  

Thus, eq 9 becomes

\[ y_n = w_n u_0. \] (12)

One can sharpen or loosen the system by playing with the exponent in eq 10; a value of 3 seems to suppress unwanted responses thoroughly without being unduly sensitive to small differences between the spectra being compared.

This overall approach has worked well in preliminary tests. The upper curve in Figure 14 (O'Neill 1988) shows a portion of a record obtained over a river ice sheet with the 3102 antenna. The lower curve is \(y_n\), the result obtained with this filtering system. On the basis of the time separation between the spikes and an assumed ice dielectric constant of 3.2, the estimated ice thickness is about 48 cm, which is within about 10% of values measured in the field in the vicinity of the radar readings. The system appears to be quite robust in the presence of noise (O'Neill 1988). More extensive testing of this system on actual field records with representative selections of both common and challenging cases is warranted.

**Thin Ice Layers**

When the ice is thin relative to the length of the wavelet in ice, an altogether different approach must be employed. In thin ice returns, wavelets convolved with the \(R_i\) series overlap, and the results from a system designed to recognize separated wavelets will deteriorate. This case is of practical importance because a good deal of ice in seasonally cold areas either remains thin through the cold season or in any case is thin at least during the early part of the season. In such instances safety, scientific or engineering investigations may be frustrated when the thicknesses are too small to be distinguished with ordinary time domain filtering methods.

To test approaches to signal processing for thin ice, we carried out a laboratory program in which a thin sheet of ice was grown on top of a foil sheet. Its thickness was gradually increased by adding water to its surface. After each addition was frozen solid, we used
where all "real" echoes were obtained from the ice sheet in the lab. Example (a) at the bottom of the figure is a return from bare foil, and it provided the reference wavelet and the \( S(f) \) used in eq 6 to compute all synthetic return shapes. As the ice increases in thickness (bottom to the top of the figure), one sees \( R_0 \) and \( R_2 \) gradually move away from the larger central reflection \( R_1 \). (Note the magnitudes of the successive spikes in Figure 12.) Synthetic example (f) at the top corresponds to 24 cm of ice, and one sees that it is just about at the thickness where \( R_0 \), \( R_1 \), and \( R_2 \) become completely distinct. It is clear that the real and synthetic signals in the figure agree quite well, as they did at a variety of other thicknesses between 0 and about 17 cm.

Given any real signal to be processed, one performs a cross-correlation with each member of the synthetic library of possible matches. For each possible thickness \( d \), the correlation function \( c_t \) is examined, and the value of its peak noted. These peak values are assembled into a maximum correlation value function \( a(d) \), and the value of \( d \) for which \( a(d) \) itself is a maximum constitutes the best estimate of \( d \) for the real signal. This procedure assumes that all synthetic echoes have been scaled to have the same total energy, which is automatically the case when \( G_1(f,d) \) is constructed for an air-ice-foil system. Because material 3 is a perfect reflector, all incident energy must ultimately be reflected. This also assumes essentially one-dimensional plane wave propagation back and forth between the layer system and the antenna, an assumption that would be compromised by an uneven surface with higher antenna elevation.

Spelling out the above in equations, we first construct synthetic echoes \( g_t \) for each prospective \( d \)

\[
g_t(d) = \text{Re} \int \frac{G_t(f,d)}{f_1} S(f)e^{2\pi i f t} df \quad (13)
\]

We then compute the cross-correlation \( c_t \) of each of these with the signal to be analyzed, \( x_t \)

\[
c_t(d) = \sum_{j=\pm \infty} g_t(jd)x_j \quad (14)
\]

The maximum value of \( c_t \) is recorded in \( a(d) \) for each \( d \)

\[
a(d) = \max_{\text{over } t} \left\{ c_t(d) \right\} \quad (15)
\]

and the peak in \( a(d) \) should identify the \( d \) value for which an auto-correlation with zero time shift has been obtained.

Figure 16 shows \( a(d) \) for the two real returns shown
Figure 16. Correlation maxima function $a(d)$ for a 17.1-cm-thick ice sheet (dashed) and for a 7.3-cm-thick ice sheet (solid line).

in Figure 15 corresponding to 7.3 and 17.1 cm. In each case, the peak in $a(d)$ is in the right locale, although the maximum in the lower (7.3 cm) curve is barely above the next highest peak. Figure 17 shows the $d$ values for which $a(d)$ was a maximum, as a function of measured thickness. That is, the best estimates of ice thickness based on the processing system are plotted on the vertical axis, versus the actual $d$ values applying in each case. Perfect agreement would place all points on the 45° line. Clearly, the level of resolution is quite good. This represents an order of magnitude improvement over the level of estimation possible on the basis of the

Figure 17. Thin ice sheet thickness as determined from signal processing (peak locations in $a(d)$ function) versus actual thicknesses of laboratory ice sheets from which reflections were recorded.
raw data or obtainable from conventional time domain deconvolution. Further testing of the method on thin ice growing on a pond is reported elsewhere (O’Neill, in prep.). Results are equally good in both studies, with some degeneration of accuracy at greater thicknesses. This was traced to a spreading of the wavelet in the $R_1$ return; it is speculated that this is attributable to variation of frequency content with angle within the incident beam (O’Neill, in prep.).

Some signal processing exercises were also conducted to gain some feel for the robustness of the method. Figure 18 shows three synthetic returns for a hypothetical 17-cm ice sheet with air above and medium 3 underneath, having $\varepsilon_3 = 1, 12$ and 352. Thus, the cases cover an $\varepsilon_3/\varepsilon_2$ ratio both less than and greater than 1, and both less than and greater than what would occur in ice over water. Each curve is scaled relative to its peak amplitude, producing a relative rise in prominence of the center $R_1$ component in the progression from the bottom to the top of the figure. Signals at various intermediate values of $\varepsilon_3/\varepsilon_2$ were also constructed, and in all cases some artificial noise was thrown in, for a whiff of realism.

Each of these signals was then processed in the manner described above, using the library developed for the air–ice–foil system. Figure 19 shows the results of this exercise, indicating that the method correctly estimated 17 cm as the subject signal thickness whenever there is substantial contrast between $\varepsilon_2$ and $\varepsilon_3$, with less success where they are comparable ($\log 3.2 = 0.5$).

Going back to Figure 13, we note that the amplitude curve of the layer system transfer function for the air–ice–foil system would represent the limiting case for an increase in $\varepsilon_3$, and would be flat at a value of 1. The shape of the phase curve however is similar to those shown. Thus, by using prospective matching signals based on the foil system, we have wiped out all information on the amplitude response of the system. Nevertheless, despite substantial alteration of the phase characteristics of the test returns caused by variation of $\varepsilon_3$ (contrast phase patterns shown below when $\varepsilon_3 < \varepsilon_2$ with those in Fig. 13), this processing scheme appears to react adequately to phase information only. While we have shown success in constructing an algorithm to determine ice thickness based only on phase information (Riek et al. 1990), the method presented here should have the advantage that, for physical instances encountered in the field, both amplitude and phase information is present. Our deletion here of amplitude information suggests that the performance shown in

Figure 18. Synthetic echoes with added noise for 17-cm-thick ice layer when $\varepsilon_3 = 1.0$ (bottom), $\varepsilon_3 = 12$ (middle) and $\varepsilon_3 = 352$ (top).
Figure 19. Thickness calculated for various 17-cm layer systems with $\varepsilon_3$ varied. Log 3.2 = 0.5.

Figure 19 may be a worst case. The method warrants testing on field records from river ice surveys, including heterogeneous ice accumulations under solid ice covers. This will help determine whether the processing system can distinguish the ice sheet when dielectric properties of the sublayer are complex and variable.

**Thin layers of water**

In some of our river ice surveying during the season approaching ice breakup, bright surface reflections from wet ice prevented meaningful profiling of the ice cross section. In what follows we offer some perspective on the effects on radar returns of thin and not-so-thin water layers on top of ice. These considerations are intended to outline likely feasibility limits and guide interpretation of surveying data.

To begin, we consider a smooth water layer of thickness $d$ at 0°C. To isolate the effects of the water layer, we consider the underlayer of ice to be infinitely deep and, as before, we treat normal incidence on a horizontally extensive surface to keep the analysis in one dimension. The same expressions apply as in eq 4 and 5, except that the materials corresponding to media 2 and 3 are reversed. Additionally, the dielectric constant of the water (medium 2) is now considered to be complex, which also corresponds to complex values for $\eta_2$ and for the wave number $k (=k_2)$. The dielectric constant model includes the effects of dc conductivity ($\sigma$) and of dielectric relaxation, expressed through the relaxation time $\tau_r$ (about $1.8 \times 10^{-11}$ at 0°C).

$$\varepsilon = \varepsilon_\infty + \frac{\varepsilon_3 - \varepsilon_\infty}{1 + i \omega \tau_r} - \frac{\sigma}{\omega \varepsilon_0}$$

where $\omega = \text{angular frequency}$

$\varepsilon_3 = \text{static dielectric constant (88.3)}$

$\varepsilon_\infty = \text{high frequency dielectric constant (5.28)}$

$\varepsilon_0 = \text{permittivity of free space}$.

This formula assumes that time dependency implied by all phasor quantities is expressed through a positive exponent, i.e., $\exp(i \omega t)$. Using this to calculate quantities in eq 4 and 5 produces curves for $G_1$ (Fig. 20) similar in many respects to those for the ice layer over water shown in Figure 13.

If one assumes a lossless water layer then the amplitude spectra in Figures 13 and 20 have the same overall shape relative to $kd$. The phase spectra differ in that the zero delay upper surface return $R_0$ is dominant in the case of water over ice, with $R_1$ and trailing multiples interfering with $R_0$ so as to cause curvature in the pattern around the zero delay position. Despite differences, the phase curves in both lossless cases show similar periodicity in curvature patterns. This is reflected in the results shown in Figure 19; large variation in $\varepsilon_2/\varepsilon_3$ values from well below to well above 1.0 does not prevent the
correlation system from recognizing the underlying structure in quite different returns for a given \((kd)\).

The lossy curves in Figures 20 and 21 illustrate the difference between the water layer vs ice layer cases. While the lossy and lossless amplitude curves are similar in the low frequency range, at higher frequencies a diminution of returns from the water/ice interface causes the lossy curve to flatten out, converging on a value approximately equal to that for an infinitely deep water body. It is interesting to note that for relatively thin layers the reflectivity of the water layer is greater than the reflectivity of water alone over most frequencies.

While the dc conductivity plays some role in flattening out the response at higher frequencies (a value of \(5.0 \times 10^{-3}\) S/m was assumed), by far the greatest part of this effect is ascribable to the dielectric relaxation, as noted above. Figure 21 shows a similar effect on the curvature in the phase pattern. In principle this curvature could be analyzed in the same manner as has been done successfully to calculate thin ice thickness based on phase curves of the type in Figure 13 (Riek et al. 1990). However, we note that in the lossy case the multiples have diminishing effect with higher frequency such that the curve flattens out around \(\pm \pi\), ultimately remaining at \(\pi\).

**Figure 20.** Reflection coefficient of water layer on ice \((d = 3.5\ cm)\).

**Figure 21.** Phase of \(G_e\) versus frequency for water layer on ice \((d = 3.5\ cm)\).
Figure 22. Two-way transmission coefficient for layer of water on ice (d = 3.5 cm).

Related effects are evident in curves for transmission coefficients through the water layer. Starting from standard expressions for the wave fields in each medium (Cheng 1983), one can solve for an air to water to ice transmission coefficient equal to

\[ T_{13}^{\text{lossless}} = \frac{T_{12} T_{23} e^{-4kd}}{1 + G_{12} G_{23} e^{-2kd}} \]  

(17)

where as before it is understood that \( k = k_2 \). Like \( k \), all interface reflection and transmission coefficients in this expression are complex; at the same time, the dominant factors in reducing the magnitude of \( T_{13} \) are the exponential terms containing complex \( k \). Ultimately, to detect objects or interfaces below the upper surface of the ice, we must be concerned with additional effects on the signal from the return trip through the water layer, i.e., we must calculate the round trip coefficient \( T_{\text{ret}} \), where

\[ T_{\text{ret}} = T_{13} T_{31} = \frac{\eta_1}{\eta_3} T_{13}^2. \]  

(18)

Figure 22 shows the magnitude of \( T_{\text{ret}} \) over the same domain as in Figure 20, assuming a water layer 3.5 cm deep on top of the semi-infinite ice. Especially because values in the lossy case reflect two-way losses, the magnitudes of peak transmission values decline relatively rapidly with frequency (compare relative effects of lossy parameters in Fig. 20). In any real application losses would be even greater, given additional losses within the ice and at the interface or object to be detected below the ice. As a reference for comparison, we note that a transmission coefficient of 0.1 corresponds to about a 20-dB power loss.

In the lossless case, \( k \) is only a linear function of frequency and \( G_1 \) and \( T_{\text{ret}} \) are functions only of the real valued product \( kd \), for a given set of dielectric properties in all three strata. In this case those functions can be plotted more revealingly as functions of \( kd \) instead of frequency. Referring to Figure 13, one sees therefore that, for a fixed upper limit on frequency, decreasing the value of \( d \) restricts one to a smaller portion of the left side of the plot. Alternatively, plotting the functions versus frequency, one would stretch the graphs out horizontally by decreasing \( d \). Thus, for a small enough value of \( d \) and a fixed frequency limit, one would only see a portion of the first cycle in the graph. In the lossy case, complex values of the water’s dielectric constant have their greatest effect through the exponential factors, through the product \( kd \). That product ceases to be a linear function of frequency and in any case becomes complex, preventing the same simple scaling of the graph as was possible in the lossless case. Nevertheless, as illustrated in the figures, the approximate equivalence between the two cases holds for relatively small values of \( kd \) where lossy mechanisms have less effect. Over most of the first cycle in these graphs the value of \( kd \) may be represented by its real part.

Consideration of “wet” as opposed to inundated ice restricts one’s view to this first cycle, given the frequency characteristics of our antennas. Figure 23 shows curves for the magnitude of \( T_{\text{ret}} \) for water layer thicknesses of 3.5 and 0.35 mm. These were calculated for the same values of \( \sigma \) and \( \tau _r \) as before, producing values essentially indistinguishable from the lossless case. Also shown is the magnitude spectrum for a wavelet transmitted by our 101C antenna; superposition of this on Figures 21 and 22 is also informative. The two water
depths used in Figure 23 roughly bracket the “yes–no” threshold for detecting objects or interfaces below the surface of the ice with this radar or others operating over a similar band. The 3102 antenna spectrum is similar in shape to the one shown but would be shifted correspondingly to the left. Corresponding signals therefore would likely survive a round trip through the 3.5-mm layer, but could not be used for much thicker layers. It is interesting to note that the inability of our system to surmount wet or inundated ice transmission loss is not due simply to the high reflectivity of the upper surface of the water or to the lossiness of the water layer. Rather, the two-interface interference effects raise the reflection coefficient above that of water alone over the first cycle (Fig. 20) and correspondingly decrease the value of $T_{ret}$ drastically.

If we are not concerned about probing the ice but only characterizing the water layer, we still encounter considerable constraints. With our two antennas and their wavelet durations of about 4 and 6 ns, one requires a water depth greater than about 6 and 9.5 cm, respectively, for complete separation of returns from the top and bottom of the layer. The results shown in this section and Figure 7 suggest that this is a dubious prospect for processing, given that only the lowest frequency portion of the wavelet at best would survive two-way transmission through layers greater than these thicknesses. For processing returns to detect thin layers, one requires a portion of the $G_1$ vs $(kd)_2$ curve that corresponds to at least a significant portion of the first cycle (see Fig. 20). Under ideal conditions using the phase processing algorithm (Riek et al. 1990), it might be possible to succeed with information over only about the first quarter cycle, but realistically at least a half cycle would be required for reliable results in practice. If we require information at $k_2$ values up to a limit corresponding to about 1.5 GHz, then each cycle in Figure 20 corresponds to an increment in the thickness $d$ of about 1 cm; meaning, a thickness of 2 cm would allow us to perceive the first two cycles, 3 cm the first three, etc. To view at least half the first cycle we need a water depth of at least 5 mm. Inspection of Figure 20 for 3.5 cm and our examination of similar plots not shown suggest that processing with thin layer techniques is likely to be unreliable for water layer depths much above about 4 cm because of fading of the $R_1$ response.

To sum up, these graphs suggest the following observations.

1. For inundated ice (water depth on the order of centimeters), the two-way transmission losses above about 500 MHz are likely to foil attempts at detecting features below the ice surface. A sufficiently broadband system might still furnish some information on the depth of the water layer itself by revealing the location of the first two peaks (troughs) in the magnitude of $T_{ret}$ ($G_1$), to the extent that those were detectable for depths up to several centimeters.

2. For merely wet ice, the ice could readily be probed with our radars for water depths less than about 4 mm. Even for the 3102 antenna, Figure 22 shows that losses above the center frequency would be greater than about 10–20 dB for a water layer exceeding about 4 mm.
3. For estimating the depth of a water layer on ice, the considerations above suggest that, for our radar systems, processing by standard methods assuming complete separation of interface echoes is unlikely to succeed at depths required for complete separation to occur. For shallower water layers our various thin layer algorithms may succeed but are likely to be restricted to layer thicknesses of at least about 0.5 cm and not greater than that.

OBSERVATIONS

Airborne surveying is most successful in terms of both data collection and interpretation when smooth, extensive horizontal layers are present, with low electrical conductivity and reasonably (piecewise) homogeneous permittivity. Exposed rock or dry soil strata and clean ice layers are ideal. Uncluttered upper ground or ice surface with low reflectivity also helps. Superior results are often obtained when antennas are shielded and when low altitude and air speed can be maintained.

Both established and developing signal processing techniques should be able to help interpret records from cleanly layered ice (and perhaps other) strata, even for quite thin layers. Wet and rough surfaces and heterogeneous ice–water mixtures still pose many interesting challenges. A new direction is represented by the use of pulse code modulation, such as has been developed at CRREL (Wills 1987). In this promising approach, one cross-correlates two transmitted complementary codes with stored replicas and then adds the two correlations to produce a highly compressed pulse, giving enhanced penetration at low power.

Ideally, satellite positioning systems should be integrated into the enhanced processing and display capabilities. This should be done so that flight paths and positions within them can be identified accurately during both real time and subsequent examination of data. At present, accurate positioning of airborne radar profiles is a serious problem.

More agreeable designs for higher power, lower frequency short-pulse antennas are required for situations in which penetration is of greater importance. Preferably, these systems should feature center frequencies in the 50-MHz range, and should weigh no more than about 50 kg. Substantial weight can be accommodated by a sling arrangement beneath a helicopter. However, extensive shielding and flight safety considerations limit one’s options, especially when high power work requires separate transmit and receive antennas.

Extensive shielding in airborne work might be avoided by the use of filtering and polarization techniques. The effectiveness of filtering strategies is likely to be compromised by the variety of waveforms in reflections from the helicopter, as well as by distortions of the primary wavelet itself by the aircraft. In cross polarization techniques one receives at a polarization orthogonal to that of the transmitted signal. Because smooth, flat surfaces do not rotate the polarization on reflection, this would diminish reflections from smooth surfaces and allow greater attention to rough surfaces, volume inhomogeneities and reflective three-dimensional objects. Such schemes must await serious improvements in the signal-to-noise ratio of short-pulse radar systems, which are now limited by sampling noise in the receiver. Circular polarization would enhance detection of surfaces. The rotational sense of the transmitted wave would be reversed by the air/water interface, but not by the water/bottom interface, thus allowing these to be distinguished by alternative receive modes.

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An overview is presented of recent activities and results in the use of commercially available short-pulse UHF radar for surveying ice conditions on freshwater bodies. Improvements in radar systems have made it possible to increase ice thickness resolution by as much as one third relative to that in past attempts, and some new signal processing approaches shown here may offer an order of magnitude improvement. Results from airborne surveying are shown in which the varieties of ice character are reflected. Given the lack of ground coupling, one can rely upon a reasonably well-defined wavelet structure for enhanced signal processing and interpretation possibilities. An algorithm is presented that locates returns from interfaces in the presence of noise for a non-minimum delay wavelet. The method performs a simple inversion in the frequency domain, enhanced by a time dependent weight designed to recognize the shape of the wavelet amplitude and phase spectra. Thin ice layers are resolved down to a few centimeters and are distinguished from an ice free condition by means of a matched filter system designed to recognize the interference pattern from parallel interfaces close to one another. The effects and constraints imposed by water layers on wet ice are discussed, as are general attenuation, sloping bottom, and critical angle effects in deeper water. In closing, observations on the problems and prospects of this sort of surveying are offered.