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The Accuracy of Water Table Elevation Estimates

Determined from Ground Penetrating Radar Data

by

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Abstract

In hydrogeologic investigations, water table elevations measured at sparsely distributed monitoring wells are used to interpolate the water table surface for a field site. Ground penetrating radar is one non-invasive technique which have been used to estimate water table elevations between monitoring wells. The accuracy of these estimates is effected by numerous sources of error, including surface elevations, GPR wave velocity, spatial velocity variations, capillary rise estimates and the time sampling interval. We have identified and quantified these uncertainties at two sites. When the sources of error are minimized, the uncertainty in water table elevation estimates from GPR data is ± 0.10 to 0.15 m. At a third site, we were unable to obtain a reflection from the water table on the GPR sections. However, a non-linear, inverse relationship between the depth to the brackish water table and the apparent conductivity values measured with a Geonics EM-31 system was found, showing that the unsaturated zone has a conductivity of approximately 6 mS/m. Therefore, poor GPR data quality at this site cannot be attributed to the conductivity of the unsaturated zone.

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Table of Contents

Approval Page	ii
Abstract	iii
Acknowledgements.....	iv
Table of Contents.....	v
List of Tables.....	viii
List of Figures.....	ix
List of Symbols and Abbreviations.....	xii
 Chapter 1: Introduction.....	 1
1.1 The Scientific Problem.....	1
1.2 Thesis Objectives.....	2
1.3 Datasets Used.....	3
1.4 Hardware and Software.....	4
 Chapter 2: Methodology and Background Literature.....	 5
2.1 Introduction.....	5
2.2 Electromagnetic Theory.....	7
2.2.1 Maxwell Equations.....	7
2.2.2 Reflection and Transmission Equations for GPR - the Fresnel Equations.....	14
2.3 GPR velocity analysis and depth determination.....	18
2.3.1 Velocity Analysis.....	18
2.3.2 Depth Determination.....	22
2.4 GPR and Hydrogeology.....	24
2.5 Summary.....	33
 Chapter 3: Sources of Error in Water Table Elevation Estimates.....	 35
3.1 Uncertainty Propagation.....	35
3.2 Sources of Uncertainty in Water Table Elevation Estimates from Ground Penetrating Radar.....	36
3.2.1 Surface Elevation Uncertainties.....	37
3.2.2 Uncertainty in the Capillary Fringe Depth from Ground Penetrating Radar Data.....	41
3.2.2.1 Uncertainty in the Picked Depth Values.....	42

3.2.2.2 Velocity Uncertainty on a Single Common Midpoint Gather.....	43
3.2.2.3 Velocity Uncertainty due to Heterogeneity and Spatially Sparse CMP locations.....	44
3.2.2.4 Vertical Resolution.....	54
3.2.2.5 Total Uncertainty in the Depth to the top of the Capillary Fringe.....	55
3.2.3 Capillary Fringe Height Estimation and Uncertainty.....	55
3.3 Sources of Uncertainty in Water Table Elevation Measurements from Monitoring Wells.....	56
3.4 Summary.....	57
Chapter 4: Data Processing of Ground Penetrating Radar.....	60
4.1 General Processing Techniques.....	60
4.2 Common Midpoint Gathers.....	62
4.3 Common Offset Lines.....	64
4.4 Deconvolution.....	68
4.5 Summary.....	69
Chapter 5: Gilby Waterflood, Alberta.....	70
5.1 Study Site.....	70
5.2 Dataset.....	70
5.3 Processing Characteristics.....	74
5.4 Results.....	76
5.4.1 Common Midpoint Gathers.....	76
5.4.2 Common Offset Lines.....	85
5.4.3 Apparent Conductivity Measurements.....	101
5.5 Summary.....	104
Chapter 6: Strathroy, Ontario.....	108
6.1 Study Site.....	108
6.2 Dataset.....	108
6.3 Processing Characteristics.....	111
6.4 Results.....	112
6.4.1 Common Midpoint Gathers.....	112
6.4.2 Common Offset Lines.....	113

6.5 Summary.....	122
Chapter 7: North Caicos Island, Turks and Caicos Islands, British West Indies.....	124
7.1 Study Site.....	124
7.2 Dataset.....	126
7.3 Processing Characteristics.....	128
7.4 Results.....	129
7.4.1 Common Midpoint Gathers.....	129
7.4.2 Common Offset Lines.....	131
7.4.2.1 Area A - Agricultural Farm.....	131
7.4.2.2 Area C - Swale and elliptical depressions.....	133
7.4.3 Apparent Conductivity Measurements.....	139
7.5 Summary.....	145
Chapter 8: Conclusions.....	147
References.....	152
Appendix A: GPR Data, Gilby Waterflood Alberta.....	158
Appendix B: GPR Data, Strathroy Ontario.....	179
Appendix C: GPR Data, North Caicos Island, Turks and Caicos Islands, British West Indies.....	184

List of Tables

2.1 Typical Electromagnetic Constants for Common Geologic Materials.....	8
2.2 Capillary Rise for Various Sediments.....	27
3.1 Uncertainties due to Surveying.....	39
3.2 Mean and Standard Deviation of Hydraulic Parameters for Various Textural Classes.....	47
5.1 Acquisition Information for the GPR Data Collected at the Gilby Waterflood Site.....	73
5.2 GPR Data Collection Parameters for the Gilby Waterflood Site.....	74
5.3 Processing Parameters for the Gilby Waterflood GPR Data.....	76
5.4 Summary of RMS Velocities to the Capillary Fringe.....	82
5.5 Comparison of Tie Points between the Main GPR Lines.....	96
5.6 Comparison of Tie Points between Line 1 and the Cross Lines.....	97
5.7 Summary of Uncertainties in the GPR Data.....	99
5.8 Comparison of Water Table Elevation Estimates from the 100 MHz GPR Sections with the Water Table Elevation Measurements from the Wells.....	100
5.9 Comparison of Water Table Elevation Estimates from the 200 MHz GPR Sections with the Water Table Elevation Measurements from the Wells.....	101
6.1 GPR Data Collection Parameters for the Strathroy Site.....	110
6.2 Processing Parameters for the Strathroy GPR Data.....	112
6.3 Direct Wave Velocity and RMS Velocity to the Capillary Fringe.....	113
6.4 Comparison of GPR Tie Points for Cross Lines at 4 and 29 m.....	117
6.5 Summary of Uncertainties.....	118
6.6 Comparison of Water Table Elevation Estimates from the GPR Sections with the Water Table Elevation Estimates from the Wells.....	119
7.1 Measurements from Various Wells on North Caicos Island.....	127
7.2 Apparent Conductivity Data Collected on North Caicos Island.....	141

List of Figures

2.1 Common Data Collection Modes for GPR data.....	5
2.2 Typical Common Offset GPR Section.....	6
2.3 Typical Common Midpoint Gathers.....	7
2.4 Field Vector Orientations for EM Waves Incident on a Plane.....	15
2.5 Semblance Velocity Analysis.....	22
2.6 Schematic Diagram of the GPR Response to the Water Table and Capillary Fringe.....	26
2.7 Volumetric Water Content/Pressure Head Relationship.....	29
2.8 Moisture Profile Changes during Infiltration.....	30
2.9 Capillary Fringe Model for Determining GPR Response.....	32
3.1 Errors Caused by Interpolation between Survey Stations.....	40
3.2 Pressure Head as a Function of Depth For Sand.....	49
3.3 RMS Velocity and Two-way Travel Time to the Capillary Fringe.....	51
3.4 Error in Depth due to Unaccounted Lateral Velocity Variations.....	53
3.5 Vertical Resolution.....	54
4.1 Wavelet Dispersion in GPR Data.....	61
4.2 Airwave and Direct Wave Picks for a CMP Gather.....	63
4.3 Processing Flow for Common Midpoint Gathers.....	64
4.4 Processing Flow for GPR Common Offset Lines.....	65
4.5 Effect of Constant Q Compensation on Wavelet Dispersion.....	66
4.6 Amplitude Spectra for GPR data Before and After Q Compensation.....	67
5.1 Site Map for the Gilby Waterflood Site.....	71
5.2 Water Table Elevations from the Gilby Waterflood Site.....	72
5.3 Surface Contour Map for the Gilby Waterflood Site.....	75
5.4 RMS Velocity Profiles for Lines 2, 3 and 4 (100 MHz data).....	77
5.5 RMS Velocity Profiles for Lines 2, 3 and 4 (200 MHz data).....	78
5.6 RMS Velocity Profiles for Line 1 and the Cross Lines (100 MHz data).....	79
5.7 RMS Velocity Profiles for Line 1 and the Cross Lines (200 MHz data).....	81
5.8 Spatial Velocity Variation.....	84
5.9 Time Migrated GPR Sections for Line 4.....	87
5.10 Depth Corrected Sections and Apparent Conductivity Data for Line 4.....	88

5.11 Depth Corrected Sections and Apparent Conductivity Data for Line 3.....	90
5.12 Depth Corrected Sections and Apparent Conductivity Data for Line 2.....	91
5.13 Depth Corrected Sections and Apparent Conductivity Data for Line 1.....	93
5.14 Depth Corrected Sections for the Cross Line at 20 m	95
5.15 Surface Contour Map of the Water Table Elevation Estimates from the 100 MHz GPR Data.....	102
5.16 Surface Contour Map of the Water Table Elevation Estimates from the 200 MHz GPR Data.....	103
6.1 Site Map for the Strathroy Field Site.....	109
6.2 Water Table Elevations from the Strathroy Site.....	111
6.3 RMS Velocity Profiles from Strathroy.....	114
6.4 Time and Depth GPR Sections from Line 10.....	115
6.5 Depth Corrected Sections for the Cross Lines at 4, 29 and 59 m.....	117
6.6 Surface Contour Map of the Water Table Elevation Estimates from the GPR Data...	120
7.1 North Caicos Island Field Sites.....	125
7.2 RMS Velocity Profiles from Areas A and C on North Caicos Island.....	130
7.3 Time and Depth Sections and Apparent Conductivities for Area A, Line 2.....	132
7.4 Time and Depth Sections and Apparent Conductivities for Area C, Lines 1 and 2.....	135
7.5 Cross-section Through a Depression.....	136
7.6 Time and Depth Sections and Apparent Conductivities for Area C, Line 3.....	137
7.7 Depth Section for Area C, Line 5.....	138
7.8 Time and Depth Sections and Apparent Conductivities for Area C, Line 6.....	140
7.9 Apparent Conductivity Measurements for Area E, Line 1.....	143
7.10 Water Table Depth Versus Apparent Conductivity from North Caicos Island.....	144
A.1 Depth and elevation corrected section from Gilby line 1, 100 MHz....	159
A.2 Depth and elevation corrected sections from Gilby line 1, 200 MHz..	160
A.3 Depth and elevation corrected sections from Gilby line 2.....	161
A.4 Depth and elevation corrected sections from Gilby line 3.....	162
A.5 Depth and elevation corrected sections from Gilby line 4	163
A.6 Depth and elevation corrected sections from Gilby cross line 10.....	164
A.7 Depth and elevation corrected sections from Gilby cross line 20.....	165
A.8 Depth and elevation corrected sections from Gilby cross line 30.....	166
A.9 Depth and elevation corrected sections from Gilby cross line 40.....	167

A.10 Depth and elevation corrected sections from Gilby cross line 50.....	168
A.11 Depth and elevation corrected sections from Gilby cross line 70.....	169
A.12 Depth and elevation corrected sections from Gilby cross line 85.....	170
A.13 Common midpoint gathers from Gilby line 1, 100 MHz.....	171
A.14 Common midpoint gathers from Gilby line 1, 200 MHz.....	172
A.15 Common midpoint gathers from Gilby line 2.....	173
A.16 Common midpoint gathers from Gilby line 3.....	174
A.17 Common midpoint gathers from Gilby line 4.....	175
A.18 Common midpoint gathers from Gilby cross lines 10, 20 and 30.....	176
A.19 Common midpoint gathers from Gilby cross lines 40, 50 and 70.....	177
A.20 Common midpoint gathers from Gilby cross line 85.....	178
B.1 Depth and elevation corrected sections from Strathroy lines 2 and 4.....	180
B.2 Depth and elevation corrected sections from Strathroy lines 6 and 8.....	181
B.3 Depth and elevation corrected sections from Strathroy line 10.....	182
B.4 Common midpoint gathers from Strathroy	183
C.1 Common Offset Sections from North Caicos Island, Area A.....	185
C.2 Common Midpoint Gathers from North Caicos Island, Area A.....	186
C.3 Common Offset Sections from North Caicos Island, Area B.....	187
C.4 Common Midpoint Gathers from North Caicos Island, Area B.....	188
C.5 Common Offset Section from North Caicos Island, Area C, lines 1 and 2.....	189
C.6 Common Offset Sections from North Caicos Island, Area C, lines 3 and 4.....	190
C.7 Common Offset Sections from North Caicos Island, Area C, lines 5 and 6.....	191
C.8 Common Midpoint Gathers from North Caicos Island, Area C.....	192
C.9 Common Offset Sections from North Caicos Island, Area D.....	193
C.10 Common Midpoint Gathers from North Caicos Island, Area D.....	194
C.11 Common Offset Sections from North Caicos Island, Area E.....	195
C.12 Common Midpoint Gathers from North Caicos Island, Area E.....	196

List of Symbols and Abbreviations

ϵ	permittivity
ϵ_0	permittivity of free space - 8.85×10^{-12} farad/m
ϵ_r	dielectric constant or relative permittivity
θ	volumetric water content
μ	magnetic permeability
μ_0	magnetic permeability of free space - 1.26×10^{-6} henry/m
μ_r	relative permeability
σ	electrical conductivity
ϕ	porosity
ψ	pressure head
ψ_s	saturated pressure head - equivalent to the capillary rise above the water table
B	magnetic induction
b	pore-size distribution index
CF_{elev}(GPR)	capillary fringe elevation estimated from ground penetrating radar
CMP	common midpoint
D	dielectric displacement
d_{cap}	depth to the top of the capillary fringe
E	electric field intensity
E	Easting (m)
H	magnetic field intensity
h_{cap}	height of the capillary rise above the water table
J	current density
$\bar{x}$	mean of x
N	Northing (m)
NMO	normal moveout
q	charge density
t_{cap}	two-way travel time to the top of the capillary fringe
v_{direct}	direct wave velocity
$v_{H\ cap}$	harmonic mean of the interval velocities above the capillary fringe reflection
$v_{RMS\ N}$	RMS velocity to layer N
WTD	water table depth below a datum measured in a well
WT_{elev}(GPR)	water table elevation estimated from ground penetrating radar
WT_{elev}(well)	water table elevation measured from a monitoring well
Z	surface elevation

1. Introduction

1.1 The Scientific Problem

In hydrogeologic investigations, many different methods are used to determine water table elevations. These methods often include contouring water table elevation data from sparsely distributed monitoring wells or piezometers to give a water table surface. The result is a smoothed surface with an unknown deviation from the actual water table away from monitoring well locations. An inexpensive, non-intrusive method to estimate water table elevations between monitoring wells would be very useful. Ground penetrating radar (GPR) is one possible geophysical method which could provide water table elevations.

GPR uses electromagnetic waves in the MHz to GHz range to investigate the electromagnetic properties of the shallow subsurface. In the last decade its use has become widespread, with applications in numerous fields including sedimentology, stratigraphy, geomorphology, archeology, contaminant hydrogeology, forensics, and engineering geophysics. For comprehensive overviews of GPR applications, see Davis and Annan (1989), conference proceedings from GPR'94, GPR'96, and GPR'98, conference proceedings from the Symposium on the Application of Geophysics to Engineering and Environmental Problems (SAGEEP) or one of two special GPR issues from the Journal of Applied Geophysics (Volume 33 and 40). Radar waves are transmitted through the subsurface, reflected at interfaces caused by changes in the dielectric constant, and are attenuated by high conductivity materials. GPR data is interpreted by measuring the time it takes for reflections to return from the subsurface. If the velocity of the GPR wave in the subsurface is also known, the time can be converted to depth below the surface.

In low conductivity materials, the volumetric water content of a material is the major contributor to the dielectric constant (Topp et al. 1980). At or near the water table, there is a change from unsaturated to saturated conditions in the subsurface. This change often

results in a strong reflection on ground penetrating radar data. Under the right conditions, ground penetrating radar can and has been used to map water table elevations. Extensive water table mapping using ground penetrating radar data has been completed by Johnson (1992), Iivari and Doolittle (1994), and van Overmeeren (1994). Beres and Haeni (1991) and van Overmeeren (1998) have used GPR in hydrogeologic investigations, identifying the water table as well as examining the radar patterns which might be produced in various sedimentary environments. Harari (1996) used GPR to investigate the internal structure of sand dunes. Harari identified the water table and showed evidence of wetting fronts and preferential flow paths which are controlled by the stratigraphic features. The question we will address in this thesis is the accuracy of water table elevation estimation from ground penetrating radar data.

In the preceding discussion, we mentioned the need for low conductivity materials in the subsurface for good GPR results. In the presence of higher conductivity materials in the subsurface, ground penetrating radar data quality tends to decrease. The magnitude of the decrease depends on the conductivity and dielectric constant of the subsurface, as well as the frequency of the radar waves. At the various field sites discussed in this thesis, high conductivity materials such as silts, clays and saline water occasionally cause a decrease in GPR data quality. A Ground Conductivity Meter was used to quantify any conductivity changes along the ground penetrating radar lines.

1.2 Thesis Objectives

This thesis investigates the accuracy with which the water table elevations can be determined with GPR. In order to determine the accuracy of the water table elevations, the following steps have been taken. We have identified and quantified the sources of error in water table elevation estimates from ground penetrating radar data. Where possible, we have collected data in such a way as to minimize the effect of these uncertainties. The GPR data sets were processed to best resolve the water table reflection and create the best possible depth

sections. Finally, the water table elevation estimates from the ground penetrating radar sections have been compared to elevation values from nearby wells. A secondary objective was to determine the comparability between apparent conductivity data collected with a Geonics EM-31 device and ground penetrating radar data quality.

1.3 Datasets Used

Three ground penetrating radar datasets were collected between June and November 1997. In June 1997 the Gilby Waterflood dataset was collected. It includes more than 500 m of radar lines collected in two frequencies at a station spacing of 0.25 m. Common midpoint gathers were collected approximately every 10 m along these lines to provide subsurface radar velocity information. Four monitoring wells were sampled throughout data collection to provide control data for the water table elevation at the site. These wells also provide general geologic data on the site. The nearby river was monitored qualitatively for water level changes as it is the major control on the local unconfined aquifer. To provide elevation data, the GPR lines were surveyed at 2 to 2.5 m station spacings. Finally, conductivity data was collected using an electromagnetic induction system in June 1999, along the approximate position of the original radar lines.

A dataset was collected on North Caicos Island, Turks and Caicos Islands, British West Indies in July 1997. Five sites on the island were chosen to test the method. Overall, three different frequencies were used to collect over 1000 m of radar lines at 0.25 m station intervals. Thirty-nine common midpoint gathers were collected where appropriate to provide radar velocity data, and surveying was completed in three of the five sites. Where possible, geologic data and water table elevations were collected, as well as water conductivity and temperature data. Electromagnetic induction data was collected along each of the radar lines to give an indication of the conductivity of the near surface.

In December 1997, a final dataset was collected near Strathroy, Ontario. This dataset was

collected in conjunction with a feasibility study on using GPR to find preferred infiltration paths in a field used for nitrate testing. Five parallel, 70 m long lines were collected using 200 MHz central frequency antennas with a station spacing of 0.10 m. Three cross lines were collected, along with three CMPs. Water table elevation measurements from nearby sampling wells were collected within days of the GPR data collection. The GPR line elevations were not surveyed, however kriging was used to interpolate elevation information from survey control points to the GPR lines. An electromagnetic survey was conducted on a different field in the same area, providing some indication of the conductivity of the soil.

1.4 Hardware and Software

The ground penetrating radar data used in this thesis was collected using a Sensors and Software Limited PulseEKKO IV radar system with 50, 100, and 200 MHz central frequency antennas. A Geonics Limited EM-31 Ground Conductivity Metre was used to collect the conductivity data at all three sites. Water table depths were determined using a Solinst Water Level Metre at the Gilby and North Caicos Island research sites. Surveying of the radar lines was accomplished using a WILD T1 theodolite and D14 distomat system with a single prism. Advanced data processing of the GPR data was performed in Advance Geophysical Corporation's PROMAX 6.2 seismic data processing package. Additional processing and data analysis were completed using Matlab 5 from Mathworks Incorporated.

2 Methodology and Background Literature

2.1 Introduction

Ground penetrating radar has become increasingly popular as a non-invasive technique for high resolution geophysical mapping of the shallow subsurface. In this technique, a short pulse of high frequency (10 to 1000 MHz) electromagnetic energy is transmitted into the ground. The electromagnetic wave is reflected, refracted and attenuated depending on the electromagnetic properties of the subsurface (Davis and Annan, 1989). A receiver amplifies and digitizes the arriving signal, which is then stored as a function of the time elapsed since the electromagnetic pulse was transmitted. Similarities between GPR and reflection seismic in terms of general wave propagation, data collection, and storage allow many of the processing techniques designed for seismic reflection data to be applied to GPR data.

GPR data is collected using a pair of identical antennae for the transmitter and receiver. Typically, the data is collected in common-offset profiles (fig. 2.1A) where the transmitter and receiver separation is fixed and they are moved across the area of interest in regular

Figure 2.1 Common data collection modes for GPR. **A** Common-offset profiling. The transmitter and receiver are kept at a fixed offset and moved along the line of interest in regular steps. **B** Common midpoint gather. The transmitter and receiver are centred equidistant from a common surface point and moved apart at a constant step size.

Figure 2.2 Typical common offset section. The airwave, direct wave and capillary fringe/water table reflector have been identified. (Data from the Gilby Waterflood site, cross line 85. 100 MHz central frequency antenna, 1.0 m antenna separation, 0.25 m station spacing.)

steps. To improve the signal-to-noise ratio, the data is usually stacked between 32 and 128 times (Sensors and Software Inc., 1993) at each station. Common-offset profiling results in a cross-section of reflections from the electrical impedance contrasts in the subsurface with the vertical axis as two way travel time (fig. 2.2). In order to obtain depth information, the velocity of the radar wave as it travels through the subsurface must be known. Velocity information cannot be determined from a common offset profile. A second data collection method, the common-midpoint (CMP) gather, is needed to extract radar velocity information. Common-midpoint data is collected by centring the transmitting and receiving antennae equidistant from a common surface point and moving them apart at a constant step size (fig. 2.1B). Figure 2.3 is an example of three CMP gathers collected every 10 m along the same line as the common-offset profile in fig. 2.2. In both figures, the first and second arrivals are the coupled air and ground waves. On the CMP gathers the inverse of the slope of the ground wave can be used to determine the near-surface wave velocity. Hyperbolic events below the air and ground waves correspond to changes in the electrical impedance in the subsurface.

Figure 2.3 Typical common midpoint gathers. The airwave, direct wave and capillary fringe/ water table reflector have been identified. (Data from the Gilby Waterflood site, cross line 85, CMPs @ 11, 21, and 31 m. 100 MHz central frequency antenna, 0.10 m station spacing.)

The hyperbolic reflectors in fig. 2.3 can be used to determine wave propagation velocities at depth. The airwave travels at the speed of light (0.3 m/ns) while electromagnetic waves in the ground can travel at velocities ranging from 0.03 to 0.15 m/ns. Table 2.1 provides propagation velocities, dielectric constants, conductivities and attenuation constants for common geological materials (Davis and Annan, 1989). The theory and relationships between these values will be discussed in Section 2.2.

2.2 Electromagnetic Theory

2.2.1 Maxwell Equations

As in all geophysical techniques based on electromagnetic theory, the theory behind ground penetrating radar begins with the Maxwell equations,

Table 2.1 Typical dielectric constant (ϵ_r), electrical conductivity (σ), velocity (v), and attenuation (α) observed in common geological materials at 100 MHz (Davis and Annan, 1989).

Material	ϵ_r	σ (mS/m)	v (m/ns)	α (dB/m)
Air	1	0	0.30	0
Distilled water	80	0.01	0.033	2×10^{-3}
Fresh water	80	0.5	0.033	0.1
Sea water	80	3×10^4	0.01	10^3
Dry sand	3 - 5	0.01	0.15	0.01
Saturated sand	20 - 30	0.1 - 1.0	0.06	0.03 - 0.3
Limestone	4 - 8	0.5 - 2.0	0.12	0.4 - 1
Shales	5 - 15	1 - 100	0.09	1 - 100
Silts	5 - 30	1 - 100	0.07	1 - 100
Clays	5 - 40	2 - 1000	0.06	1 - 300
Granites	4 - 6	0.01 - 1	0.13	0.01 - 1
Dry salt	5 - 6	0.01 - 1	0.13	0.01 - 1
Ice	0.01	0.01	0.16	0.01

$$\nabla \times \mathbf{E} = - \frac{\partial \mathbf{B}}{\partial t} \quad (2.1)$$

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t} \quad (2.2)$$

$$\nabla \cdot \mathbf{D} = q \quad (2.3)$$

$$\nabla \cdot \mathbf{B} = 0 \quad (2.4)$$

where $\mathbf{E}$ is the electric field intensity, $\mathbf{B}$ is the magnetic induction, $\mathbf{H}$ is the magnetic field intensity, $\mathbf{J}$ is the current density, $\mathbf{D}$ is the dielectric displacement and q is the charge density. In order to use these equations meaningfully, we must consider the constitutive relations, which relate the fields to the physical properties of the medium.

$$\mathbf{D} = \epsilon \cdot \mathbf{E} \quad (2.5)$$

$$\mathbf{B} = \mu \cdot \mathbf{H} \quad (2.6)$$

$$\mathbf{J} = \sigma \cdot \mathbf{E} \quad (2.7)$$

In the above equations, the assumption has been made that dielectric permittivity ϵ , magnetic permeability μ , and electrical conductivity σ are independent of the electric and magnetic field strengths and directions. While this assumption is rarely completely valid, it is necessary in order to bring the electromagnetic theory to a manageable level. Annan and Daniels (1998) state that “GPR has yet to be able to discern the complexity” of field dependent ϵ , μ , and σ “except in a few cases”.

Substituting the constitutive equations (eq. (2.5) to (2.7)) into the Maxwell equations (eq. (2.1) to (2.4)) gives

$$\nabla \times \mathbf{E} = -\mu \frac{\partial \mathbf{H}}{\partial t} \quad (2.8)$$

$$\nabla \times \mathbf{H} = \sigma \mathbf{E} + \epsilon \frac{\partial \mathbf{E}}{\partial t} \quad (2.9)$$

$$\nabla \cdot \mathbf{D} = \epsilon \nabla \cdot \mathbf{E} = q \quad (2.10)$$

$$\nabla \cdot \mathbf{B} = \mu \nabla \cdot \mathbf{H} = 0 \quad (2.11)$$

From these equations, the two symmetrical wave equations which describe the propagation of electromagnetic fields through a medium can be developed. Beginning with the electric field equation, taking the curl of both sides of eq. (2.8) gives

$$\nabla \times \nabla \times \mathbf{E} = \nabla \times \left(-\mu \frac{\partial \mathbf{H}}{\partial t} \right) = -\mu \frac{\partial}{\partial t} (\nabla \times \mathbf{H}) \quad (2.12)$$

Equation (2.9) can be substituted into eq. (2.12) to give

$$\nabla \times \nabla \times \mathbf{E} = -\mu \sigma \frac{\partial \mathbf{E}}{\partial t} - \mu \epsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad (2.13)$$

Using the vector identity

$$\nabla \times \nabla \times \mathbf{A} = \nabla (\nabla \cdot \mathbf{A}) - \nabla^2 \mathbf{A} \quad (2.14)$$

gives

$$\nabla (\nabla \cdot \mathbf{E}) - \nabla^2 \mathbf{E} = -\mu \sigma \frac{\partial \mathbf{E}}{\partial t} - \mu \epsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad (2.15)$$

In a source free region, $q = 0$, causing eq. (2.10) to equal zero. Consequently, the first term in eq. (2.15) is zero.

$$\nabla^2 \mathbf{E} = \mu\sigma \frac{\partial \mathbf{E}}{\partial t} + \mu\epsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad (2.16)$$

A similar derivation results in the magnetic field equation

$$\nabla^2 \mathbf{H} = \mu\sigma \frac{\partial \mathbf{H}}{\partial t} + \mu\epsilon \frac{\partial^2 \mathbf{H}}{\partial t^2} \quad (2.17)$$

Equations (2.16) and (2.17) represent the wave equations for electromagnetic fields. These equations can be expressed in the frequency domain by substituting

$$\begin{aligned} \mathbf{E} &\rightarrow e^{i\omega t} \mathbf{E} \\ \mathbf{H} &\rightarrow e^{i\omega t} \mathbf{H} \end{aligned} \quad (2.18)$$

into eq. (2.16) and (2.17). Differentiating the result gives

$$\nabla^2 \mathbf{E} + (\epsilon\mu\omega^2 - i\mu\sigma\omega) \mathbf{E} = \nabla^2 \mathbf{E} + \gamma^2 \mathbf{E} = 0 \quad (2.19)$$

$$\nabla^2 \mathbf{H} + (\epsilon\mu\omega^2 - i\mu\sigma\omega) \mathbf{H} = \nabla^2 \mathbf{H} + \gamma^2 \mathbf{H} = 0 \quad (2.20)$$

for the frequency domain electric and magnetic fields. The propagation constant, γ , is defined as

$$\gamma^2 = \epsilon\mu\omega^2 - i\mu\sigma\omega = \epsilon\mu\omega^2 \left(1 - i \frac{\sigma}{\epsilon\omega} \right) \quad (2.21)$$

The propagation constant is complex and is often written as

$$\gamma = \beta - i\alpha \quad (2.22)$$

where α is the attenuation constant and β is the phase constant. The result of separating γ into its real and imaginary components gives

$$\alpha = \sqrt{\omega^2 \mu \epsilon \left(\frac{(1+D^2)^{1/2} - 1}{2} \right)} \quad (2.23)$$

$$\beta = \sqrt{\omega^2 \mu \epsilon \left(\frac{(1+D^2)^{1/2} + 1}{2} \right)} \quad (2.24)$$

D is the dissipation constant and $\tan l$ is the loss tangent. As D increases, dissipation effects become more important to field propagation in the medium.

$$D = \tan l = \frac{\sigma}{\omega \epsilon} \quad (2.25)$$

Equations (2.23) and (2.24) can be simplified further by converting ϵ and μ to their relative values using $\epsilon = \epsilon_r \epsilon_o$ and $\mu = \mu_r \mu_o$ where r subscripts represent relative values and o subscripts represent free space values. As well, the speed of light, c , can be written in terms of ϵ_o and μ_o ($c = (\epsilon_o \mu_o)^{-1/2}$).

$$\alpha = \frac{\omega}{c} \sqrt{\frac{\mu_r \epsilon_r}{2} \left((1+D^2)^{1/2} - 1 \right)} \quad (2.26)$$

$$\beta = \frac{\omega}{c} \sqrt{\frac{\mu_r \epsilon_r}{2} \left((1+D^2)^{1/2} + 1 \right)} \quad (2.27)$$

Typically, $\mu_r \approx 1$ except when metallic objects or minerals are present in abundance (Telford et al., 1990).

For electromagnetic fields propagating in the z direction, the solutions to eq. (2.18) and (2.19) in terms of α and β are

$$E = E_o e^{-i\gamma z} e^{i\omega t} = E_o e^{-\alpha z} e^{i(\omega t - \beta z)} \quad (2.28)$$

$$H = H_o e^{-i\gamma z} e^{i\omega t} = H_o e^{-\alpha z} e^{i(\omega t - \beta z)} \quad (2.29)$$

These are the expressions for a damped plane wave which propagates through the medium with a phase velocity of

$$v = \frac{\omega}{\beta} \quad (2.30)$$

In a good conductor, displacement currents are negligible compared to conduction currents. The imaginary part of γ^2 dominates the real part and the electromagnetic fields do not propagate as waves. This is the case which is assumed in induction electromagnetic systems such as the Geonics EM-31.

In low conductivity materials with high frequency EM fields, the displacement current dominates the conduction current. The electromagnetic fields propagate as waves. In this case the dissipation constant D is very small.

$$D = \tan \delta = \frac{\sigma}{\omega \epsilon} \ll 1 \quad (2.31)$$

It is under these conditions that ground penetrating radar is most successful. When eq. (2.31) is true the attenuation and phase constants can be simplified to

$$\alpha = \frac{\sigma}{2} \sqrt{\frac{\mu}{\epsilon}} = \frac{1}{2c\epsilon_o} \frac{\sigma}{\sqrt{\epsilon_r}} \quad (2.32)$$

$$\beta = \frac{\omega}{c} \sqrt{\mu_r \epsilon_r} \quad (2.33)$$

The phase velocity of the electromagnetic waves in a low conductivity medium can be simplified to

$$v = \frac{c}{\sqrt{\mu_r \epsilon_r}} \approx \frac{c}{\sqrt{\epsilon_r}} \quad \mu_r \approx 1 \quad (2.34)$$

Equations (2.32), (2.33) and (2.34) give the low loss approximations to the attenuation constant, phase constant and phase velocity. It is interesting to note that both α and v are no longer dependent on the frequency of the EM wave. This is contrary to observations, which show that attenuation of the radar signal increases with increasing frequency. As well, we have assumed throughout this derivation that ϵ , μ , and σ are frequency, space and time independent. While this is not entirely true, Davis and Annan (1989) show that the frequency dependence of α and v is small in the frequency range used in GPR as long as we remain within the low loss conditions of eq. (2.31).

2.2.2 Reflection and Transmission Equations for GPR - the Fresnel Equations

The reflection of electromagnetic waves at changes in electrical impedance is governed by the Fresnel equations. While a complete discussion of these equations will not be given here, interested persons are referred to any text on electrodynamic theory, for example, Jackson, (1975). The Fresnel equations depend on the orientation of the electric field compared to the plane of incidence. Figure 2.4 shows the physical relationships between field vector orientations, the interface between media 1 and 2, and the plane of incidence. The electrical impedance, Z , of a medium is defined as

$$Z = \frac{E}{H} = \frac{\omega \mu}{\gamma} = \sqrt{\frac{\mu}{\epsilon \left(1 - i \frac{\sigma}{\epsilon \omega} \right)}} \quad (2.35)$$

Figure 2.4 Field vector orientations for **A** an incident wave with E perpendicular to the plane of incidence; **B** an incident wave with E parallel to the plane of incidence.

When the electric field vector is normal to the plane of incidence, the Fresnel equations for

reflection and transmission are

$$\left(\frac{E_{Rm}}{E_{Im}} \right)_\perp = \frac{Z_2 \cos(\theta_I) - Z_1 \cos(\theta_T)}{Z_2 \cos(\theta_I) + Z_1 \cos(\theta_T)} \quad (2.36)$$

$$\left(\frac{E_{Tm}}{E_{Im}} \right)_\perp = \frac{2Z_2 \cos(\theta_I)}{Z_2 \cos(\theta_I) + Z_1 \cos(\theta_T)} \quad (2.37)$$

In the equations above, E_{Im} , E_{Rm} , and E_{Tm} are the magnitudes of the incident, reflected and transmitted electric field, Z_1 and Z_2 are the electrical impedances of the first and second media, and θ_I , and θ_T are the incident and transmitted angles. This case is depicted in fig. 2.4A. When the GPR antennae are positioned with their long axes perpendicular to the survey line, the electric field vector is normal to the plane of incidence (Turner, 1992).

When the electric field vector is parallel to the plane of incidence, the Fresnel equations for reflection and transmission are

$$\left(\frac{E_{Rm}}{E_{Im}} \right)_\parallel = \frac{Z_2 \cos(\theta_T) - Z_1 \cos(\theta_I)}{Z_2 \cos(\theta_T) + Z_1 \cos(\theta_I)} \quad (2.38)$$

$$\left(\frac{E_{Tm}}{E_{Im}} \right)_\parallel = \frac{2Z_2 \cos(\theta_T)}{Z_2 \cos(\theta_T) + Z_1 \cos(\theta_I)} \quad (2.39)$$

This case is depicted in fig. 4.2B. The GPR antenna orientation which corresponds to this case occurs when the antennae are positioned with their long axes parallel to the survey line (Turner, 1992).

In the case of normal incidence, all the E and H vectors are parallel to the interface, and the two cases above become identical, giving

$$\frac{E_{Rm}}{E_{lm}} = \frac{Z_2 - Z_1}{Z_2 + Z_1} \quad (2.40)$$

$$\frac{E_{Tm}}{E_{lm}} = \frac{2Z_2}{Z_2 + Z_1} \quad (2.41)$$

Under low loss conditions, conduction losses which are represented by the imaginary part of the denominator in eq. (2.35) are negligible and Z becomes

$$Z = \sqrt{\frac{\mu}{\epsilon}} = \sqrt{\frac{\mu_o}{\epsilon_o}} \sqrt{\frac{\mu_r}{\epsilon_r}} = Z_o \sqrt{\frac{\mu_r}{\epsilon_r}} = \frac{Z_o}{\sqrt{\epsilon_r}} \text{ for } \mu_r = 1 \quad (2.42)$$

When $\mu_r = 1$, Z can be substituted into eq. (2.40) and (2.41) to give the Fresnel equations as a function of the relative dielectric permittivity, ϵ_r , for each layer. Simplifying results in the normal incidence Fresnel equations

$$\frac{E_{Rm}}{E_{lm}} = \frac{\sqrt{\epsilon_{r1}} - \sqrt{\epsilon_{r2}}}{\sqrt{\epsilon_{r1}} + \sqrt{\epsilon_{r2}}} \quad (2.43)$$

$$\frac{E_{Tm}}{E_{lm}} = \frac{2\sqrt{\epsilon_{r1}}}{\sqrt{\epsilon_{r1}} + \sqrt{\epsilon_{r2}}} \quad (2.44)$$

From eq. (2.42), we see that in low loss conditions where $\mu_r = 1$ the electrical impedance of a material is inversely proportional to the square root of the dielectric constant, ϵ_r . Under these conditions, radar reflections originate at a changes in the dielectric constant. The strength of the returning GPR signal is dependent on the difference in dielectric constants, as shown by eq. (2.43) and (2.44).

2.3 GPR velocity analysis and depth determination

2.3.1 Velocity analysis

Due to the ease of interpretation and short duration of typical ground penetrating radar investigations, advanced processing of GPR data is still rarely accomplished. However, a number of papers have been published which outline the use of seismic processing techniques on GPR data (Fisher et al., 1992a; Young et al., 1995; Fisher et al., 1996). Fisher et al. (1992a) have taken GPR data acquisition and processing to the next level by collecting and processing multi-offset GPR data. The processing techniques used in this thesis are discussed in more detail in Chapter 4.

One processing technique which is very important to extracting depth estimates from ground penetrating radar is the analysis of radar propagation velocities. Velocity analysis is typically performed by analysing the direct wave and hyperbolic reflectors on common midpoint gathers (Fisher et al., 1992a; Huggenberger et al., 1994; van Overmeeren, 1994; Tillard and Dubois, 1995; Young et al., 1995; Greaves et al., 1998). When numerous strong diffractions are present in a common offset section, the moveout of diffraction curves can also be used to determine the average velocity of the overlying material (Young et al., 1995; Harari, 1996).

Direct wave velocities from CMP gathers are calculated from the inverse slope of the direct wave move out.

$$v_{direct} = \frac{\Delta x}{\Delta t} = \frac{x_2 - x_1}{t_2 - t_1} \quad (2.45)$$

In eq. (2.45) Δx is the difference in antenna offsets and Δt is the one-way travel time difference of the direct wave. This velocity is the very near-surface radar wave velocity. As an example, fig. 2.3 shows a series of CMP gathers collected at 11, 21, and 31 m along the

profile shown in fig. 2.2. The direct wave velocities calculated at these locations are

$$\begin{aligned} \text{CMP @ 11 m} \quad v_{\text{direct}} &= 0.122 \text{ m/ns} \\ \text{CMP @ 21 m} \quad v_{\text{direct}} &= 0.152 \text{ m/ns} \\ \text{CMP @ 31 m} \quad v_{\text{direct}} &= 0.132 \text{ m/ns} \end{aligned}$$

Analysis of the hyperbolic reflection curves on CMP gathers is somewhat more complicated than the direct wave calculation. Assuming that the reflector is a horizontal reflecting plane in a homogeneous medium, the relationship between two-way travel time, t , offset, x , and propagation velocity, v is

$$t^2 = \frac{x^2 + 4h^2}{v^2} \quad (2.46)$$

where h is the depth of the reflector. Looking at the problem in another way, the difference in time between the peak or zero-offset time, t_o and any other time on the hyperbola can be described as

$$\Delta t_c = t - t_o = t_o \left(\sqrt{1 + \frac{x^2}{t_o^2 v^2}} - 1 \right) \quad (2.47)$$

where Δt_c is also called the normal move-out (NMO) correction. In seismic processing, NMO corrections are used to flatten a CMP hyperbola prior to creating a zero-offset or stacked section by stacking the traces in each CMP together. In ground penetrating radar studies, the NMO is used primarily as a velocity analysis tool and to correct traces to zero offset equivalents. When multiple layers are present, the velocity determined for each hyperbola can be approximated as the root mean square (RMS) velocity of a wave which has travelled through all the layers above the reflecting layer. The equation for the RMS velocity is given as

$$(v_{RMS})_N = \left[\frac{\sum_{n=1}^N t_n v_n(z)^2}{\sum_{n=1}^N t_n} \right]^{\frac{1}{2}} \quad (2.48)$$

where $(v_{RMS})_N$ is the RMS velocity to the base of the Nth layer, t_n is the vertical one-way travel time through layer n, and v_n is the wave propagation velocity in layer n.

In order to determine the RMS velocities for a CMP, a combination of hyperbolic curve fitting, NMO analysis, and semblance analysis was used. NMO analysis involves examining the flattening effect of a chosen RMS velocity. Semblance is defined as the normalized output-to-input energy ratio (Yilmaz, 1987) defined as

$$S = \frac{1}{M} \frac{\sum_t s_{t(i)}^2}{\sum_t \sum_{n=1}^M f_{i,t(i)}^2} \quad (2.49)$$

where M is the number of traces in a CMP gather, and $f_{i,t(i)}$ is the amplitude of the ith trace at two-way travel time $t(i)$. The stacked amplitude, $s_{t(i)}$ is the amplitude at two way travel time $t(i)$ of a trace resulting from stacking the CMP traces at velocity v_{st} . The two way travel time $t(i)$ lies along the trial stacking hyperbola such that

$$t(i) = \left[t^2(0) + \frac{x_i^2}{v_{st}^2} \right]^{1/2} \quad (2.50)$$

Semblance analysis is generally displayed as a contour plot or a gated row plot of semblance with axes of two way travel time and velocity. An example is seen on the left panel in fig.

2.5. The maximum semblance is indicated in dark grey with decreasing semblance shown as lightening contours. Velocity picks on this CMP are indicated by the dark line on the semblance plot. The middle panel in fig. 2.5 is the CMP data from which the semblance panel has been determined. Notice that a mute has been applied to remove the air and ground waves. The right hand panel is the NMO corrected CMP showing the degree of flattening the chosen velocities have achieved. If the velocity is too low, the hyperbola will be overcorrected, causing the larger offsets to curve up. If the velocity is too high, the hyperbola will be under-corrected and the large offsets will curve down. Figure 2.5 is a 100 MHz central frequency CMP centred at 50 m on line 2 at the Gilby Waterflood site. Above 50 ns, the mute gate has caused the semblance calculations to be smeared and unreliable. Without the mute gate, this section of the semblance plot would be contaminated by the direct airwave and ground wave. Between the two-way times of 80 to 100 ns and offsets of 7 to 10 m on the NMO corrected panel, the reflector appears to have been under corrected. This is actually part of the direct wave which was not within the applied mute gate.

Once an RMS velocity function in time has been determined, the interval velocities of each layer can be found by using the Dix equation (Dix, 1955) given by

$$v_N^2 = \frac{t_N v_{RMS\ N}^2 - t_{N-1} v_{RMS\ N-1}^2}{t_N - t_{N-1}} \quad (2.51)$$

where v_N is the interval velocity, $v_{RMS\ N}$ is the RMS velocity down to the bottom of layer N, $v_{RMS\ N-1}$ is the RMS velocity down to the bottom of layer N-1, and t_N and t_{N-1} are the vertical two way travel times to the bottom of layers N and N-1 respectively. Interval velocity functions in time can then be used to convert GPR time sections to depth sections.

Throughout this discussion, it has been assumed that the reflecting layers are horizontal. While this is rarely the case in reality, it is generally true for the primary target in this study, the water table. Non-horizontal reflectors cause the RMS velocity determined from CMP

Figure 2.5 Semblance velocity analysis. The left panel shows the semblance plot in RMS velocity (m/ns) versus time (ns). The centre panel shows the CMP from which the velocities are determined. The right panel shows the NMO corrected CMP gather. (Data from the Gilby Waterflood site, line2, CMP @ 50 m. 100 MHz antenna frequency, 0.10 m station spacing.

gathers to be less than the actual RMS velocity value by a factor of $1/\cos\phi'$, where ϕ' is the apparent dip of the reflector along the survey line. For dip angles which do not exceed 15° , the ratio of NMO velocity to the actual RMS velocity is close to unity (Yilmaz, 1987).

2.3.2 Depth determination

Reflector depth estimation from ground penetrating radar has been accomplished through a number of different methods. The most basic depth estimation method is the use of a single velocity to create a depth scale on a time section. The methods for determining the velocities used in published material are quite varied and include; the use of known reflector depth and two way travel time (Johnson, 1992; Iivari and Doolittle, 1993; Harari, 1996), velocity

assumed from expected or known ϵ_r (Beres and Haeni, 1991; Arcone et al., 1998), velocity assumed from expected subsurface materials (Soldal et al., 1994), and velocity analysis of numerous CMPs to determine a single average velocity (van Overmeeren, 1994, 1998).

More detailed velocity analysis has been completed by Fisher et al. (1992a), Huggenberger et al. (1994), Tillard and Dubois (1995), and Greaves et al. (1996), using NMO or semblance analysis to determine detailed velocity functions at numerous locations. Huggenberger et al. (1994) completed detailed velocity analyses in a sand and gravel aquifer system, but didn't use the interpreted velocities to convert their sections to depth. Tillard and Dubois (1995) also performed velocity analysis on a number of CMP's in a granite quarry. They determined that using velocity analysis in place of extrapolating direct wave velocities to depth reduced the apparent depth error from ± 4 m to ± 2 m for a reflector approximately 24 m below the surface.

In order for more detailed velocity analysis to be used in depth determinations, migration is usually required to correctly position the reflectors in space. Fisher et al. (1992b) used reverse-time migration to position the GPR reflections in space and generally improve the interpretability of single fold radar data. See Fisher et al. (1992b) or Yilmaz (1987) for a complete explanation of the reverse-time migration technique. Using seismic processing techniques, Fisher et al. (1992a) found a great improvement in interpretability due to noise suppression when comparing a single-offset section to a multi-offset stacked section. They saw further improvement in the section following reverse-time migration. Most of the main lithologic features from a number of core logs near the line were positively correlated to specific reflectors on the migrated depth section. Greaves et al. (1996) took the data from Fisher et al. (1992a) a further step by examining the velocity variations and estimating water content from the Fisher et al. (1992a) dataset. They showed that the variation in radar velocity can be quite significant over a 450 m profile and that the quality of the stacked section improves as the number of velocity analysis locations increases. Lateral velocity variations of as much as 30 % were encountered in the more detailed areas. Interval

velocities calculated using eq. (2.51) were used to estimate water content and convert the water content profile to depth. The velocity analysis completed by Greaves et al. (1996) differed in final maximum depth to bedrock from the original work done by Fisher et al. (1992a) by approximately 5 m. Basement depths reported by Kiley and Annan (1989) (from Greaves et al., 1996) range from 0 to 30 m, making either the Fisher et al. (1992a) or Greaves et al. (1996) interpretation possible. The water table was not identified in either the Fisher et al. (1992a) or the Greaves et al. (1996) paper. However, Greaves et al. (1996) identified an area of high water content (30–40 %). The top of this zone may delineate the approximate water table surface.

2.4 GPR and Hydrogeology

The use of ground penetrating radar in hydrogeology has been well documented over the last decade (Beres and Haeni, 1991; Johnson, 1992; Iivari and Doolittle, 1994; van Overmeeren, 1994; Harari, 1996; van Overmeeren, 1998 and others). Detailed water table mapping using ground penetrating radar data have been completed by Johnson (1992), Iivari and Doolittle (1994), and van Overmeeren (1994, 1998). In these studies, water table depths were determined by using a single velocity for the entire area, based on one of the methods introduced in Section 2.3. Detailed velocity estimation has not been used for water table mapping using ground penetrating radar data. As the lateral velocity variations encountered in an area can be significant (Greaves et al., 1996), an understanding of the cause of variations and their significance is important.

The water table reflection occurs when the GPR wave encounters a change in ϵ_r , which is the result of a change from unsaturated to saturated conditions in the subsurface. Equations 2.34, 2.43, 2.44 show how the velocity, reflection and transmission coefficients for GPR waves depend on the dielectric constant of the material through which the wave is passing. Topp et al. (1980) showed that the dielectric constant of a soil is primarily dependent on the volumetric water content, θ of the soil. The difference in dielectric constant from a dry sand

to a saturated sand is from $(\epsilon_r)_{\text{dry sand}} = 3-5$ to $(\epsilon_r)_{\text{saturated sand}} = 20-30$ (see Table 2.1). The range in ϵ_r for a saturated sand is due to the range of porosities, and therefore saturated water contents, possible in sands. The change in dielectric constant as an electromagnetic wave travels from unsaturated to saturated conditions tends to result in a strong radar reflection, as long as the boundary between unsaturated and saturated conditions is fairly sharp compared to the central wavelength of the GPR signal.

The water table is defined in Freeze and Cherry (1979) as the surface on which the fluid pressure in the pores of a medium is exactly atmospheric. Above the water table, there commonly exists a capillary fringe or tension-saturation zone, where surface tension causes the pore space to be saturated even though the fluid pressure is less than atmospheric. As GPR is sensitive to changes in water content rather than pressure changes, the reflection associated with the water table actually occurs at the top of the capillary fringe. Figure 2.6 illustrates the GPR response to the water table and capillary fringe zone. The monitoring well on the left provides the depth to the water table. In the centre, a hypothetical volumetric water content profile is shown. The water content, θ , is relatively constant throughout the vadose zone, increases rapidly at the top of the capillary fringe zone, and then is constant and equal to the porosity of the soil through the capillary fringe and beneath the water table. On the right, a GPR trace is shown, with the coupled air and ground waves near the surface, and a reflection from the change in θ at the top of the capillary fringe zone. The polarity of the capillary fringe reflection is 180° out of phase with the air wave. This is due to the value of the reflection coefficient. At its most extreme, the dielectric constants from the unsaturated and saturated zone are approximately $(\epsilon_r)_{\text{dry soil}} = 4$ to $(\epsilon_r)_{\text{saturated soil}} = 25$. Using eq. (2.43) for the Fresnel reflection coefficient at normal incidence, the reflection coefficient at the top of the capillary fringe would be

$$R_{\text{capillaryfringe}} = \frac{E_{Rm}}{E_{Im}} = \frac{\sqrt{4} - \sqrt{25}}{\sqrt{4} + \sqrt{25}} = -\frac{3}{7} \quad (2.52)$$

The negative sign indicates that the reflected wave will have a 180° polarity shift from the

Figure 2.6 Schematic diagram of the GPR response to the water table and capillary fringe zone. The water table is the flat surface near the bottom of the figure. The top of the capillary fringe zone is shown above the water table. From left to right, decreasing pore size results in an expansion of the capillary fringe zone. The monitoring well on the left samples the water table depth only. In the centre, a hypothetical volumetric water content profile is depicted. The water content, θ , is relatively constant throughout the vadose zone, increases rapidly at the top of the capillary fringe zone, and then is constant and equal to the porosity of the soil through the capillary fringe and beneath the water table. On the right, a hypothetical GPR trace is depicted, with the coupled air and ground waves near the surface and a reflection from the top of the capillary fringe zone.

incident wave.

The height of the capillary fringe zone above the water table is controlled by a number of factors, including pore size distribution and the wetting and drying history of the site. The main control on the capillary fringe height is the pore size distribution of the sediments. In fig. 2.6, pore size is decreasing to the right, causing a corresponding expansion of the capillary fringe zone. The height of capillary rise, h_c , can be calculated using

$$h_c = \frac{2\sigma \cos \lambda}{\rho_w g R} \quad (2.53)$$

where σ is the surface tension of the fluid, λ is the angle of the meniscus with the capillary tube, ρ_w is the density of the fluid, and R is the radius of the capillary tube or pore size. For water at 18 °C, σ is 73 g/s², ρ_w is 0.999 g/cm³, and λ may be taken as 0. Then eq. (2.53) becomes

$$h_c = \frac{2 \times 73 \text{ g/s}^2 \times \cos 0}{0.999 \text{ g/cm}^3 \times 980 \text{ cm/s}^2 R} = \frac{0.15}{R} \text{ cm}^2 \quad (2.54)$$

When identical spherical grains are subject to cubic packing, the radius of the pore spaces is equal to 0.2 times the grain diameter. Table 2.2 displays the average grain diameter, radius of pore spaces, and capillary rise predicted from eq. (2.54) for various sediments. In a real environment, the actual capillary rise above the water table can be irregular due to the spatial variations in sediment texture. The largest diameter pore space dictates the height of capillary rise above the water table.

Table 2.2 Capillary rise for various sediments (Fetter, 1994).

Sediment	Grain Diameter (cm)	Pore Radius (cm)	Capillary Rise (cm)
Fine silt	0.0008	0.0002	750
Coarse silt	0.0025	0.0005	300
Very fine sand	0.0075	0.0015	100
Fine sand	0.0150	0.003	50
Medium sand	0.03	0.006	25
Coarse sand	0.05	0.010	15
Very coarse sand	0.20	0.040	4
Fine gravel	0.50	0.100	1.5

From this discussion, it is obvious that there is an additional difficulty in determining water table depths from ground penetrating radar. The capillary fringe reflection occurs at an unknown distance above the water table. While it is possible to estimate the capillary rise at a logged monitoring well, extrapolating this estimate to the rest of the site requires the assumption that the soil is homogeneous at the water table. This is rarely the case. For instance at the Gilby Waterflood site, the three monitoring wells show fine sand with trace fine gravel, medium sand, and sand and gravel at the water table. Thus the theoretical capillary rise ranges from 4 cm for very coarse sand to 50 cm where fine sand is located. Estimating the capillary rise will be important in determining the water table elevations from ground penetrating radar.

Figure 2.6 shows an example of static conditions for the volumetric water content and capillary rise in an unconfined aquifer. Static conditions rarely occur in reality. Rising or falling water tables, wetting history of the soil, and even temperature affect the actual height of the capillary fringe zone. A rising water table results in a smaller fringe zone, while a falling water table tends to stretch the capillary fringe zone beyond 'static' levels. The wetting history of the soil affects not only the capillary fringe height, it also affects the volumetric water content profile throughout the unsaturated zone. Soil-water retention curves are used to explain this phenomenon. A soil-water retention curve such as the one in fig. 2.7 shows the relationship between pressure head, ψ , and volumetric water content θ . The pressure head, ψ , which is also called the moisture potential or matric suction, is a negative pressure which results from soil-water attraction (Fetter, 1994). It increases with decreasing amounts of soil moisture and is measured with tensiometers. Figure 2.7 illustrates the complications which can arise from the wetting history of the soil. There is a hysteresis effect in the relationship between pressure head and volumetric water content. At a specific pressure head, a drying soil will have a larger θ than a wetting soil. As well, the capillary fringe is more extensive in a drying soil than in a wetting soil. For a more complete discussion of the cause of hysteresis in soil-water retention curves, see Hillel (1971).

Figure 2.7 Idealized curve showing the relationship between volumetric water content, θ , and the soil-moisture pressure head, ψ (Fetter, 1994).

As a further complication to both unsaturated flow studies and the use of GPR in hydrogeologic studies, even fig. 2.7 is an oversimplification which does not include the effects of infiltration pulses. Figure 2.8 depicts infiltrated water as it moves downward through the vadose zone (Fetter, 1994). In part A, the volumetric water content has increased to saturation, θ_{sat} , at the surface. Following infiltration (part B), the slug of water begins to move deeper, and θ drops below θ_{sat} . The slug of water continues to move downward, and the volumetric water content continues to decrease throughout the column (part C). In part D, the infiltrated water reaches and raises the water table. The near surface volumetric water content has been reduced to the field capacity. Field capacity is the volumetric water content of a soil following gravity drainage. Precipitation frequency and duration, the velocity of vertical water movement in the unsaturated zone and thickness of the unsaturated zone all

Figure 2.8 Moisture profiles showing the downward migration of a wave of infiltrated water. $\theta_{sat} = 0.29$ and $\theta_{field\ capacity} = 0.06$ (adapted from Fetter, 1994).

affect the actual θ profile. As volumetric water content changes are the primary cause of GPR reflections, the θ profile in the unsaturated zone is very important. For instance, if there has been a significant amount of precipitation for a long enough time or the water table is close to the surface, it is possible for the vadose zone to be near saturation, with only minor variations in the volumetric water content throughout the column. This would result in a poor radar response from the water table. As well, since the volumetric water content has a large effect on the dielectric constant, ϵ_r , and square root of ϵ_r is inversely proportional to the radar wave velocity (eq. (2.34)), the velocity profile of the subsurface is not constant. Instead, it is constantly changing as the volumetric water content profile in the unsaturated zone changes.

Annan et al. (1991) used a simplified model for the change in volumetric water content near the top of the capillary fringe zone to examine the GPR response from this region. The model used for the variation in θ with depth was

$$\begin{aligned} \theta &= m & z < T - A - W \\ \theta &= m + (n - m) \times F((z - (T - A - W))/W) & T - A - W < z < T - A \\ \theta &= n & z > T - A \end{aligned} \quad (2.55)$$

where z was depth, m was the field capacity, n was the saturated volumetric water content, T was the depth of the water table, A was the air entry pressure head or height of the capillary fringe, W was the thickness of the transition zone and F was a site specific function which describes the drop in θ from saturated to field capacity levels. Annan et al. (1991) used a simple cosine drop off

$$F(x) = \frac{1 - \cos(\pi x)}{2} \quad (2.56)$$

to model the transition zone between saturated and field capacity conditions. This is a significant simplification when compared to the empirical formulas which have been developed in hydrogeology. Volumetric water content versus depth was converted to

dielectric constant versus depth (fig. 2.9A) using the Topp et al. (1980) relationship. Synthetic radargrams were computed for various transition zone thicknesses (W) and radar wavelengths, λ . Figure 2.9B shows the normalized reflection amplitude versus the ratio of transition thickness to radar wavelength, W/λ . Annan et al. (1991) point out that the data in fig. 2.9B are specific to the model used, but it can be seen that there is a significant decrease in reflection amplitude when the W/λ ratio exceeds 0.3. Therefore Annan et al. (1991) suggest that longer wavelength (lower frequency) antennas should be chosen when water table detection is the goal. However, since longer wavelengths decrease the GPR resolution, there is a trade off between the strength of the reflection and the resolution of the target. While the unsaturated zone water content profile was oversimplified, Annan et al. (1991) have shown the importance of a relatively sharp transition from unsaturated to saturated conditions compared to the radar wavelength in order to effectively image the water table.

Figure 2.9 A Model of dielectric constant, ϵ_r , versus depth for various transition widths, W . B Change in peak water table reflection amplitude versus ratio of transition zone width to radar wavelength (W/λ) (adapted from Annan et al., 1991).

2.5 Summary

In this chapter, we have discussed various aspects of ground penetrating radar and hydrogeology theory, interspersed with previous work in these fields. An overview of the electromagnetic theory which describes ground penetrating radar has lead to equations for the velocity of a GPR wave

$$v = \frac{c}{\sqrt{\epsilon_r}} \quad \mu_r \approx 1$$

and reflection and transmission coefficients for normal incidence waves of

$$R = \frac{E_{Rm}}{E_{Im}} = \frac{\sqrt{\epsilon_{r1}} - \sqrt{\epsilon_{r2}}}{\sqrt{\epsilon_{r1}} + \sqrt{\epsilon_{r2}}}$$

$$T = \frac{E_{Tm}}{E_{Im}} = \frac{2\sqrt{\epsilon_{r1}}}{\sqrt{\epsilon_{r1}} + \sqrt{\epsilon_{r2}}}$$

Previous work on velocity analysis, depth estimation, and water table mapping with GPR was discussed. The strong reflection from the boundary between the unsaturated and saturated zones occurs at the top of the capillary fringe, and is 180° out of phase with the airwave. As the position of the water table is our primary concern, the height of the capillary fringe above the water table must be determined. The capillary rise can be estimated, given the grainsize distribution of the sediments at the water table. Sites are rarely composed of homogeneous material, however, and capillary rise must be extrapolated from well locations where it can be estimated to GPR locations.

The actual height of the capillary rise is dependent on more than the sedimentary grainsize. Rising and falling water tables, the wetting history of the soil and temperature also effect the

capillary rise. These conditions also affect the unsaturated zone, causing variability in the water content throughout the unsaturated zone. Variable moisture content caused variations in the dielectric constant, which directly affect the velocity of the GPR wave and the reflections patterns seen in the unsaturated zone.

Finally, we discussed a paper by Annan et al. (1991) which showed that a relatively sharp transition between unsaturated and saturated conditions, compared to the radar wavelength, is needed to effectively image the capillary fringe. Survey design for water table investigations should take into account grainsize distributions, antenna frequencies, as well as recent weather conditions and precipitation events in order to get the best possible target response.

3. Sources of Error in Water Table Elevation Estimates

The objective of this thesis is to estimate water table elevations and determine the uncertainty in the estimate. At a particular site, there are two sources of water table elevations, direct measurement from monitoring wells, and estimation from ground penetrating radar sections. Both of these water table elevations have a number of sources of error which contribute to the uncertainty in the elevation estimates. In this chapter, we will investigate each of these sources of error to determine their effect on the accuracy of the water table elevation estimated from ground penetrating radar and measured from monitoring wells.

3.1 Uncertainty Propagation

In a given experiment, a number of values are often measured and combined in some way to estimate the value of some parameter. Each measured value has an uncertainty or estimated error, such as the standard deviation. In order to determine the uncertainty in the estimated parameter, we must have some way of combining the uncertainties in the measured values to give the uncertainty in the estimated parameter. A common method for doing this is outlined in Bevington (1969) and works as follows. Suppose x is a function of at least two variables, u and v , which have been measured and which have an uncertainty given by their variances, σ_u^2 and σ_v^2 .

$$x = f(u, v) \quad (3.1)$$

Assuming that the mean value for x can be determined using the mean values for u and v , it can be shown that the variance of x , σ_x^2 is equal to

$$\sigma_x^2 = \sigma_u^2 \left(\frac{\partial x}{\partial u} \right)^2 + \sigma_v^2 \left(\frac{\partial x}{\partial v} \right)^2 + \sigma_{uv}^2 \left(\frac{\partial x}{\partial u} \right) \left(\frac{\partial x}{\partial v} \right) \quad (3.2)$$

In eq. (3.2), each of the partial derivatives is evaluated with the other variables fixed at their

mean values. The first two terms in eq. (3.2) are the averages of the squares of the deviations and usually dominate the expression. The third term introduces the covariance, σ_{uv}^2 , between the variables u and v . If we can assume that the fluctuations in u and v are uncorrelated, then the covariance term can be neglected. The expression for σ_x^2 becomes

$$\sigma_x^2 = \sigma_u^2 \left(\frac{\partial x}{\partial u} \right)^2 + \sigma_v^2 \left(\frac{\partial x}{\partial v} \right)^2 \quad (3.3)$$

If x is a function of more than two variables, these simply add variance and covariance terms to eq. (3.2) or (3.3).

Variances have been used in the preceding discussion as an expression of uncertainty. Other values for uncertainty can be used in this propagation method by substituting the square of the uncertainty for the variance in equation 3.3. For instance if Δu , Δv , and Δx are the uncertainties in u , v , and x , then the uncertainty in x is equal to

$$\Delta x^2 = \Delta u^2 \left(\frac{\partial x}{\partial u} \right)^2 + \Delta v^2 \left(\frac{\partial x}{\partial v} \right)^2 \quad (3.4)$$

3.2 Sources of Uncertainty in Water Table Elevation Estimates from Ground Penetrating Radar

Water table elevations from ground penetrating radar are determined in the following manner. The depth from the surface to the top of the capillary fringe, d_{cap} , is determined from the ground penetrating radar data. The capillary fringe elevation, $CF_{elev}(GPR)$, is determined by subtracting d_{cap} from the surface elevation, Z .

$$CF_{elev}(GPR) = Z - d_{cap} \quad (3.5)$$

In order to determine the water table elevation, the height of the capillary rise above the water table, h_{cap} , must be estimated. The water table elevation from GPR is then

$$WT_{elev}(GPR) = CF_{elev}(GPR) - h_{cap} = Z - d_{cap} - h_{cap} \quad (3.6)$$

Therefore, the water table elevation estimated from GPR data is determined from surface elevations, capillary fringe depths from GPR depth sections and the estimated capillary fringe height. Each of these values has sources of error which contribute to an uncertainty in the final result. We will see that the height of the capillary rise above the water table is our greatest uncertainty. Therefore, knowing the capillary fringe and water table elevation estimates from ground penetrating radar sections, along with their uncertainties, is important.

Assuming the uncertainties in Z , d_{cap} , and h_{cap} can be quantified, the uncertainty propagation method described in section 3.1 can be used to determine the uncertainty in the CF_{elev} and the WT_{elev} .

$$\Delta CF_{elev}(GPR) = \sqrt{\Delta Z^2 + \Delta d_{cap}^2} \quad (3.7)$$

$$\Delta WT_{elev}(GPR) = \sqrt{\Delta CF_{elev}(GPR)^2 + \Delta h_{cap}^2} = \sqrt{\Delta Z^2 + \Delta d_{cap}^2 + \Delta h_{cap}^2} \quad (3.8)$$

In the next sections, each of the sources of error in the estimation of the water table elevations from GPR data will be examined. We will quantify the uncertainties each of these sources of error brings to the overall uncertainty in the water table elevation estimates from ground penetrating radar.

3.2.1 Surface Elevation Uncertainties

In order to make accurate depth estimates and to compare these estimates to information from wells, it is necessary to have accurate surface elevation data. Survey data has an uncertainty which is based on the equipment used. As well, ground penetrating radar data is generally collected on a fine station spacing compared to typical survey data. For instance,

GPR station spacings ranged from 0.05 to 1.00 m in the three datasets. Survey station spacings generally ranged from 1.0 to 2.5 m. Interpolation of surface elevations for the unsurveyed GPR stations can introduce an added uncertainty to the surface elevations.

Survey data has an uncertainty in its measurements which is dependent on the equipment and method used. With the theodolite and distomat system, the uncertainty in measured horizontal and vertical angles from the theodolite to the reflector is 3 arc seconds. The standard deviation of the slope from theodolite to reflector was measured to be ± 0.005 m for a distance of 200 m. For most of the surveyed stations (distances < 100 m) ± 0.005 m is the maximum uncertainty that might occur in the measured slope.

The uncertainty method outlined in section 3.1 can be used to combine the uncertainties in the measured horizontal and vertical angles and slope to give the uncertainty in Northing, Easting and elevation values for the surveyed points. Equation 3.9 shows the equations used to calculate ΔN , ΔE , and ΔZ .

$$\begin{aligned}\Delta N &= \sqrt{(\Delta s \sin V \cos H_z)^2 + (s \Delta V)^2 (\sin^2 V \sin^2 H_z + \cos^2 V \cos^2 H_z)} \\ \Delta E &= \sqrt{(\Delta s \sin V \sin H_z)^2 + (s \Delta V)^2 (\cos^2 V \sin^2 H_z + \sin^2 V \cos^2 H_z)} \\ \Delta Z &= \sqrt{(\Delta s \cos V)^2 + (s \sin V \Delta V)^2 + \Delta h^2}\end{aligned}\tag{3.9}$$

In eq. (3.9) ΔN , ΔE , and ΔZ are the uncertainties in the northing, easting and elevation respectively. The vertical and horizontal angles, V and H , and their uncertainty, $\Delta V = \Delta H = 3'' = 0.00083^\circ$, were measured from the theodolite, and the slope, s was determined by the distomat reflector system. The uncertainty in the slope, Δs was taken to be ± 0.005 m, and the uncertainty in the height of the reflector and distomat, Δh , was taken to be ± 0.005 m. Equation (3.9) can be evaluated for each of the survey stations, or it can be evaluated using the mean values for V , H_z and s to give overall values for ΔN , ΔE and ΔZ . Table 3.1 gives the uncertainties in Northing, Easting and elevation values for the North Caicos and Gilby datasets. The Strathroy GPR lines were not surveyed, but survey points

Table 3.1 Uncertainties in Northing, Easting and Elevation values at survey stations (Gilby and North Caicos Island) and kriging (Stathroy).

	ΔN (m)	ΔE (m)	ΔZ (m a.s.l.)
Gilby	± 0.003	± 0.004	± 0.005
North Caicos Is.	± 0.004	± 0.003	± 0.005
Strathroy			± 0.037

from the field and Northing and Easting values for the line ends were provided. The N and E values for the GPR stations were linearly interpolated from the end points. The field survey values were used with a standard kriging algorithm to determine elevations along the line. Kriging is an interpolation technique which uses a weighted average of the nearest known values to estimate the value at an unknown point. The square root of the kriging variance can be used to provide an uncertainty in the elevation estimates and is listed in Table 3.1.

GPR data was collected at station spacings ranging from 0.05 to 1.00 m for the three datasets. The survey station spacing along the GPR lines ranged from 1.0 to 2.5 m for the Gilby and North Caicos Island surveys. It was therefore necessary to interpolate between survey stations to provide Northing, Easting, and elevation data for each of the GPR stations. This can introduce an additional uncertainty in the surface elevations at the GPR stations which were not surveyed. We chose a linear interpolation due to the nearness of the survey points and the generally smooth surface changes along the lines. To investigate the elevation uncertainties which could be introduced by linear interpolation, we took a cosine wave to approximate surface features and chose various station spacings, L , to simulate surveying an undulating surface. The maximum error caused by interpolating at cosine curve occurs when the surveyed points are $L/2$ to either side of a peak or a trough in the cosine curve. The maximum error, $E_{\max}$ is then equal to

$$E_{\max} = \pm A \left[1 - \cos\left(\frac{2\pi}{\lambda} \frac{L}{2}\right) \right] \quad (3.10)$$

where λ is the wavelength of the cosine wave which approximates the wavelength of surface undulations, A is the amplitude of the cosine wave and L is the station spacing. Figure 3.1 shows lines of constant $E_{\max}$ for survey station spacings from 0.5 to 5 m and wavelengths from 0 to 360 m. Given the amplitude and wavelength which could simulate the surface topography, fig. 3.1 can be used to determine the survey station spacing needed to achieve a certain accuracy for linearly interpolated locations. As traditional surveying takes a considerable amount of time, fig. 3.1 can provide quantitative information to help in determining what station spacing is needed to achieve an acceptable interpolation error. When a typical site is surveyed for surface elevations, station spacings are generally much greater than the maximum of 5 m used in fig. (3.1). Depending on the surface topography,

Figure 3.1 Lines of constant maximum error, $E_{\max}$, for a range of survey station spacings and surface undulation wavelengths, λ . The maximum error in metres is dependent on the amplitude of the surface undulations, A .

a survey station spacing greater than 5 m is likely to result in maximum errors in interpolated elevations of between 0.01 and 0.1 m. For the purposes of this thesis, we would like to reduce avoidable errors such as this as much as possible.

For the Gilby Waterflood data, the general topography can be represented by a sine wave with an amplitude between 1 and 1.2 m and a wavelength of 240 to 300 m. One line had a greater change in topography, and would require a wavelength of 100 m. Survey data at the Gilby study site was collected with station spacings between 1 and 2.5 m, with a 2 m station spacing for most of the data. Using an amplitude of 1 m, a station spacing of 2 m and wavelengths of 100 and 300 m, the maximum error in linearly interpolated surface elevations is $E_{\max, \lambda=100 \text{ m}} = 0.002 \text{ m}$ and $E_{\max, \lambda=300 \text{ m}} = 0.0002 \text{ m}$. The North Caicos Island surface topography has a larger variability in amplitude and wavelength. The smallest interpolation error of $E_{\max} = 0.0009 \text{ m}$ occurred when $L = 2.5 \text{ m}$, $\lambda = 100 \text{ m}$ and $A = 0.3 \text{ m}$. The largest $E_{\max}$ value on the island occurred when $L = 1.0 \text{ m}$, $\lambda = 44 \text{ m}$ and $A = 1.1 \text{ m}$, giving an interpolation error of $E_{\max} = 0.002 \text{ m}$. For both the Gilby and North Caicos Island datasets, the $E_{\max}$ values are less than the elevation uncertainty of the surveyed points themselves. Therefore, any uncertainty due to interpolation can be ignored. The surface elevations from the Strathroy site were all interpolated using kriging, which provided an estimate of the uncertainty in the estimated elevations.

3.2.2 Uncertainty in the Capillary Fringe Depth from Ground Penetrating Radar Data

The uncertainty in the depth to the capillary fringe determined from ground penetrating radar has a number of sources. These include uncertainty in the depth to a reflector due to uncertainty in the picked depth values, uncertainty in the velocities determined from a CMP gather and uncertainties due to interpolation of the velocity profile between CMP gathers. As well, the resolution possible due to the frequency of the GPR signal can affect the estimate of the depth to the capillary fringe from ground penetrating radar.

3.2.2.1 Uncertainty in the Picked Depth Values

The uncertainty in the picked depth values is dependent on two values which provide the minimum and maximum extremes of the uncertainty. As ground penetrating radar is collected by sampling the returning radar signal at a specific rate, the minimum uncertainty in any time pick, $\Delta t_{\text{pick min}}$ is equal to half the sampling interval. All of the GPR data used in this thesis was sampled at a rate of 0.8 ns, giving a minimum uncertainty in any time pick of $\Delta t_{\text{pick min}} = 0.4$ ns. The maximum uncertainty in the time picks comes from intuitive reasoning. Intuitively, the accuracy of a pick is related to the frequency of the radar wave and the signal-to-noise ratio of the data. For seismogram data, Aki and Richards (1980) have given a formula to quantify their intuitions. Stewart et al. (1984) converted this equation to give the time picking error, t_e of

$$t_e = \frac{1}{f_m \log_2 \left[1 + \left(\frac{S}{N} \right)^2 \right]} \quad (3.11)$$

where f_m is the frequency of the radar wave, S is the RMS amplitude of the signal, and N is the RMS amplitude of the noise. As the capillary fringe reflection is our primary target, we used the RMS amplitude of the capillary fringe reflection for the signal, and the RMS amplitude below the last identifiable reflection as the noise. The signal to noise ratio was determined to be 6.69, giving a time picking error of

$$\Delta t_{\text{max } 100\text{MHz}} = t_{e \text{ } 100\text{MHz}} = 1.8 \text{ ns} \quad \Delta t_{\text{max } 200\text{MHz}} = t_{e \text{ } 200\text{MHz}} = 0.91 \text{ ns} \quad (3.12)$$

Stewart et al. (1984) mention that these values are probably realistic for a single trace, but are likely too large for hand-picked sections where events can be correlated from trace to trace. Therefore, the time picking error determined using the frequency and signal-to-noise ratio is the maximum uncertainty in the time picks. The actual uncertainty lies somewhere between these two extremes.

The uncertainty in the time picks can be converted to the uncertainty in the depth picks using the interval velocity, v

$$\Delta d_{pick} = \frac{v \Delta t_{pick}}{2} \quad (3.13)$$

For the minimum and maximum values Δt_{pick} values above and an interval velocity of 0.10 m/ns, eq. (3.13) gives

$$\begin{aligned} \Delta d_{pick \min} &= \pm 0.02 \text{ m} \\ \Delta d_{pick \max 100 \text{ MHz}} &= \pm 0.09 \text{ m} \quad \Delta d_{pick \max 200 \text{ MHz}} = \pm 0.04 \text{ m} \end{aligned} \quad (3.14)$$

As there is no reasonable method to determine the actual value for Δd_{pick} between these two extremes, the maximum uncertainty in the depth pick will be used in the calculation of the uncertainty in the water table elevation estimates from ground penetrating radar.

3.2.2.2 Velocity Uncertainty on a Single Common Midpoint Gather

As discussed in Chapter 2, the hyperbolic reflections on a single CMP gather can be used to determine the root-mean squared (RMS) GPR wave velocity to the peak two-way time for the hyperbola. When picking these velocities, there is an uncertainty in the velocity which can lead to an uncertainty in depth when the GPR time section is converted to depth. The uncertainty in the RMS velocity can be determined from the semblance plot, as shown in fig. 2.5. The half width of the maximum semblance contour can provide a method for quantifying the uncertainty in the picked velocities. For the 100 MHz central frequency data in fig. 2.5, the average half width of the maximum semblance contours for reflections above the water table is ± 0.003 m/ns (3 mm/ns). This is typical of the 100 MHz data collected. The maximum semblance contours for the 200 MHz central frequency data tend to be slightly smaller, with an average half width of ± 0.0025 m/ns (2.5 mm/ns).

Propagating the uncertainty in the radar velocities determined from the semblance plots to an uncertainty in depth is not a simple task. A max/min approach was chosen as the most efficient method. In the max/min analysis method, the maximum and minimum velocity values are determined for all of the velocity functions using the original velocity functions as a guide. The original velocity functions are used in all processing steps until the time-to-depth conversion is performed. The time-to-depth conversion is performed three times, with the actual velocity functions, the maximum velocity functions and the minimum velocity functions. The maximum uncertainty in the capillary fringe depth due to velocity uncertainty is then equal to half the maximum depth minus the minimum depth. This results in an uncertainty in the depth to the capillary fringe of $\Delta d_{\text{cap}} = \pm 0.05$ to ± 0.15 m.

3.2.2.3 Velocity Uncertainty due to Heterogeneity and Spatially Sparse CMP Locations

The second cause for velocity uncertainty results from the need to collect CMP spreads less frequently than the common offset station frequency. This can create an uncertainty in the velocity along a line if there is a change in near surface conditions which occurs more frequently than the CMP data collection along the line. This is not generally a problem in areas which have a relatively homogeneous body such as in a granite or limestone quarry, or within saturated sediments, where the velocity range is relatively small. However, in a near surface composed of unsaturated sediments, changes in water content can cause considerable changes in velocity. These changes are expected in the vertical direction, as the water content increases towards saturation at the capillary fringe. Lateral velocity changes can also occur due to stratigraphic changes, variations in grain size distribution, variable precipitation and uneven wetting fronts moving through the subsurface. The extent to which lateral velocity variations can occur is an important issue in a field where depth calculations are usually carried out using a single velocity or velocity function. As discussed in Section 2.4, Greaves et al. (1996) showed that lateral velocity variations as high as 30 % could be seen at a relatively heterogeneous site.

An investigation was conducted to determine the extent to which changes in the near surface will effect the RMS velocity at the capillary fringe surface. Velocity variations are primarily dependent on changes in water content and soil characteristics. For simplicity, it is assumed that the soil characteristics are relatively homogeneous, remaining in the same general classification group. If the water content as a function of distance above the water table can be estimated, then the dielectric constant, ϵ_r , can be determined from empirical equations developed by Topp et al. (1980). Interval radar velocities can be determined from the estimated dielectric constant, and the RMS velocity of a radar wave at the top of the capillary fringe can be calculated. The variation in RMS velocities for a specific soil classification group will give us an indication of the variation likely to be seen at a particular site.

Clapp and Hornberger (1978) and Cosby et al. (1984) approximated the relationship between pressure head (ψ) and volumetric water content (θ) as a power law.

$$|\psi(S)| = |\psi_s| S^{-b} \quad S = \frac{\theta}{\phi} \quad (3.15)$$

where $\psi(S)$ is the pressure head as a function of S , S is the degree of saturation, θ is the volumetric water content, ϕ is the porosity, b is the pore-size distribution index, and ψ_s is the saturated pressure head or the height of capillary rise above the water table. Equation (3.15) neglects the complications resulting from the hysteresis effects discussed in Chapter 2. Numerous equations have been used to explain the moisture characteristics of a soil. Cosby et al. (1984) point out that for their data “the limited number of measurements taken for each sample and the large amount of variability in the available data suggest that a simpler representation of the hydraulic properties is desirable”. They also point to studies (eg. Ghosh, 1980) which show that the power law provides a sufficient relationship between ψ and θ in many cases. In our case, we are looking for a method to approximate the magnitude of moisture content variations in the unsaturated zone, so the power law should suffice.

Equation (3.15) contains two empirical parameters - the saturated pressure head, ψ_s , and the

slope of the retention curve on a logarithmic graph, b . Clapp and Hornberger (1978), and Cosby et al. (1984) show that ψ_s and b depend primarily on soil texture. They analyzed a set of data from Holtan et al. (1968) and Rawls et al. (1976) containing 1873 soil samples taken from 35 localities in 23 states in the United States. In both papers they only considered samples for which both moisture characteristics and saturated conductivities had been determined. As well, they eliminated any samples which had been texturally classed as rock fragments or identified as the R horizon. This left 1446 and 1448 samples respectively. For each sample, ψ_s and b were determined using eq. (3.15). Saturated hydraulic conductivity, K_s and saturated water content, θ_s were also available for each sample. The assumption has been made that θ_s and ϕ are equivalent. The samples were then grouped according to their textural class and the mean and variance were determined for each of the hydraulic parameters. Table 3.2 shows the results of this analysis for the various textural classes from Cosby et al. (1984).

Rewriting eq. (3.15), the pressure head can be expressed as a function of θ , ϕ , ψ_s , and b .

$$|\psi(\theta)| = |\psi_s| \phi^b \theta^{-b} \quad (3.16)$$

This equation can then be rearranged to express the water content as a function of ϕ , pressure head $\psi(z)$, ψ_s , and b .

$$\theta(z) = \left(\frac{|\psi_s|}{|\psi(z)|} \right)^{\frac{1}{b}} \phi \quad (3.17)$$

From Chapter 2, ψ changes with depth below the surface, z , giving rise to a change in θ with depth. Using the data from Cosby et al. (1984), which gives ψ_s , θ_s , and b for a given soil texture, and assuming a $\psi = \psi(z)$ function, we can determine the volumetric water content versus depth from eq. (3.17).

Topp et al. (1980) provide an empirical relationship between water content and dielectric

Table 3.2 Mean and standard deviation for hydraulic parameters b , $\log \psi_s$ (ψ_s is measured in centimetres), and θ_s for various textural classes (adapted from Cosby et al., 1984).

Class	n	$\bar{b}$	σ_b	$\overline{\log \psi_s}$	$\sigma_{\log \psi_s}$	$\bar{\theta}_s$	σ_{θ_s}
sandy loam	124	4.74	1.40	1.15	0.73	43.4	8.8
sand	14	2.79	1.38	0.84	0.56	33.9	7.3
loamy sand	30	4.26	1.95	0.56	0.73	42.1	7.2
loam	103	5.25	1.66	1.55	0.66	43.9	7.4
silty loam	394	5.33	1.72	1.88	0.38	47.6	5.4
sandy clay loam	104	6.77	3.39	1.13	1.04	40.4	4.8
clay loam	147	8.17	3.74	1.42	0.72	46.5	5.4
silty clay loam	325	8.72	4.33	1.79	0.58	46.4	4.6
sandy clay	16	10.73	1.54	0.99	0.56	40.6	3.2
silty clay	43	10.39	4.27	1.51	0.84	46.8	6.2
light clay	148	11.55	3.93	1.67	0.59	46.8	3.5
all classes	1448	7.22	3.86	1.59	0.70	45.7	6.1

constant, ϵ_r , which can be used to determine the dielectric constant at depth from the volumetric water content values determined using eq. (3.17).

$$\epsilon_r(z) = 3.03 + 9.3\theta(z) + 146.0\theta(z)^2 - 76.7\theta(z)^3 \quad (3.18)$$

The dielectric constant can be converted to a radar wave velocity at depth using eq. (2.34), which is repeated here,

$$v(z) = \frac{c}{\sqrt{\epsilon_r(z)}} \quad (3.19)$$

Equation (3.19) gives the discrete interval velocities, v , for each depth interval. The RMS

velocity to the top of the capillary fringe can be determined from the RMS velocity equation (previously eq. (2.48))

$$(v_{RMS})_N = \left[\frac{\sum_{n=1}^N t_n v_n^2}{\sum_{n=1}^N t_n} \right]^{\frac{1}{2}} \quad t_n = \frac{\Delta z_n}{v_n} \text{ for each interval} \quad (3.20)$$

where $(v_{RMS})_N$ is the RMS velocity to the base of the Nth layer, t_n is the vertical one-way travel time in layer n, and v_n is the interval velocity in layer n as calculated using eq. (3.19).

Using the equations (3.15) to (3.20) the RMS velocity to the top of the capillary fringe can be determined for a range of plausible soil characteristics. From Table 3.2 we can see that for their textural class ‘sand’, Cosby et al. (1984) determined that $\log(\psi_s) = 0.84 \pm 0.56$ where ψ_s is measured in centimetres, $b = 2.79 \pm 1.38$, and $\theta_s \approx \phi = 33.9 \pm 7.3$. The water table depth for this model was chosen to be 4.0 m to approximate the average water table depth encountered at the study sites. At the surface, $z = 0$. The pressure head versus depth from the surface to the water table was assumed to be

$$\psi(z) = -4.0m + z \quad (3.21)$$

A step size of 0.02 m was used for the depth intervals in order to capture the capillary rise above the water table. A Matlab 5 function was created to calculate $\theta(z)$, $\epsilon_r(z)$, $v(z)$ and $t(z)$ for the mean values of ψ_s and b given above, and for one half and one standard deviation to either side of the mean. The saturated water content, θ_s , was assumed to be constant at $\theta_s = 33.9$. Varying θ_s by one standard deviation (± 7.3) had negligible effect on the final results. The results of this analysis are shown in fig. 3.2. Graphs A and B show the change in volumetric water content, θ , with pressure head, ψ , calculated using eq. (3.17). In fig. 3.2A, b is constant at its mean value of $\bar{b} = 2.79$, and ψ_s is varied by one standard deviation in either direction from the mean, $-0.0191 \text{ m} < \bar{\psi_s} = -0.0692 \text{ m} < -0.2512 \text{ m}$. In fig. 2.3B, ψ_s

Figure 3.2 Results of θ , ϵ_r , and v versus $\psi(z)$ for sand. Graphs A, C, and E show θ , ϵ_r , and v versus $\psi(z)$ when $b = 2.79$ and ψ_s varies in $\sigma/2$ steps from $\bar{\psi}_s - \sigma_{\psi_s} = -0.0191$ m to $\bar{\psi}_s + \sigma_{\psi_s} = -0.2512$ m. Graphs B, D, and F show θ , ϵ_r , and v versus $\psi(z)$ when $\psi_s = \bar{\psi}_s = -0.0692$ m and b varies in $\sigma/2$ steps from $b - \sigma_b = 1.41$ to $b + \sigma_b = 4.17$. The bold curves on each graph represent the results using mean values for both $b = \bar{b} = 2.79$ and $\psi_s = \bar{\psi}_s = -0.0692$ m.

is constant at $\overline{\psi_s} = -0.0692$ m, and b is varied by one standard deviation in either direction from the mean, $1.41 < \overline{b} = 2.79 < 4.17$. The bold curves indicate the result using mean values for both b and ψ_s . Figures 3.2C and 3.2D show the change in dielectric constant, ϵ_r , with pressure head, ψ , calculated using eq. (3.18). As in fig. 3.2A and 3.2B, fig. 3.2C has $\overline{b} = 2.79$ and ψ_s varying by one standard deviation on either side of the mean, while fig. 3.2D has $\overline{\psi_s} = -0.0692$ m and b varying by one standard deviation on either side of the mean. Finally, fig. 3.2E and 3.2F show the change in interval velocity, v , with pressure head, ψ , calculated using eq. (3.19).

The RMS velocity to the top of the capillary fringe, $(v_{\text{RMS}})_{\text{cap}}$, and the two-way travel time to the top of the capillary fringe, t_{cap} were calculated for each of the permutations of b and ψ_s described above. The calculated $(v_{\text{RMS}})_{\text{cap}}$ and t_{cap} versus b are shown in fig. 3.3. The lines indicate constant ψ_s values ranging from $\overline{\psi_s} - \sigma_{\psi_s}$ to $\overline{\psi_s} + \sigma_{\psi_s}$. From this graph, the maximum range in RMS velocities at the capillary fringe due to the range of ψ_s and b values for sand is from 0.0850 m/ns to 0.1638 m/ns. The RMS velocity calculated from the mean values for ψ_s and b is 0.1205 m/ns. Therefore, assuming that varying b and ψ_s by one standard deviation would account for the variation in moisture conditions at a single site, the RMS velocity could be expected to vary by up to 36 %. This is similar to the lateral velocity variations seen by Greaves et al. (1996) of up to 30 %. From the same analysis, the two-way time to the capillary fringe due to the range in ψ_s and b values for sand ranges from 48.87 ns to 88.65 ns. The t_{cap} calculated from the mean values for ψ_s and b is 65.81 ns. The two-way time to the capillary fringe has almost doubled as conditions change from low ψ_s and b to high ψ_s and b . Changing ψ_s and b affects the velocity in the near surface, as well as the actual depth to the top of the capillary fringe. The two effects can be separated out by keeping the capillary rise, or ψ_s , constant and only looking at the effect of changing b . This can be seen in fig. 3.3 where the bold lines show the change in $(v_{\text{RMS}})_{\text{cap}}$ and t_{cap} as b changes, while the capillary rise is held constant at its mean value of $\psi_s = \overline{\psi_s} = -0.0692$ m. The RMS velocity at the capillary fringe has changed by 0.0513 m/ns and the two-way time to the capillary fringe has changed by 26.04 ns due to a change in b of one standard deviation in either

Figure 3.3 Results of RMS velocity and two-way travel time to the capillary fringe versus b . In both graphs, the lines indicate constant values for ψ_s . The bold line indicates the mean value of $\psi_s = \bar{\psi}_s = -0.0692$ m. The other lines range in $\sigma/2$ steps from $\bar{\psi}_s - \sigma_{\psi_s} = -0.0191$ m to $\bar{\psi}_s + \sigma_{\psi_s} = -0.2512$ m.

direction from its mean of 2.79. As b increases, the velocity decreases and the two-way travel time increases. This indicates that lateral velocity variations caused by variations in grain size distribution, stratigraphic changes, or variable precipitation can significantly affect the two-way travel time to the capillary fringe even when the water table and capillary fringe are constant. Lateral changes in two-way time to the capillary fringe could be due to changes in the depth to the water table, changes in the height of the capillary fringe due to changes in ψ_s , or changes in the velocity due to changes in water content in the unsaturated zone.

The effect of lateral velocity variations on capillary fringe depth estimates from ground penetrating radar can be examined using the data from the preceding analysis. Assuming that the interval velocity profile determined using $\overline{\psi_s}$ and $\overline{b}$ was measured from a CMP gather, the range of error in the depth to the capillary fringe from using this velocity profile when ψ_s and b vary by one standard deviation from the mean values can be determined. The incorrect depths will be calculated using

$$(d_{cap})_{\psi_s, b} = \sum_{n=1}^M (t_{\psi_s, b})_n (v_{\overline{\psi_s}, \overline{b}})_n \quad (3.22)$$

where $(d_{cap})_{\psi_s, b}$ is the depth to the capillary fringe calculated using the incorrect velocity structure, $(t_{\psi_s, b})_n$ is the one way time through each interval n for a specific ψ_s and b , $(v_{\overline{\psi_s}, \overline{b}})_n$ is the interval velocity in each interval n for the mean values $\overline{\psi_s}$ and $\overline{b}$ and M is the interval above the top of the capillary fringe. This depth can be compared to the actual capillary fringe depth, given by

$$(d_{cap})_{true} = WTD + \psi_s = 4\text{ m} + \psi_s \quad (3.23)$$

where $(d_{cap})_{true}$ is the actual depth to the capillary fringe, and WTD is the water table depth which is equal to 4 m in this problem. The error in capillary fringe depth, $\epsilon_{d_{cap}}$, when velocity variations are not accounted for is then

$$\epsilon_{d_{cap}} = (d_{cap})_{\psi_s, b} - (d_{cap})_{true} \quad (3.24)$$

Figure 3.4 shows the error in depth, $\epsilon_{d_{cap}}$ versus b for b values ranging from $\bar{b} - \sigma_b$ to $\bar{b} + \sigma_b$. The lines indicate constant ψ_s values ranging from $\bar{\psi}_s - \sigma_{\psi_s}$ to $\bar{\psi}_s + \sigma_{\psi_s}$. For the simplest case, where ψ_s has remained equal to $\bar{\psi}_s$ and b has varied by one standard deviation from the mean, an error in depth to the capillary fringe of ± 0.8 m has been caused by using incorrect velocities. If both ψ_s and b vary, the depth error can be as great as -1.7 m. This shows that if the large velocity variations seen in this experiment exist at a field site and are not accounted for, large errors can result in depth estimates from ground penetrating radar.

Figure 3.4 The effect of lateral velocity variations on capillary fringe depth estimates. $\epsilon_{d_{cap}}$ is the error in depth caused by ignoring velocity variations caused by variations in ψ_s and b . The lines indicate constant values of ψ_s , where the bold line represents the depth errors when $\psi_s = \bar{\psi}_s = -0.0692$ m. the other lines range in $\sigma/2$ steps from $\bar{\psi}_s - \sigma_{\psi_s} = -0.0191$ m to $\bar{\psi}_s + \sigma_{\psi_s} = -0.2512$ m.

3.2.2.4 Vertical Resolution

The vertical resolution which can be achieved by a GPR signal also limits the accuracy with which the capillary fringe depth can be determined. The vertical resolution is the thickness of the beds or events that can be resolved and it is typically assumed to be one quarter of the central wavelength of the radar wave (Widess, 1973). This rule of thumb has developed from seismic wave theory, but can also be applied to GPR waves. The central wavelength of the GPR wave, λ_c , is dependent on the central frequency emitted by the antenna and the velocity of the radar wave in the subsurface. Figure 3.5 provides the relationship between velocity and vertical resolution for specific antenna frequencies which have been used at the three field sites. For reflections above the capillary fringe, the radar wave velocity is generally greater than 0.10 m/ns. Therefore, the vertical resolution is between 0.25 and 0.5 m when the 100

Figure 3.5 The dependence of vertical resolution on material velocity for 50, 100, and 200 MHz antenna frequencies (adapted from Jol, 1995).

MHz antennae are used. Similarly, the vertical resolution above the capillary fringe is between 0.125 and 0.25 m for the 200 MHz antennae. If two reflectors are closer than the vertical resolution in fig. 3.5, the two reflectors will not be distinguishable. Specifically, if another reflector lies within the vertical resolution of the top of the capillary fringe, distinguishing two separate reflections from one another on the resulting GPR section would be difficult if not impossible. We will see in a number of the GPR sections that cross-cutting reflections do interfere with the capillary fringe reflection, but are often distinguishable due to their dip. In areas where an over or underlying reflector is interfering with the capillary fringe reflection, the actual location of the capillary fringe must be estimated from the surrounding data.

3.2.2.5 Total Uncertainty in the Depth to the top of the Capillary Fringe

The uncertainty in the depth to the top of the capillary fringe, Δd_{cap} is composed of two uncertainties. The first is the uncertainty in the depth of the reflection due to the sampling interval of the data, Δd_{pick} . This value was shown to be $\Delta d_{pick} = \pm 0.02$ m. The second is the uncertainty in the position of the reflector due to uncertainty in the velocities determined from the CMP gathers, Δd_{vel} . This uncertainty ranges from $\Delta d_{vel} = \pm 0.05$ to 0.15 m depending on the dataset. These two uncertainties can be combined using

$$\Delta d_{cap} = \sqrt{\Delta d_{pick}^2 + \Delta d_{vel}^2} \quad (3.25)$$

to give the uncertainty in the depth to the top of the capillary fringe estimated from ground penetrating radar data. Velocity variations between the CMP gathers can add further error to the depth estimates from ground penetrating radar.

3.2.3 Capillary Fringe Height Estimation and Uncertainty

In order to convert the capillary fringe elevation from ground penetrating radar data into a

water table elevation, the height of the capillary rise above the water table must be estimated. Cosby et al. (1984) provide one possible way to make this estimation. As discussed above, the logarithm of the saturated pressure head, $\log \psi_s$, for sand samples in their dataset is $\log \psi_s = 0.84 \pm 0.56$. The saturated pressure head, ψ_s , is equivalent to the height of the capillary fringe above the water table under hydrostatic, vertical flow conditions. When the Cosby et al. (1984) data for sand is converted to ψ_s in metres to give an estimate for the capillary fringe height, h_{cap} for a sand has a mean value of 0.0692 m above the water table. One standard deviation above the mean is 0.2512 m and one standard deviation below the mean is 0.0191 m. The uncertainty in h_{cap} determined in this way can be expressed as one half the difference between these two values, giving $\Delta h_{cap} = \pm 0.1161$ m.

The capillary fringe height can also be estimated if the sedimentary grain size is known using eq. (2.54) from Chapter 2

$$h_c \approx \frac{0.15}{R} \text{ cm}^2 \approx \frac{0.15}{0.2 \cdot D} \text{ cm}^2 \quad (3.26)$$

where h_c is the height of capillary rise, R is the radius of the pore spaces and D is the grain diameter. The uncertainty in the capillary rise determined in this way would be dependent on how well the grain size of the sediments at the water table was known.

3.3 Sources of Uncertainty in Water Table Elevation Measurements from Monitoring Wells

Water table elevation measurements from monitoring wells are determined from the depth below a surveyed point, usually the top of inner casing. The water table elevation determined at a monitoring well, $WT_{elev}(\text{well})$ is equal to

$$WT_{elev}(\text{well}) = Z - WTD \quad (3.27)$$

where Z is the elevation of the top of the casing and WTD is the water table depth below that

point. Uncertainty in $WT_{elev}(well)$ is dependent on the uncertainty in the surveyed location and the water table depth measurement. The survey uncertainties discussed in section 3.2.1 for the GPR data apply to the well measurements as well, giving an uncertainty in elevation of $\Delta Z = \pm 0.005$ m for the Gilby and North Caicos Island monitoring wells. At the Strathroy study site, the monitoring wells were surveyed in a similar fashion, and can be expected to have a similar uncertainty. The uncertainty in the water table depth measurements is related to the instrument used. The standard water level meter has an uncertainty of $\Delta WTD = \pm 0.005$ m. Using the uncertainty propagation method discussed in section 3.1, the uncertainty in the water table elevation measurements from the monitoring wells is

$$\Delta WT_{elev}(well) = \sqrt{\Delta Z^2 + \Delta WTD^2} \quad (3.28)$$

Using $\Delta Z = \pm 0.005$ m and $\Delta WTD = \pm 0.005$ m in eq. (3.28), the uncertainty in the water table elevation measurements from a monitoring well is $\Delta WT_{elev}(well) = \pm 0.007$ m.

3.4 Summary

In this chapter, we have discussed the sources of error which lead to uncertainty in the water table elevations measured from monitoring wells and estimated from ground penetrating radar. The sources of error which affect the estimates from ground penetrating radar can be separated into three categories, uncertainty in the surface elevations along the GPR lines, uncertainty in the depth to the capillary fringe as measured from the GPR sections, and uncertainty in the capillary rise above the water table. We showed that the uncertainty in the surface elevation for surveyed locations at the Gilby Waterflood and North Caicos Island sites is $\Delta Z = \pm 0.005$ m. Surface elevations at the Strathroy site were estimated using kriging, which provides an uncertainty in surface elevations along the GPR lines of $\Delta Z = \pm 0.037$ m. We then investigated the uncertainty in surface elevation which would be introduced by linearly interpolating elevations between the surveyed locations. We showed that for the Gilby Waterflood and North Caicos Island sites, the uncertainty due to linear interpolation

of the elevations was negligible compared to the other sources of error.

Uncertainty in the depth to the capillary fringe from ground penetrating radar sections is dependent on a number of sources of error. GPR data is sampled at a specific time interval, which provides a minimum uncertainty in any depth pick of $\Delta d_{\text{pick}} = \pm 0.02$ m. In order to convert two-way travel time into depth, the velocity of the radar wave in the subsurface must be determined from common midpoint gathers. The uncertainty in RMS velocities determined from CMP gathers was estimated to be $\Delta(v_{\text{RMS}})_{100 \text{ MHz}} = \pm 0.003$ m/ns for the 100 MHz CMP gathers, and $\Delta(v_{\text{RMS}})_{200 \text{ MHz}} = \pm 0.0025$ m/ns for the 200 MHz CMP gathers. The uncertainty in the RMS velocities is converted to uncertainty in depth by comparing the depth sections using the actual picked velocities to depth sections created using velocities equal to $(v_{\text{RMS}})_{\text{actual}} - \Delta v_{\text{RMS}}$ and $(v_{\text{RMS}})_{\text{actual}} + \Delta v_{\text{RMS}}$. The uncertainty in the depth to the capillary fringe due to uncertainty in the velocity picks from a CMP are $\Delta d_{\text{vel}} = \pm 0.05$ m to ± 0.15 m depending on the depth section.

Variations in velocity along a GPR section can cause additional uncertainty in depth to the capillary fringe due to the sparsity of CMP gathers compared to the common offset stations. As it is not possible to determine the exact error in depth which would be caused by neglecting velocity variations, we examined the theoretical variation in velocities which might be expected at a particular site. Cosby et al. (1984) provided a mean and standard deviation for the saturated pressure head, ψ_s , and pore distribution index, b determined from sand samples, which could be used in eq. (3.15) to relate pressure head (ψ) and volumetric water content (θ). Assuming a water table depth of 4 m and that one standard deviation of variation in ψ_s and b described the expected variability in water content in the unsaturated zone, we determined that a 36 % variation in RMS velocity to the capillary fringe would result. If this variation in velocity were ignored, the resulting error in depth would be as high as ± 0.8 m if the capillary fringe height, or ψ_s , remained constant, rising to an error in depth of ± 1.7 m if the capillary fringe height changed along with changes in b . This analysis showed that velocity variations which are unaccounted for due to sparsely spaced CMP

gathers can cause large errors in depth estimates from ground penetrating radar.

In order to convert the capillary fringe elevation estimates from ground penetrating radar to water table elevations, the height of the capillary rise above the water table must be estimated. We identified two methods for determining the capillary rise. In the first, the value for ψ_s for sand from Cosby et al. (1984), which is equivalent to capillary rise, can be used to provide h_{cap} and Δh_{cap} . The second method uses known grainsize distributions at the water table to determine the theoretical capillary rise which could be expected. The uncertainty in h_{cap} determined from this method would depend on how well the grainsize distribution was known, and would tend to be more uncertain for smaller grain sizes, due to the larger capillary rise which results. Interpolating between known points can add errors on the order of 0.10 to 0.15 m in sandy, heterogeneous areas, as seen by the standard deviation of the saturated pressure head measurements for sand of approximately 0.1161 m from Cosby et al. (1984).

Water table elevations measured from monitoring wells have two sources of uncertainty, survey elevations and water table depths. Uncertainty in surveyed elevations, Z are the same for the monitoring wells as for surveyed GPR stations. Therefore, $\Delta Z = \pm 0.005$ m for all three datasets. Uncertainty in the measured water table depth, WTD is dependent on the water level meter used. In our case, $\Delta WTD = \pm 0.005$ m. Using the uncertainty propagation method discussed in section 3.1, the uncertainty in water table elevations measured from monitoring wells is $\Delta WT_{elev}(well) = \pm 0.007$ m.

4. Data Processing of Ground Penetrating Radar

One of the important advantages of ground penetrating radar in its use in engineering and environmental investigations is the ability to see and interpret the data as it is being collected. However, due to similarities between reflection seismic and ground penetrating radar data, many of the processing techniques and packages developed for use with seismic data can be applied and used to an advantage with ground penetrating radar data. Examples of this include Fisher et al. (1992a), Fisher et al. (1992b), Huggenberger et al. (1994), Young et al. (1995), Fisher et al. (1996), Greaves et al. (1996), Harari (1996), and Arcone et al. (1998). Processing techniques include gain application, bandpass filtering, normal moveout correction, migration, spectral analysis and velocity analysis.

4.1 General Processing Techniques

All three datasets discussed in this thesis were processed in the same general manner. Before the data was exported from the PulseEKKO IV data collection software, a signal saturation correction was applied to the data. The initial air and direct ground wave arrivals are very strong compared to reflection returns, and saturate the receiver. The saturation correction or de-wow function applies a DC correction to remove the effects of this initial saturation. The data was then converted into seg-y format and imported into ProMAX.

The first step in processing the GPR data was to import geometry information into ProMAX and create geometry databases for all of the data. Where appropriate, data files were merged to speed up processing. This mainly involved merging CMPs collected along the same line into one file to facilitate velocity analysis. When GPR data is collected, there is a delay time prior to the arrival of the air wave which is dependent on the length of the optical fibre. In order to perform accurate velocity analysis and depth calculations, the data needs to be shifted to its proper zero time. For a CMP, this involved picking the first breaks on the raw data, determining the zero offset time of the airwave arrival and shifting each trace by that

Figure 4.1 Seven traces from Gilby Waterflood Line 2 - 100 MHz data. The artifact prior to the airwave, the airwave first break and the capillary fringe reflection are indicated. Dispersion of the GPR wavelet is seen in water table reflection.

amount. For a common offset profile, each trace was shifted so that the first break time for each trace was equal to $t_{\text{shift}} = x/c$, where x is the source-receiver offset and c is the speed of light. This time, t_{shift} is the time it takes the airwave to travel from the transmitter to the receiver. For instance, if the source-receiver offset is 1.0 m, then the airwave should arrive at the receiver at $t = 1.0 \text{ m} / 0.3 \text{ m/ns} = 3.33 \text{ ns}$ after the wave was generated. This positions the rest of the trace at its correct time relative to the initial transmission.

Determining the first break on radar data can sometimes cause some confusion. Due to the initial signal saturation and subsequent correction, GPR data tend to have an initial artifact prior to the rise of the airwave, as seen in fig. 4.1. If the first break is chosen as the beginning of the artifact instead of the beginning of the airwave, problems arise in the interpretation of the polarity of reflections compared to the airwave. First breaks were chosen as the zero crossing between the artifact and initial airwave rise on all the data. In Chapter 2, we determined that the polarity of the capillary fringe reflection will be opposite to the airwave. Therefore the first arrival of the capillary fringe reflection will be the zero crossing prior to the positive half cycle, as seen in fig. 4.1.

Spectral analysis of the data was completed to determine the most appropriate bandpass filter, and to help identify any areas where problems with signal quality might be suspected. The GPR antennae had central frequencies of 50, 100, and 200 MHz with a bandwidth approximately equal to the central frequency. High frequencies are attenuated more rapidly than low frequencies, so the recorded spectrum tends to be greatly reduced from the output wavelet's bandwidth. From this point, processing streams for the CMP gathers and the common offset profiles differ, so they will be discussed separately.

4.2 Common Mid-point Gathers

The common mid-point gathers were processed with the goal of extracting as much velocity information as possible from them. Time zero adjustment was accomplished by picking the first breaks and exporting that information to Matlab. A best fit line was determined from the data, giving the unshifted time for the first breaks at zero offset. For some of the data, positive and negative offset data was collected, and in this case, the average of the first break time at zero offset was determined. Each trace was shifted up in time by t_{shift} equal to the zero offset first break time. The direct wave was also picked and best fit lines were used to determine the very near-surface radar wave velocity. Figure 4.2 shows the best fit lines for the airwave and direct wave picks for the CMP gather at 85 m on line 1 - 200 MHz. The

Figure 4.2 Airwave (+) and direct wave (x) picks for a CMP at 85 m on line1, Gilby Waterflood. The best fit lines for each are shown, as well as the zero offset time (*).

direct wave velocity will be used in velocity analysis, NMO correction, migration and time-to-depth conversion as the very near-surface velocity.

After spectral analysis was complete, the CMP gathers were processed using the flow in fig. 4.3 prior to semblance velocity analysis. The processing techniques in fig. 4.3 are discussed in more depth below. Velocity analysis was performed as outlined in Section 2.3.1. Figure 2.5 provides an example of the semblance velocity analysis. The resulting RMS velocity function for each CMP includes the direct wave velocity at $t = 0 \text{ ns}$, and a final RMS velocity of between 0.070 and 0.090 m/ns at $t = 300 \text{ ns}$. This final velocity was picked in order to reduce velocity induced artifacts below the capillary fringe. These artifacts resulted from a

Figure 4.3 Processing flow for common midpoint gathers.

decreased ability to pick reflections below the capillary fringe. RMS velocities between 0.070 to 0.090 m/ns are a good approximation of the RMS velocity of a wave travelling for some distance in a saturated sand, after travelling through relatively unsaturated sediments. The specific value chosen for each CMP depended on the velocities chosen from the hyperbolic reflectors. The RMS velocity values were transferred to the common offset line processing stream and used for NMO correction, and spherical gain corrections. The RMS velocity functions were also converted into interval velocities using the Dix equation (eq. (2.51)), and used in time migrations and time-to-depth conversions.

4.3 Common Offset Lines

The common offset lines were processed with the goal of accurately identifying the capillary fringe reflection and determining the depth to the capillary fringe. A flow diagram of the processing stream is provided in fig. 4.4. The first breaks or airwave arrivals were picked, and shifted to $t_{\text{shift}} = x/c$, as outlined in Section 4.1.

The next step was to apply normal moveout (NMO) correction and spherical and exponential gains using RMS velocity functions developed from the CMP gathers. Normal moveout corrections were discussed in Chapters 2 and 3 as a velocity analysis method for CMP gathers. NMO correction removes the slight time shift due to antenna offset, placing the traces at zero offset. Spherical and exponential gain applies a time variant gain function

Figure 4.4 Processing flow for GPR lines.

which attempts to compensate for amplitude losses due to wavefront spreading and inelastic attenuation. This processing step was followed by Q compensation, which attempts to compensate for the dispersion of the wavelet as it travels through the subsurface. This dispersion is seen in fig. 4.1 in the capillary fringe reflection. In this case, the ground surface was rising, causing the capillary fringe reflection to occur deeper in the time section. As the time increases, a minor increase in the width of the reflection, as well a decrease in the amplitude of the reflection are an indication of dispersion. In fig. 4.1, spherical and exponential gains and an Ormsby bandpass filter have been applied. Turner and Siggins (1994) and Bano (1996) show similar examples of attenuation and dispersion of GPR waves. Both papers demonstrate that in some materials, dispersion effects can be approximated by a constant Q model similar to the constant Q attenuation found in seismic waves. Constant Q compensation appears to remove some of the dispersion effects present in the GPR data, as shown in fig. 4.5. The traces are the same as those in fig. 4.1, however constant Q compensation has been applied after the spherical and exponential gain and before the bandpass filter. The amplitude

Figure 4.5 Seven traces from Gilby Waterflood Line 2 - 100 MHz data. The artifact prior to the airwave, the airwave first break and the capillary fringe reflection are indicated. These are the same traces as in fig. 4.1, however constant Q compensation has been applied (see text for details).

of most of the reflectors has been greatly enhanced, and the dispersive effects have been reduced. The effect of constant Q compensation on the frequency of the data is illustrated in fig. 4.6. Figure 4.6A shows the frequency spectrum above and below the capillary fringe reflector, prior to Q compensation and bandpass filtering. Figure 4.6B shows the frequency spectrum above and below the capillary fringe reflector following Q compensation and bandpass filtering. The high end frequencies have been enhanced above and below the capillary fringe by the constant Q compensation function. The spectra has been split into data

Figure 4.6 Spectra from data above and below the capillary fringe on line 2 - 100 MHz from Gilby Waterflood. **A** Line 2 has been processed with spherical and exponential gain and a bandpass filter (5-20-150-180 MHz). **B** Line 2 has been processed with spherical and exponential gain, constant Q compensation, and a bandpass filter (5-20-150-180 MHz).

above and below the capillary fringe because reflector amplitude and bandwidth is often greatly reduced below the capillary fringe reflector. This is seen in both fig. 4.1 and 4.6A. Q compensation has helped to remove dispersive effects from the data and increase the amplitude levels above and below the capillary fringe.

The next step in the common offset line processing stream was to apply a bandpass filter based on the results of the spectral analysis completed at the beginning of processing. An Ormsby minimum phase bandpass filter was used on all the datasets. Bandpass filtering was followed by common midpoint stacking to allow post-stack processes such as migration and time-to-depth conversion to be applied. As common offset profiles are single fold, stacking has no effect of the data.

Steep dip explicit finite difference time migration was used to migrate the reflectors to their proper spatial location. This was followed by a time-to-depth conversion using the same velocity function. According to Yilmaz (1987), depth migration is only necessary when the stacked section contains structural dip and large lateral velocity gradients. Time migration

is suitable for mild lateral velocity variations. While lateral velocity variations did occur in the various datasets, they generally did not exceed 10 % on any one line, so a time migration was deemed suitable. Interval velocities in time were used to complete the time-to-depth conversion.

Finally, elevation corrections were applied to each of the depth sections. In seismic reflection processing, elevation corrections are performed prior to the NMO correction as a time correction. The amount of correction applied is determined from surface elevation data and near surface velocity(s), which are used to fill in or strip of topography depending on the location of the source with respect to a final datum. This process did not seem to provide a final depth corrected section which had the proper topography. The solution to this problem was to shift the data to the correct elevation on the depth section, removing the possibility of introducing error through the use of near surface velocities. As well, Lehmann et al. (1998) show that for GPR data in which the target depth is on the same order as the elevation variations, applying static corrections prior to migration can cause the dip of the target to be incorrect. Applying static corrections after migration produces the correct dip in the target reflector, but may not correctly locate it. Their research suggests that topographic migration, where source and receiver elevation are incorporated into the migration algorithm, reproduces the true dip and location for a target reflector. Topographic migration was not performed on the data in this thesis, as it was not available in ProMAX 6.2 and the capillary fringe target was deeper than the elevation variations along any one line.

4.4 Deconvolution

One of the key processing techniques used in seismic data processing is deconvolution. Deconvolution is performed to increase temporal resolution by compressing the basic seismic wavelet to approximately a spike and suppressing reverberating wave trains. This involves whitening the frequency spectrum to correct for the more rapid attenuation of higher frequencies. A white spectrum has constant power at all frequencies. There are a few

examples of successful applications of deconvolution to GPR data (Huggenberger et al., 1994, Fisher et al., 1996, Arcone et al., 1998). However, success with deconvolution is very dataset dependent. For instance, Fisher et al. (1992a) state that deconvolution was not performed because the “radar source already produces a compact wavelet”. In the same paper, most standard seismic processing techniques were used to process one of the first large, multi-offset GPR datasets.

Deconvolution was not applied in the data processing, because it often degraded the quality of the results. Deconvolution algorithms are based on a series of simplifying assumptions commonly referred to as the convolution model (Margrave, 1996). The result of these assumptions is that the seismic or GPR data to be deconvolved is expected to be stationary. A stationary series is one for which the main spectral features do not change in time. Q compensation increases the high frequency content of the data deeper in the section without decreasing the data quality. Since constant Q compensation has whitened the spectrum of the data and improved interpretability (fig. 4.6), it was used in the processing stream and deconvolution was deemed unnecessary.

4.5 Summary

This chapter outlines the processing techniques used on the ground penetrating radar data. Constant Q compensation was used to amplify the high end frequencies throughout the GPR sections, removing some of the dispersion effects commonly seen in GPR data. A steep dip explicit finite difference time migration was chosen to move the reflections to their proper spatial location using velocity information obtained from common midpoint gathers. After converting the GPR sections to depth, elevation corrections were applied to ensure the depth sections had the proper topography.

5 Gilby Waterflood, Alberta

5.1 Study Site

The Gilby Waterflood site was our initial study location, and contains our most complete set of data. The Gilby site is located on the west bank of the Red Deer River, 3 km west of Red Deer, Alberta. This site has been extensively modified by aggregate extraction activities which have removed 10 to 15 m of overburden. In general, the shallow subsurface consists of 10 to 15 m of well-graded fine-to-coarse sands and fine-to-coarse gravels with sands (Wills, 1993). Minor silt and clay deposits are also found throughout the site, and stockpiles of topsoil up to 10 m thick overlie the sands and gravels in places. The bedrock consists of a weathered siltstone to fine sandstone of the Paskapoo formation. The upper 30 cm of bedrock is extremely weathered, resulting in its appearance as a clay rather than a consolidated media. The hydrogeology of the site consists of a local unconfined aquifer in the sand and gravel which is connected to the Red Deer River. The water table lies between 3 and 6 m below the ground surface across the site. Figure 5.1 shows a site map, including monitoring wells, GPR common offset lines, and CMP locations. Subsurface data is provided by four existing monitoring wells, installed to determine the effect of a series of pumping wells on the local shallow unconfined aquifer. The close proximity of GPR lines and wells allow correlations between the GPR data and the subsurface.

5.2 Dataset

The Gilby dataset contains GPR common offset lines, common midpoint gathers, apparent electrical conductivities, water table elevations, and survey data. The ground penetrating radar, survey and water table elevation data were collected from the ninth to the eighteenth of June, 1997. Late spring weather was encountered, which generally consisted of a relatively clear morning and early afternoon, with late afternoon thundershower buildup. There were a number of occasions when data collection had to be halted in the late afternoon due to an

Figure 5.1 Site map for the Gilby Waterflood site. The locations of GPR common-offset profiles, CMP gathers, and monitoring wells are indicated. The limit of the gravel pit excavation, location of the spill pile and the Red Deer River are also indicated. The gravel pit excavation and spill pile rise up to 10 m above the ground surface in the pit at the GPR lines.

approaching thunderstorm. Light rainfall was frequently noted during the evening. Under these conditions, moisture values in the unsaturated zone are expected to be high.

Water table elevations were collected from the four monitoring wells at least twice a day.

Figure 5.2 shows water table elevation data collected from each of the four wells over the 10 days of data collection. For the first five days, the water table elevation remained relatively constant. On the night of June 13, 1997, a heavy rain was noted, and a 32.5 cm rise in river level was observed between 15:58 on June 13 and 08:16 on June 14. Water table elevations continued to rise, stabilizing by June 18. Due to the loss of river stakes during data collection, only a qualitative description of the Red Deer River levels is available.

The GPR data consists of four main GPR lines ranging in length from 48 to 90 m, plus seven connecting or cross lines of between 6 and 34 m in length. Common midpoint gathers were generally collected every 10 m along the common offset lines. For each line, 100 and 200 MHz central frequency antennas were used. Table 5.1 shows the length, number of CMP gathers collected, and collection date for each of the common offset lines, separated by frequency. The 100 MHz data was collected from June 10 to 15, while the 200 MHz data was collected from June 16 to 18, 1997. Due to the rise in water table elevations seen in fig. 5.2,

Figure 5.2 Water table elevations collected from four monitoring wells at the Gilby Waterflood site.

Table 5.1 Acquisition date, length, and number of CMP gathers collected along each of the common offset GPR lines collected at the Gilby Waterflood site.

	100 MHz			200 MHz		
	date	length (m)	no. of CMP's	date	length (m)	no. of CMP's
line 1	06/10/97 12:30	90	8	06/17/97 15:00 06/18/97 09:30	91 91	4 10
line 2	06/14/97 09:00	59	7	06/17/97 11:00	70	7
line 3	06/14/97 14:00	50	5	06/17/97 09:00	50	5
line 4	06/15/97 10:00	65	6	06/16/97 16:00	56	4
xline 10	06/11/97 14:20	6	0	06/18/97 13:40	6	1
xline 20	06/11/97 14:00	29	1	06/18/97 13:55	30	3
xline 30	06/11/97 15:00	27	2	06/18/97 15:00	28	2
xline 40	06/12/97 10:30	21	1	06/18/97 15:20	22	2
	06/13/97 10:00	21	2			
xline 50	06/13/97 11:00	21	2	06/17/97 17:00	22	2
xline 70	06/13/97 13:00	10	1	06/17/97 18:20	10	1
xline 85	06/13/97 14:00	34	3	06/17/97 16:00	34	4

the capillary fringe elevations from the 200 MHz data should be higher than the 100 MHz data. Data collection parameters for each frequency are shown in Table 5.2.

Each line was surveyed at intervals between 1 and 2.5 m. Figure 5.3 provides a surface elevation contour map for the site. The survey data locations used to create the surface contours are also indicated on the map.

In June 1999, an EM-31 electromagnetic induction system was used to collect apparent conductivity data along the approximate location of the GPR profiles. Additional excavation had occurred in the area of line 1 and cross lines 10 through 50. The EM-31 data was

Table 5.2 GPR data collection parameters for the Gilby Waterflood site.

	100 MHz	200 MHz
common offset antenna separation	1.00 m	0.75 m
common offset station spacing	0.25 m	0.25 m
CMP step size	0.10 m	0.10 m
CMP maximum offset	10.00 m	10.00 m
sampling interval	0.8 ns	0.8 ns
number of stacks	128	128
pulser voltage	400 V	400 V

collected for horizontal and vertical pole alignments with 1 m station spacings along all existing lines. The monitoring wells used to collect water table elevations had been removed. The water level in the Red Deer River was approximately the same as during the 1997 data acquisition program.

5.3 Processing Characteristics

The Gilby Waterflood data were processed using the processing flows discussed in Chapter 4. Table 5.3 gives the actual processing parameters for the 100 MHz and 200 MHz data. As shown in Table 5.2, the 100 MHz common-offset profiles had an antenna separation of 1.0 m, while the 200 MHz profiles had a separation of 0.75 m. Therefore the airwave picks were flattened to 3.3 and 2.5 ns respectively. The Q Compensation values for both frequencies were picked by experimentation, as were the bandpass filter ramps. The maximum frequency to migrate in the time migration was based on the high pass limit of the bandpass filter. Elevation corrections were applied by determining the difference in elevation between each station and a datum of 866.5 m above sea level (a.s.l.). This is the maximum elevation at which GPR data was collected.

Table 5.3 Processing Parameters for the Gilby Waterflood GPR data.

	100 MHz	200 MHz
Airwave Flattening *	$t = 3.3 \text{ ns}$	$t = 2.5 \text{ ns}$
Q Compensation	$Q = 30$	$Q = 40$
Ormsby Bandpass filter	5-20-150-180 MHz	20-40-260-280 MHz
Steep Dip Explicit Finite Difference Time Migration *	maximum frequency = 150 MHz maximum dip = 70°	maximum frequency = 260 MHz maximum dip = 70°
Elevation correction *	Datum = 866.50 m a.s.l.	Datum = 866.50 m a.s.l.

* processing parameters which were applied only to the common-offset data

5.4 Results

5.4.1 Common Midpoint Gathers

Thirty-eight 100 MHz and forty-five 200 MHz common midpoint gathers were collected at the Gilby Waterflood site. A few of the locations were in poor data areas, resulting in little or no velocity information. Direct wave velocities were determined for all the CMP gathers. The mean and standard deviation of the direct wave velocities is $0.125 \pm 0.008 \text{ m/ns}$ with minimum and maximum values of 0.108 and 0.152 m/ns. The direct wave velocities are remarkably similar for the entire dataset, considering the change in surface conditions throughout data collection. A typical field day began with wet ground conditions due to overnight rains. By midday, the ground had usually dried considerably, and by midafternoon on a sunny day, the ground was dry to the touch. The direct wave velocities do not show the variation expected from the change in surface conditions. This indicates that the direct wave velocity samples the subsurface at a depth which is less affected by precipitation events.

Figures 5.4 and 5.5 show the RMS velocity sections along lines 2, 3 and 4 for the 100 and 200 MHz antennae, respectively. The RMS velocity profiles for line 1 did not contour well

Figure 5.4 RMS velocity profiles for the 100 MHz CMP gathers from **A** line 2, **B** line 3, and **C** line 4.

Due to the lack of data from the CMP gathers collected between 40 and 60 m. Figures 5.6 and 5.7 provide line plots of RMS velocity versus time for line 1 and the cross lines at 100 and 200 MHz central frequencies. Examining fig. 5.4 to 5.7, we can see some general trends which are similar for all of the data. The RMS velocities tend to be relatively high at the surface, decrease somewhat below the surface, then increase at the capillary fringe reflection. The velocities then decrease below the water table. As mentioned in Chapter 4, there is very

Figure 5.5 RMS velocity profiles for the 200 MHz CMP gathers from A line 2, B line 3, and C line 4.

little data below the water table. The actual velocity of the radar wave decreases once the wave is within a saturated medium. For instance, the velocity in saturated sand is 0.06 m/ns versus 0.15 m/ns for dry sand (see Table 2.1). The RMS velocity will not immediately fall to this value, but will gradually decrease as the total amount of time within the saturated medium increases. Based on the velocities determined from hyperbolic curves above 200 ns, an RMS velocity at 300 ns was chosen to continue the decreasing trend in RMS velocity with

Figure 5.6 RMS velocity profiles for the 100 MHz CMP gathers from A line 1, and B the cross lines.

time. In determining this RMS velocity, the interval velocity which would result from the picked values had to be considered. The Dix interval velocity equation (formerly eq. (2.51)) is used to convert RMS or stacking velocities into interval velocities,

$$v_N^2 = \frac{t_N v_{RMS\ N}^2 - t_{N-1} v_{RMS\ N-1}^2}{t_N - t_{N-1}} \quad (5.1)$$

Interval velocities are used to convert two way times into depths. The right side of eq. (5.1) must be positive to give a real value for v_N . Consequently, changes in v_{RMS} must honour

$$v_{RMS\ N} > v_{RMS\ N-1} \sqrt{\frac{t_{N-1}}{t_N}} \quad (5.2)$$

The constraint eq. (5.2) places on the decrease in v_{RMS} with time had to be kept in mind during velocity analysis, as a real value for v_N is necessary for time to depth conversion. This is especially true below the water table, where the RMS velocities are always decreasing.

One example where the constraint on the decrease in RMS velocity caused problems during velocity analysis is seen on line 4 (fig. 5.9). Deep in the section at the east (50 m) end of the line, a relatively strong reflection was found. The reflector's velocity was very low, consistent with a clay or silt layer. In the CMP gathers collected at 40 and 50 m on line 4, it was impossible to pick the first reflection which exhibited the low velocities below the water table, as the Dix equation would have resulted in imaginary interval velocities. In this case, the earliest time which would result in positive interval velocities was chosen. Tillard and Dubois (1995) and Greaves et al. (1996) indicate that they also experienced similar problems using the Dix equation with decreasing RMS velocities.

Figures 5.4 to 5.7 show that there is considerable range in the RMS velocities through the sections. The higher RMS velocities which occur between 50 and 100 ns are from the capillary fringe reflection. Above the capillary fringe, the RMS velocities range from 0.110 to 0.135 m/ns. Analysis of the range of RMS velocities measured at the capillary fringe can be used to judge the homogeneity of the site. Table 5.4 provides the mean RMS velocity to

Figure 5.7 RMS velocity profiles for the 200 MHz CMP gathers from A line 1, and B the cross lines.

the top of the capillary fringe reflection for each of the lines, and for the entire dataset. The mean and standard deviation of the RMS velocity at the capillary fringe from all the CMP gathers is 0.129 ± 0.004 m/ns. The maximum percent difference from the mean RMS velocity is 9.5 %. This can be compared to the results in Chapter 3, where a 36% variation in the RMS velocity at the top of the capillary fringe was calculated due to presumed

Table 5.4 Mean, $\overline{v_{RMS}}$ and standard deviation, $\sigma_{v_{RMS}}$ of the RMS velocity at the capillary fringe for various lines at the Gilby Waterflood site. The percent difference is calculated as the difference from the mean of the maximum and minimum values in each set of RMS velocities.

	100 MHz			200 MHz		
	no. of CMPs	$\overline{v_{RMS}} \pm \sigma_{v_{RMS}}$	% difference	no. of CMPs	$\overline{v_{RMS}} \pm \sigma_{v_{RMS}}$	% difference
line 1	5	0.129 ± 0.004	4.5 % 3.4 %	6	0.127 ± 0.004	4.4 % 3.0 %
line 2	6	0.130 ± 0.004	3.9 % 4.9 %	7	0.128 ± 0.004	4.0 % 5.5 %
line 3	5	0.127 ± 0.005	3.4 % 5.3 %	5	0.129 ± 0.003	3.2 % 3.3 %
line 4	6	0.126 ± 0.003	2.9 % 2.5 %	4	0.127 ± 0.001	1.0 % 1.1 %
all CMPs at one freq.	31	0.129 ± 0.005	8.2 % 6.6 %	34	0.129 ± 0.004	9.5 % 5.8 %
ALL DATA	65	0.129 ± 0.004	9.5 % 6.6 %			

variations in the saturated hydraulic head, ψ_s , and the pore size distribution index, b . The calculated variation was the maximum expected variation in RMS velocity at the capillary fringe given that ψ_s and b varied by one standard deviation from the mean values reported in Cosby et al. (1984). The RMS velocity to the capillary fringe using the mean values for ψ_s and b was calculated to be $v_{rms} = 0.120$ m/ns. Considering the variability in moisture content which can occur in the vadose zone, the experimental and theoretical values are consistent.

Another way to examine the velocity variations which occur at this site is to examine the statistical and spatial distribution of the measured velocities. As the interval velocities are used to convert the two-way time sections to depth, the average interval velocity to the capillary fringe for each CMP gather will be used to describe the spatial variation in

velocities. The average interval velocity to the capillary fringe is equivalent to the weighted harmonic mean velocity to the capillary fringe, $v_{H\text{cap}}$.

$$v_{H\text{cap}} = \frac{\sum_{i=1}^N d_i}{\sum_{i=1}^N \frac{d_i}{v_i}} \quad (5.3)$$

In eq. (5.3), d_i is the length of the interval i , v_i is the interval velocity in interval i , and N is the layer above the capillary fringe reflection. Figure 5.8A shows that the $v_{H\text{cap}}$ values for the 200 MHz data have a normal distribution, with a mean and standard deviation of 0.128 ± 0.003 m/ns using 33 values. It should be noted that the $v_{H\text{cap}}$ determined for the CMP gather collected at 6 m on cross line 40 appeared to be an outlier, and was removed from the analysis. In geostatistics, the spatial distribution of the data is defined by a semivariogram which graphs the difference in data values defined by the semivariogram function, $\gamma(|h|)$ versus the distance between the data values, or lag ($|h|$). The semivariogram function, $\gamma(|h|)$ is defined as (Isaaks and Srivastava, 1989)

$$\gamma(|h|) = \frac{1}{2N(|h|)} \sum_{i=1}^{N(|h|)} [z(x_i) - z(x_i + h)]^2 \quad (5.4)$$

where h is the lag, or separation between the data points, $z(x_i)$ and $z(x_i + h)$ are the data values, and $N(|h|)$ is the number of data pairs used in the summation. Figure 5.8B is the semivariogram which describes the spatial variation in $v_{H\text{cap}}$. The experimental semivariogram values appear as circles, with the number of data pairs making up the value for $\gamma(|h|)$ indicated above each point. As there are only 33 data points available, the experimental semivariogram is not ideal. A semivariogram model is generally matched to the experimental semivariogram to create a continuous function describing the spatial variation of the data. The spherical model chosen in fig. 7.8B gives three parameters

Figure 5.8 The statistical and spatial distribution of the harmonic mean velocity to the capillary fringe (200 MHz data). **A** Histogram showing the normal distribution of the $v_{H\text{cap}}$ velocities. **B** Experimental and model semivariograms for the $v_{H\text{cap}}$ velocities. The spherical model gives a range of 20 m for the maximum correlation length of the data.

describing the model, which could be used in a kriging algorithm to interpolate between the data points. The range of a model defines the correlation length of the data. A range of 20 m was determined for the model semivariogram, implying that the v_{Hcap} velocities are correlated to a maximum distance of 20 m. In order to capture the spatial variation of the velocity, the CMP gathers would have to be collected more frequently than 20 m. Half the correlation length is probably sufficient to capture the variations. Therefore, we can say that we have captured the velocity variations at this site with CMP gathers collected every 10 m along the common offset lines.

5.4.2 Common Offset Lines

The locations of the common offset profiles collected at the Gilby Waterflood site are shown in fig. 5.1. The lines were positioned so that GPR data was collected near three of the four wells. On an earlier feasibility study, we experienced poor GPR data quality near well MW3 at the northern end of the site, so GPR data was not collected there. The well was monitored throughout data collection to provide additional information on water table fluctuations. The position of the lines was also dependent on the ground cover. Again on an earlier study, we discovered that areas which were more densely vegetated tended to have little or no GPR reflections. This was also seen in various sections of the data, especially along line 1 and cross lines 10 through 50. We believe that the vegetation density increased in areas where the clay content in the near surface increased. The clay increased the water retention of the soil enough for vegetation to grow. For this reason, more heavily vegetated areas of the site were avoided during data collection.

Line 1 and its cross lines were collected in an attempt to collect a 3D dataset. Daniels et al. (1997) lay out some field procedures which are needed for three-dimensional GPR data. According to Daniels et al. (1997) the spacing between lines must follow the same rules which are used to avoid spatial aliasing of the data along the lines. Therefore, the line spacing must be less than one quarter of the wavelength of the radar wave in the host material to

avoid spatially aliasing the data. If this spatial sampling interval is exceeded, steeply dipping events and diffraction tails can become difficult to interpret. As well, reflectors can have different characteristics with different antenna orientations. For all the GPR data in this thesis, the antennae were oriented perpendicular to the collection line. Therefore, there is a 90° difference in antennae orientation between line 1 and the cross lines. Since the water table is a relatively flat and near horizontal surface, antennae orientation should not result in a drastic change in reflector characteristics, but it is a potential problem. As well, with a distance of 10 m between the cross lines, it would be very difficult to correlate reflectors other than the water table between lines. While line 1 and the cross lines can not be considered a 3D dataset, they do provide useful information on the northwest side of the line 1. Just to the southeast of line 1 data quality decreases rapidly.

The cross line at 85 m on line 1 was used to tie line 1 and line 2. Line 2 passes by monitoring well MW 1 at 22 m. Line 4 passes by monitoring well MW 5 at 52 m and line 3 is used to tie lines 2 and 4. Figure 5.9 shows the time migrated sections for the 100 and 200 MHz common-offset profiles along line 4. The capillary fringe reflection is the one and a half cycle reflection with its first positive rise between 60 and 70 ns for the majority of the line. At the east end (0 m) of the 100 MHz profile (fig. 5.9A), interference between the first peak of the capillary fringe reflection and a reflector above it has distorted the actual location of the first half-cycle of the capillary fringe reflection. This interference is greatly reduced in the 200 MHz profile. At the west end (42 to 57 m), the capillary fringe reflection gradually gets deeper in time. Figure 5.10 shows the final depth sections, which have been converted to depth and corrected for elevation changes. In both the 100 and 200 MHz depth sections, the capillary fringe reflection appears relatively flat, showing that elevation changes were the cause of the change in time of the capillary fringe reflection in fig. 5.9. There are a few areas where the first arrival of the capillary fringe reflection on the time and depth sections is not well imaged. These areas are more obvious in the 200 MHz data, especially between 15 to 19 m and 26 to 31 m. Between 26 and 31 m the profile crosses a polyethylene pipe with magnetic tracer tape. The diffraction curve from the pipe can still be seen in the migrated

Figure 5.9 Time migrated sections from line 4. **A** 100 MHz section. **B** 200 MHz section. The capillary fringe reflection is indicated. The diffraction from a pipe is circled near the centre of the sections.

Figure 5.10 Depth and elevation corrected sections from line 4. **A** 100 MHz section. **B** 200 MHz section. **C** Apparent conductivity data. The ground surface, capillary fringe reflection, location of well MW 5, and the water table elevations are indicated on the GPR sections. The diffraction from a pipe is circled near the centre of the GPR sections, and is responsible for the large anomaly in the apparent conductivity. At MW 5, the subsurface is composed of sand and gravel.

sections. The emplacement of the pipe has likely changed the properties of the subsurface in the region of the pipe, creating a difference between the velocity of the radar wave scattering off the pipe compared to the surrounding material.

Line 4 provided the best image of the capillary fringe reflection from the Gilby Waterflood site. On some of the other lines, cross-cutting reflectors or interference from overlying reflectors made it difficult to pick the onset of the capillary fringe reflector. On line 3 (fig. 5.11) cross cutting reflectors caused particular problems in picking the initial rise of the capillary fringe reflector. Figure 5.11 shows the final depth sections for the 100 and 200 MHz data from line 3. Cross cutting reflectors interfere with the capillary fringe reflector from 8 to 20 m. While the general position of the capillary fringe reflection is still quite obvious, it is much more difficult to pick the specific time for the onset of the reflection. Between 27 and 31 m, there appears to be a sudden rise in the capillary fringe reflector on the 100 MHz section (fig. 5.11A). When we examine the same area on the 200 MHz section, the finer detail indicates that the rise on the 100 MHz section is likely due to interference between the capillary fringe reflection and an overlying reflection.

The GPR data from line 2 illustrates another difficulty encountered in interpreting the capillary fringe reflection. Figure 5.12 shows the final depth sections for the 100 and 200 MHz antennas along line 2. In both frequencies, the capillary fringe reflection is well defined for most of the line. At the south (0 m) end of line 2 the capillary fringe is much closer to the surface, and it is more difficult to pick the onset of the reflection. The apparent electrical conductivity also increases at this end of the line, as seen in fig. 5.12C. The north end of line 2 also suffers from a decreased amplitude of the capillary fringe reflection, as well as some possible interference from overlying reflectors. Otherwise, the capillary fringe reflection is quite continuous. However, this reflection does not have the characteristic flatness seen in lines 3 and 4 (fig. 5.11 and 5.10). Between 20 and 40 m, the capillary fringe reflection is significantly higher than it is between 0 to 20 m and 40 to 60 m along line 2. We considered the possibility that this rise might be caused by changes in the velocity profile along line 2.

Figure 5.11 Depth and elevation corrected sections from line 3. A 100 MHz section. B 200 MHz section. C Apparent conductivity data. The ground surface and capillary fringe reflection are indicated on the GPR sections.

Figure 5.12 Depth and elevation corrected sections from line 2. **A** 100 MHz section. **B** 200 MHz section. **C** Apparent conductivity data. The ground surface, capillary fringe reflection, location of well MW 1, and the water table elevations are indicated on the GPR sections. At MW 1, the subsurface is composed of fine sand with trace fine gravel.

Experimentation with various reasonable velocities did not result in significant changes in the capillary fringe depth profiles. As well, we can see that the same feature occurs on both the 100 and 200 MHz profiles. As the velocities for each frequency were chosen independently, the capillary fringe reflector depth changes are unlikely to be caused by errors in the velocity profiles. If we assume that the capillary fringe reflector depth variations are not due to velocity effects, then it is likely that we are seeing changes in the capillary rise due to changing grainsize distributions at the water table. Taking the water table elevation in the well MW 1 at the time of data collection as a datum, we subtracted it from the capillary fringe elevations. For the area between 20 and 45 m on line 2, this gave the experimental rise of the capillary fringe above the water table. The maximum rise in the capillary fringe was 0.53 m and 0.42 m for the 100 and 200 MHz profiles respectively. Using eq. (2.54) to determine the theoretical pore radius which would result in this capillary rise gave values of 0.0028 cm and 0.0036 cm respectively. Using the approximation that the pore radius is 0.2 times the grain diameter resulted in $D_{100\text{ MHz}} = 0.014\text{ cm}$ and $D_{200\text{ MHz}} = 0.018\text{ cm}$ which is the classification for fine sand. The drillers logs from well MW 1 at 22 m along line 2 (see fig. 5.12) indicates that the water table is within a section of fine sand with trace fine gravel.

On the above lines, the reflection from the top of the capillary fringe is identifiable, if occasionally obscured. Figures 5.13 and 5.14 show line 1 and cross line 20 using the 100 and 200 MHz antennae. On both these sections, the capillary fringe reflection is discontinuous. The spherical and exponential gains applied to the 200 MHz section (fig. 5.13B) have increased the amplitude of the high frequency, coherent noise. This noise is always present, but is generally of much smaller amplitude than returning reflections. It is only identifiable on sections which have low amplitude returns from the subsurface. The capillary fringe reflection on line 1 can be identified between 8 and 30 m as a relatively continuous reflection beginning with a positive half-cycle. From 30 to 68 m on the 100 MHz section (fig. 5.13A), the capillary fringe reflection disappears except for a brief indication of its location between 46 and 54 m. On the 200 MHz section (fig. 5.13B), the capillary fringe reflection is indistinguishable from noise between 30 and 75 m. The reason for the problems with this

Figure 5.13 Depth and elevation corrected sections from line 1. **A** 100 MHz section. **B** 200 MHz section. **C** Apparent conductivity data. The ground surface, capillary fringe reflection, location of well MW 6, and the water table elevations are indicated on the GPR sections. The drillers logs are provided at the bottom of the figure. At the water table, MW 6 indicates the subsurface is composed of medium sands.

area may be suggested in the top 1.5 to 2 m of data. On both line 1 profiles (fig. 5.13), there are two regions which are devoid of reflection below the direct wave, from 30 to 45 m and from 55 to 70 m along the lines. The CMP gathers collected at 40 and 60 m did not have identifiable reflections below the direct wave, nor was the direct wave identifiable at longer offsets. Using the 100 MHz antennae, the direct wave velocity was determined to be 0.073 m/ns at 40 m and 0.113 m/ns at 50 m. For the 200 MHz data, the only measurable direct wave velocity was 0.108 m/ns on the CMP gather collected at 50 m. These values are lower than the mean for the site of 0.125 m/ns, and consist of some of the lowest direct wave velocities measured. The direct wave velocity at 40 m is consistent with a finer grained material such as silt or clay. It is possible that these areas were filled with finer grained material following an excavation of some sort.

The decrease in velocity and loss of signal return seen on line 1 is also exhibited on many of the cross lines. Figure 5.14 shows the 100 and 200 MHz depth sections for the cross line at 20 m. From 0 to 15 m, the capillary fringe reflection is relatively well defined and continuous. At approximately 17 m, all returning signal from the subsurface is lost. In the upper part of both sections, there is a major change in the direct wave return at 16 m. As on line 1, this is likely due to a decrease in the near surface velocity. A common midpoint gather was collected in each frequency just before the change in character on cross line 20, at 15 m. The direct wave velocities measured at this location are $v_{100 \text{ MHz}} = 0.109 \text{ m/ns}$ and $v_{200 \text{ MHz}} = 0.112 \text{ m/ns}$ which are some of the lowest direct wave velocities measured at the site. We also noted increased vegetation to the south east of line 1, starting on cross line 20 at approximately 16 m. All of these observations support the hypothesis that there is a decrease in grain size in the near surface after 16 m along cross line 20, and in general to the south east of line 1.

As seen in fig. 5.1, there are a number of locations where various lines interconnect. Along with this, lines 1, 2 and 4 pass by monitoring wells MW 6, MW 1 and MW 5 respectively. Table 5.5 shows the tie points between the main GPR lines. Included in the table are

Figure 5.14 Depth and elevation corrected sections from cross line 20. **A** 100 MHz section. **B** 200 MHz section. The ground surface and capillary fringe reflection are indicated. The capillary fringe reflection is discontinuous and disappears after approximately 16 m.

collection dates and times, as there were changes in the water table elevations throughout data collection (see fig. 5.2). Table 5.6 shows the tie points between line 1 and the cross lines at 10, 20, 30, and 70 m. In general, the 200 MHz data ties together more closely than the 100 MHz data. When the acquisition dates are compared to the change in water table elevation,

Table 5.5 Comparison of tie points on the ground penetrating radar lines.

	100 MHz		200 MHz	
	Date	CF elev. (m asl)	Date	CF elev. (m asl)
line 1 @ 85 m	06/10/97 12:30	860.33	06/17/97 15:00	860.00
xline 85 @ 9.5 m	06/13/97 14:00	859.81	06/17/97 16:00	860.14
difference between line 1 and xline 85		0.52		0.14
xline 85 @ 31.5 m	06/13/97 14:00	859.60	06/17/97 16:00	860.12
line 2 @ 46 m	06/14/97 09:00	859.96	06/17/97 11:00	860.20
difference between xline 85 and line 2		0.36		0.08
line 2 @ 72.5 m	06/14/97 09:00	859.86	06/17/97 11:00	860.36
line 3 @ 2.5 m	06/14/97 14:00	860.19	06/17/97 09:00	860.40
difference between line 2 and line 3		0.33		0.04
line 3 @ 48.5 m	06/14/97 14:00	860.21	06/17/97 09:00	860.08
line 4 @ 13 m	06/15/97 10:00	860.39	06/16/97 16:00	860.24
difference between line 3 and line 4		0.18		0.16

this is not unexpected. The 100 MHz data was acquired over a period of six days from June 10 to June 15, 1997, while the 200 MHz data was acquired from June 16 to June 18, 1997. Figure 5.2 shows the water table elevations measured between June 10 and June 18, 1997. The 100 MHz data from lines 2, 3, and 4 was acquired during the large change in water levels on June 14 and 15. It is therefore unsurprising that these values tie poorly. In contrast, the 200 MHz data was collected over a relatively short period of time, during a more gradual rise in water table elevations. The tie point comparisons in Tables 5.5 and 5.6 reflect this in greatly reduced differences between the tie point elevation estimates. We will see that the differences in capillary fringe elevation estimates between the tie points are on the same order as the uncertainties in the water table elevation estimates themselves. The tie points between the cross lines and line 1 (Table 5.6) show similar values to the other line comparisons. The 100 MHz cross lines tie relatively poorly to line 1, even though the majority of the cross lines were collected during the period of stable water table elevations.

Table 5.6 Comparison of tie points between line 1 and the cross lines collected at 10, 20, 30 and 70 m along the line.

	100 MHz		200 MHz	
	Date	CF elev. (m asl)	Date	CF elev. (m asl)
Line 1 @ 10 m Xline 10 @ 4 m difference between line 1 and xline 10	06/10/97 12:30 06/11/97 14:20	859.94 860.27 0.33	06/17/97 15:00 06/17/97 13:40	860.32 860.30 0.02
Line 1 @ 20 m Xline 20 @ 15 m difference between line 1 and xline 20	06/10/97 12:30 06/11/97 14:00	860.20 860.33 0.13	06/17/97 15:00 06/18/97 13:55	860.42 860.52 0.10
Line 1 @ 30 m Xline 30 @ 20 m difference between line 1 and xline 30	06/10/97 12:30 06/11/97 15:00	859.58 859.79 0.21	06/17/97 15:00 06/18/97 15:00	860.33 860.00 0.33
Line 1 @ 70 m Xline 70 @ 2 m difference between line 1 and xline 70	06/10/97 12:30 06/13/97 13:00	859.85 859.70 0.15	06/17/97 15:00 06/17/97 18:20	859.93 859.93 0.00

Cross line 20 in fig. 5.14 illustrates the reasons for the poor tie between the cross lines and line 1. The tie between cross line 20 and line 1 is very close to the poor data area. As well, cross cutting reflections hinder the accurate identification of the onset of the capillary fringe reflection near the tie point with line 1. This interference is greatly reduced on most of the 200 MHz sections, correspondingly reducing the differences in capillary fringe elevation between the cross lines and line 1. The differences between the cross lines and line 1 at the tie points can be expected due to variable data quality and reflection interference on the cross lines and on line 1.

In Chapter 3, we determined the uncertainty in the water table elevations measured from monitoring wells and the uncertainties in the capillary fringe and water table elevations estimated from the GPR data. The equations developed to calculate capillary fringe and

water table elevations, as well as their uncertainties, are

$$WT_{elev}(well) = Z - WTD \quad (5.5)$$

$$\Delta WT_{elev}(well) = \sqrt{\Delta Z^2 + \Delta WTD^2} \quad (5.6)$$

$$CF_{elev}(GPR) = Z - d_{cap} \quad (5.7)$$

$$\Delta CF_{elev}(GPR) = \sqrt{\Delta Z^2 + \Delta d_{cap}^2} \quad (5.8)$$

$$WT_{elev}(GPR) = CF_{elev}(GPR) - h_{cap} = Z - d_{cap} - h_{cap} \quad (5.9)$$

$$\Delta WT_{elev}(GPR) = \sqrt{\Delta CF_{elev}(GPR)^2 + \Delta h_{cap}^2} = \sqrt{\Delta Z^2 + \Delta d_{cap}^2 + \Delta h_{cap}^2} \quad (5.10)$$

In these equations, WT_{elev} is the estimated water table elevation from a well or from the GPR, CF_{elev} is the capillary fringe elevation, Z is the surface elevation, WTD is the water table depth below the surface, d_{cap} is the depth of the capillary fringe below the surface, and h_{cap} is the height of the capillary rise above the water table. The uncertainty in the depth to the capillary fringe estimated from GPR sections, Δd_{cap} , is dependent on the uncertainty in the depth pick, Δd_{pick} , and the uncertainty in depth due to velocity uncertainties, Δd_{vel} . The uncertainty in d_{cap} is calculated using

$$\Delta d_{cap} = \sqrt{\Delta d_{pick}^2 + \Delta d_{vel}^2} \quad (5.11)$$

Table 5.7 summarizes the uncertainties in Δd_{pick} , Δd_{vel} , Δd_{cap} , ΔZ , and Δh_{cap} for each of the lines which tie with a monitoring well. The capillary rise, h_{cap} , was chosen to be the mean value of the saturated pressure head, ψ_s for sands from Cosby et al. (1984), or $h_{cap} = 0.069$

Table 5.7 Uncertainties which contribute to the uncertainty in the capillary fringe and water table elevation estimates from ground penetrating radar sections.

		Δd_{pick} (m)	Δd_{vel} (m)	Δd_{cap} (m)	Δh_{cap} (m)	ΔZ (m)
wells						0.005
100 MHz	line 4	0.09	0.10	0.13	0.12	0.005
	line 2	0.09	0.08	0.12	0.12	0.005
	line 1	0.09	0.15	0.17	0.12	0.005
200 MHz	line 4	0.04	0.09	0.10	0.12	0.005
	line 2	0.04	0.06	0.07	0.12	0.005
	line 1	0.04	0.11	0.12	0.12	0.005

± 0.12 m. The uncertainties in Table 5.7 will be used in eq. (5.8), and (5.10) to determine $\Delta CF_{elev}(GPR)$, and $\Delta WT_{elev}(GPR)$. The uncertainty in the water table elevation measurements from the wells is dependent on the uncertainty in the surface elevation at the well, $\Delta Z = \pm 0.005$ m, and the uncertainty in the water table depth measurement, $\Delta WTD = \pm 0.005$ m. Combining these equations using eq. (5.6) gives an uncertainty in the measured water table elevation in each well of $\Delta WT_{elev}(well) = \pm 0.007$ m.

Tables 5.8 and 5.9 provide a comparison of the measured water table elevations from the wells and the estimated capillary fringe and water table elevations from the ground penetrating radar data for the 100 MHz and 200 MHz data respectively. In all cases, the wells are located between 3 and 8 m to the side of the GPR comparison points. When the uncertainty in the water table elevation estimates from the GPR data ($\Delta WT_{elev}(GPR)$) is compared to the difference between the measured water table elevations and the GPR estimates ($\epsilon_{WT_{elev}}$) we see that in general, $\epsilon_{WT_{elev}}$ is within the uncertainty in the GPR estimate of the water table elevation. Therefore, the water table elevation measured in the well falls within the uncertainty of the water table elevation estimated from the ground penetrating radar data. There are two exceptions; between line 4 at 100 MHz and MW 5 (Table 5.8) and between line 2 at 200 MHz and MW 1 (Table 5.9). In both cases, the water table elevation

Table 5.8 Comparison of water table elevation estimates from 100 MHz GPR sections with the corresponding water table elevation measurements from the wells.

Line and Wells to compare	Date	GPR estimates		Well	$\epsilon_{WT_{elev}}$ (m)	sediment @WT
		CF _{elev}	WT _{elev}	WT _{elev}		
line 4 @ MW 5	06/15/97 10:00	860.38 ± 0.13	860.31 ± 0.18	860.085 ± 0.007	0.23	sand and gravel
line 2 @ MW 1	06/14/97 09:00	860.11 ± 0.12	860.04 ± 0.17	859.930 ± 0.007	0.11	fine sand with trace fine gravel
Line 1 - 7.5 m @ MW 6	06/10/97 12:30	859.97 ± 0.18	859.90 ± 0.21	859.960 ± 0.007	-0.06	medium sand

measured in the well falls just outside the uncertainty in the water table elevation estimate from the GPR data. In both frequencies, the water table elevation estimates from the GPR data collected along lines 2 and 4 are higher than the monitoring well measurements. This may indicate that a larger value should be used to convert from capillary fringe elevation to water table elevation. The value used ($h_{cap} = 0.069$ m) is closer to the theoretical value for medium to coarse sands, while the drillers logs identify fine sands at the water table in MW 1, and sand and gravel throughout MW 5. In comparison, the GPR data acquired along line 1 ties very well with monitoring well MW 6 using an $h_{cap} = 0.069$ m. In all cases, the water table elevations measured at MW 6 agree with the GPR estimates to within the uncertainty in the GPR water table elevation estimates.

Contoured maps of the water table elevation estimates from the ground penetrating radar data are shown in fig. 5.15 and 5.16 for the 100 and 200 MHz data respectively. For a number of reasons, the 200 MHz water table map is considered to be a more accurate representation of the water table elevation estimates than the 100 MHz map. When the tie points between the various lines were compared, the differences between the 200 MHz data fell within the uncertainty in the GPR estimates while the differences between the 100 MHz data did not. As well, the change in measured water table elevations during acquisition of the 200 MHz data was smaller than the change in water table elevations during acquisition of the 100 MHz data (fig. 5.2). Finally, the 200 MHz data had fewer incidents of interference between

Table 5.9 Comparison of water table elevation estimates from 200 MHz GPR sections with the corresponding water table elevation measurements from the wells.

Line and Wells to compare	Date	GPR estimates		Well	$\epsilon_{WT_{elev}}$ (m)	sediment @WT
		CF _{elev}	WT _{elev}	WT _{elev}		
line 4 @ MW 5	06/16/97 16:00	860.32 ± 0.10	860.25 ± 0.16	860.150 ± 0.007	0.10	sand and gravel
line 2 @ MW 1	06/17/97 11:00	860.43 ± 0.07	860.36 ± 0.14	860.140 ± 0.007	0.22	fine sand with trace fine gravel
Line 1 (17th) @ MW 6	06/17/97 15:00	860.30 ± 0.12	860.23 ± 0.17	860.150 ± 0.007	0.08	medium sand
Line 1 (18th) @ MW 6	06/18/97 09:30	860.26 ± 0.12	860.19 ± 0.17	860.165 ± 0.007	0.02	medium sand

overlying reflections and the capillary fringe reflection. On both maps, the 860.2 to 860.4 m a.s.l. contour at approximately 480 E, 170 N is interpreted as the rise in the capillary fringe reflection on line 2 which was discussed previously.

5.4.3 Apparent Electrical Conductivity Measurements

Apparent electrical conductivity measurements were collected at the Gilby Waterflood site in June, 1999. Water table elevations were not available as the monitoring wells had been removed. The Red Deer River water level was similar to the levels on June 9 and 10, 1997. The electrical conductivity measurements are shown in part C of fig. 5.10, 5.11, 5.12 and 5.13. Excluding the values associated with the pipeline, the conductivities ranged from 7.61 to 13.07 mS/m with a mean of 9.63 mS/m for the vertical dipoles and from 5.51 to 12.34 mS/m with a mean of 7.56 mS/m for the horizontal dipoles. In the figures, only the vertical dipole conductivity measurements are shown, as the horizontal dipole measurements mimicked the vertical dipole measurements at a slightly lower conductivity. The highest conductivities occurred along line 1 from 50 to 78 m, and along line 2 from 10 to 20 m. These are both areas where data quality was relatively poor. As well, these areas were characterized by a decrease in grainsize at the surface, an increase in compaction and water retention following precipitation, as well as an increase in vegetation. Due to excavations,

Figure 5.15 Surface contour map of the water table elevation estimates from the 100 MHz GPR data. Plus (+) signs indicate the location of the GPR water table elevation estimates used to contour the surface.

conductivity values from 0 to 50 m on line 1 were not collected. On line 4, a polyethylene pipe and overlying magnetic tracer tape are located at approximately 27 m. On fig. 5.10C, the pipe shows up as the large anomaly from 15 to 35 m. On the GPR profiles, the apex of the diffraction pattern from the pipe is located between 27 and 28 m along the line and

Figure 5.16 Surface contour map of the water table elevation estimates from the 200 MHz GPR data. Plus (+) signs indicate the location of the GPR water table elevation data used to contour the surface.

861.75 m a.s.l., or 2.25 m below the ground surface. Due to a loss of energy, the amplitude of the capillary fringe reflection below this area is greatly reduced.

5.5 Summary

There are a number of conclusions which can be drawn from the data collected at the Gilby Waterflood field site. A CMP gather density of one CMP gather every 10 m along the common offset lines appears to have adequately sampled the velocity variations at the site. Determining the capillary fringe elevations from the common offset GPR sections had a variety of problems, but was generally a success. There were a few areas of decreased data quality which were correlated to higher apparent conductivities and increased vegetation. All three features are likely caused by an increase in clay and silt in the near surface. The 200 MHz GPR lines tie together relatively well compared to the uncertainty in the water table elevations estimated from GPR, while the 100 MHz data does not tie as well. The capillary fringe elevation estimates were converted to water table elevation estimates using a single value for the capillary rise across the site. The water table elevation estimates were then compared with measured well values to determine the accuracy of the method.

Most of the CMP gathers collected at the Gilby Waterflood site have the same general trend in their RMS velocity profiles. The RMS velocity is relatively high at the surface, decreasing in the unsaturated zone, then increasing at the capillary fringe reflection. Where reflections exist below the capillary fringe, the RMS velocity is observed to decrease again. The mean RMS velocity to the capillary fringe was determined to be 0.129 ± 0.004 m/ns with the maximum variation being 9.5 %. Given that the area is composed of a wide range of sands and gravels, as well as clay lenses, a 9.5 % variation in the RMS velocity to the capillary fringe is lower than expected. We also examined the spatial variation of the velocities at the Gilby site by examining the semivariogram of the harmonic mean of the interval velocities above the capillary fringe. The model semivariogram showed that the velocities from the CMP gather are correlated to a maximum distance of 20 m, giving a maximum spacing between CMP gathers in order to capture the spatial variation in velocity. Therefore, a CMP gather density of one CMP gather for every 10 m of common offset line coverage appears to have captured the major velocity variations along the common offset sections.

Various difficulties were experienced in determining the location of the capillary fringe reflection on the final GPR depth sections. These included interference between the capillary fringe reflection and overlying reflectors, significant changes in the capillary fringe depth with no apparent interference from other reflectors, and loss of the capillary fringe reflection due to attenuation of the radar wave in the unsaturated zone. Interference between the capillary fringe reflection and overlying reflectors caused difficulties in identifying the first half-cycle of the capillary fringe reflection. When this occurred, the second and third half-cycles were still identifiable but the onset of the capillary fringe reflection was distorted by the overlying reflector. Areas of interference were usually easy to identify, as the overlying reflector typically cross-cut the capillary fringe reflection with a greater dip angle. As well, since the 200 MHz data had less extensive interference between dipping reflectors and the capillary fringe, the 200 MHz data could be used to clarify the interpretation of the 100 MHz data. The capillary fringe was also observed to rise and fall significantly with no apparent interference from other reflectors. If errors in the velocity can be discounted, these fluctuations in the capillary fringe elevation are likely caused by physical changes in the grainsize distribution which would result in changes in the height of the capillary fringe above the water table. Grainsize distribution changes would not be unexpected given the range of grain sizes found in the drillers logs. Finally, the capillary fringe reflection occasionally disappeared entirely for the GPR data. Loss of the capillary fringe reflection often occurred in conjunction with a decrease in data quality throughout the section. Poor data quality has been correlated to higher apparent conductivity values and an increase in vegetation, both of which are likely caused by an increase in the clay content in the near surface. Higher clay content soils tend to attenuate the GPR signal, causing poor data quality lower in the section.

The capillary fringe elevation estimates were converted to water table elevation estimates using $h_{\text{cap}} = 0.069$ m, which was the mean value for the saturated pressure head for sands from Cosby et al. (1984). When the water table elevations measured in the monitoring wells were compared with the GPR estimates, the values matched quite well. Line 1 at MW 6

matched quite closely, with the difference in water table elevations within the uncertainty in the GPR estimates. Line 2 at MW 1 and line 4 at MW 5 were not as close, with the difference in water table elevations falling just within the uncertainty in the GPR estimates for the 100 MHz line 2 and 200 MHz line 4 data, and just outside the uncertainty in the GPR estimates for the 200 MHz line 2 and 100 MHz line 4 data (Table 5.8 and 5.9). All of the comparisons between line 2 and MW 1 and line 4 and MW 5 have water table elevation estimates which are higher than the monitoring well measurements. This may indicate that a larger value should be used to convert from capillary fringe elevation to water table elevation. The value used ($h_{cap} = 0.069$ m) is closer to the theoretical value for medium to coarse sands, while the drillers logs identify fine sands at the water table in MW 1, and sand and gravel throughout MW 5. This shows that in at a heterogeneous site such as this one, using a single capillary rise value to convert capillary fringe elevations into water table elevations may not be the best solution. Each of the three wells provided information on the grainsize distribution at the water table, which could have been used to determine the capillary rise at that location. However, in order to convert all of the capillary fringe elevation estimates into water table elevation estimates using grainsize information from the wells, we would have to come up with some method for estimating how the grainsize changed between the wells. The fine sand found in MW 1 along line 2 appeared to be limited to approximately 20 m along the line, indicating that the drillers logs from the monitoring wells provide very localized information about the grainsize at this site. Extrapolating the information from the wells to the rest of the data might cause additional errors in the water table elevation estimates.

The water table surface maps determined using a constant value of $h_{cap} = 0.069$ m provide a relatively good estimate of the water table elevations across the site. From well comparisons and the GPR ties, we estimate that the 100 MHz surface contour map is within 0.30 m of the actual water table elevation, and the 200 MHz map is within 0.15 m. The general trend of the surface contour maps are accurate, but the minor fluctuations are more likely due to changes in the height of the capillary fringe than changes in water table elevation. These changes may be useful in pinpointing grainsize variations, but cannot be

taken as indicative of the actual water table elevations.

6 Strathroy, Ontario

6.1 Study Site

The Strathroy test field is located in Southern Ontario near the town of Strathroy. It has been used for hydrogeology studies involving the fate and transport of nitrates in the subsurface. A fence of piezometers is located in the northeastern corner of the field. Monitoring wells are also located throughout the field. The subsurface is composed of a relatively homogeneous fine sand which is part of the Caradoc Sand Plains (Chapman and Putnam, 1984). The sand plain thins to the west until eventually the underlying clay appears at the surface. At the location of the study site, the underlying clay is believed to be deeper than the limit of penetration with 200 MHz central frequency GPR antennae. Figure 6.1 provides a map of the study site, including monitoring well locations, ground penetrating radar lines and surface topography.

The water table is found in the sands approximately 4.4 m below the ground surface. The ground water flow in the vicinity of the GPR lines is to the SE with a gradient of 0.003. The GPR common offset lines were collected parallel to the ground water flow direction.

6.2 Dataset

The Strathroy dataset contains ground penetrating radar common offset lines and common midpoint gathers, water table elevations, survey points and an electromagnetic induction survey. The ground penetrating radar data was collected between November 28 and December 18, 1997 by the Environmental Geophysics Facility at the University of Waterloo, Waterloo, Ontario. Following preliminary testing, they selected the 200 MHz antennae for the GPR survey. Five 70 m and three 12 m common offset profiles as well as three common midpoint gathers were collected. The 70 m profiles are spaced two metres apart with the central line (line 6) perpendicularly bisecting the piezometer fence. The lines begin 1 m to

Figure 6.1 Site map for the Strathroy field site. The surface topography, locations of the GPR common-offset profiles, CMP gathers (*), and monitoring wells (●) are indicated.

the SE of the fence and extend NW in the upgradient direction of ground water flow. The 12 m profiles cross the main lines at 4, 29, and 59 m. The common midpoint gathers were collected on line 10 at 7.5 and 45 metres and on line 6 at 50 metres. The locations of the CMP gathers are indicated on Figure 6.1. Data collection parameters for the profiles and the CMP gathers are given in Table 6.1.

Table 6.1 GPR data collection parameters for the Strathroy site.

frequency	200 MHz
common offset antenna separation	0.75 m
common offset station spacing	0.10 m
CMP step size	0.10 and 0.20 m
CMP maximum offset	16.00 and 20.00 m
sampling interval	0.8 ns
number of stacks	64
pulser voltage	400 V

Water table elevations are available from October to December 1997 from a number of wells throughout the field. Four of these wells are close to the GPR lines and can be used to provide water table elevations for comparison. A fifth well is located 44 m to the NW of the 70 m end of the lines, providing an indication of the trend in water table elevations. Figure 6.2 provides the water table elevations from these five wells from October 20 to December 5, 1997. Water table elevation measurements were not collected on the same dates as the GPR data. Short term fluctuations of approximately 10 cm could be expected in the water levels.

The GPR lines were not surveyed at the Strathroy site. Easting and Northing values for each end of the lines as well as Easting, Northing and elevation values from random points throughout the field were available. Surface elevations for the GPR lines were interpolated from the surveyed points using a modified version of a GSLIB two dimensional ordinary kriging program. The square root of the kriging variance is equal to the uncertainty in the elevation estimates. This gives an uncertainty in the surface elevations along the GPR lines of $\Delta Z = \pm 0.037$ m. This is much higher than the uncertainty in the surface elevations for the Gilby Waterflood and North Caicos Island sites, which both have estimated uncertainty in

Figure 6.2 Water table elevations collected from five monitoring wells at the Strathroy field site.

the surface elevations of ± 0.005 m.

An EM-31 electromagnetic induction survey was performed on an adjacent field. The vertical dipole conductivities ranged from 5.40 to 7.74 mS/m with a mean of 6.92 mS/m. This is greatly reduced from the range seen at the Gilby Waterflood site, and gives further evidence of the homogeneity of the subsurface below the Strathroy fields.

6.3 Processing Characteristics

The Strathroy GPR data was processed using the processing flows outlined in Chapter 4. Table 6.2 provides the actual processing parameters applied to the data. The first break picks on the common offset profiles were flattened to 2.5 ns, due to the 0.75 m antenna separation.

Table 6.2 Processing Parameters for the Strathroy GPR data.

Airwave Flattening *	$t = 2.5 \text{ ns}$
Q Compensation	$Q = 50$
Ormsby Bandpass filter	20-40-270-290 MHz
Steep Dip Explicit Finite Difference Time Migration *	maximum frequency = 270 MHz maximum dip = 70°
Elevation correction *	Datum = 169.00 m asl

* processing parameters which were applied only to the common-offset data

The Q compensation value of 50 was determined by experimentation, as were the bandpass filter ramps. Constant Q compensation was not as successful at whitening the amplitude spectrum in the Strathroy data as it was in the Gilby data. Application of the constant Q compensation has also had the effect of increasing the amplitude of the high frequency noise below the capillary fringe reflection. An improvement can be seen in the unsaturated zone, as well as in the capillary fringe reflection itself however. The maximum frequency to migrate in the Kirchhoff time migration was based on the high pass limit of the bandpass filter. As the highest elevation along the GPR lines was 168.70 m a. s. l., a datum of 169.00 m a. s. l. was chosen for the elevation corrections.

6.4 Results

6.4.1 Common Midpoint gathers

Three common midpoint gathers were collected at the Strathroy site. This provides considerably less velocity information than we have for the Gilby Waterflood or North Caicos Island sites. The direct wave and capillary fringe RMS velocities are provided in Table 6.3. The direct wave velocities are all within 0.005 m/ns of each other, giving a relatively homogeneous near surface. The RMS velocities at the capillary fringe are even closer, with a difference between the maximum and minimum of only 0.003 m/ns. In fig. 6.3, the RMS velocity profiles from the CMP gathers are shown. For a variety of reasons, we

Table 6.3 Direct wave velocity and RMS velocity at the capillary fringe for CMP gathers at Strathroy, Ontario.

CMP location	v_{direct} (m/ns)	v_{RMS} capillary fringe (m/ns)
line 10 @ 7.5 m	0.105	0.107
line 10 @ 45 m	0.110	0.105
line 6 @ 50 m	0.109	0.106
average velocity function	0.108	0.104
mean of 3 CMP gathers	0.108	0.106

determined that the three CMP gathers were not sufficient to describe the velocity variations along the profiles. A mean RMS velocity function was therefore calculated and used for the entire dataset. This function is shown in bold in fig. 6.3. In Table 6.3, the RMS velocity at the capillary fringe is 0.104 m/ns, which is less than the mean value of v_{RMS} at the capillary fringe for the 3 CMP gathers. This is due to the difference in two way travel time to the capillary fringe for each of the CMP gathers. Between 50 and 75 ns, the average velocity function is determined from values above and below the capillary fringe reflection, resulting in a slightly lower average velocity at the capillary fringe for the average velocity function than the mean of the three RMS velocities at the capillary fringe reflection. On the whole the average velocity function defines the known velocities relatively well.

6.4.2 Common Offset GPR lines

The location of the common offset GPR profiles collected on the Strathroy field are shown in fig. 6.1. The lines were positioned to attempt to image the capillary fringe surface on the upgradient side of a piezometer fence. Three wells are located near the SE end of the lines, with an additional well approximately 44 m to the NW of the lines. Figure 6.4 shows the time migrated and depth sections for line 10. All five of the 70 m long lines show the same general trends as line 10. From 0 to 35 m, the capillary fringe reflection is identifiable as the one and a half cycle reflection beginning with a positive half cycle at approximately 165 m

Figure 6.3 RMS velocity profiles for the CMP gathers collected at Strathroy, and the average function used to convert the two-way time GPR sections to depth.

above sea level. Between 35 and 50 m, the capillary fringe reflection weakens until it is unidentifiable. At the beginning of line 10, there is a dip in the capillary fringe reflection. Above the capillary fringe, many of the reflections exhibit the same dip. You will recall that the GPR lines were not surveyed at this site. Instead, surveyed points from throughout the field were used to interpolate the surface elevations along the lines. While kriging has provided a good estimate of the general surface elevation changes, minor topographic changes cannot be estimated. The dip in the reflections found between 5 and 10 m on line 10

Figure 6.4 Strathroy line 10. **A** Time migrated section. **B** Depth and elevation corrected section. The ground surface, and capillary fringe reflection are indicated on the GPR sections. The deep reflection which has similar characteristics to the capillary fringe reflection is circled on each section.

is most likely caused by a minor surface elevation change which is not present in the interpolated surface elevations. This dip is not present on line 8, which shows the limited extent of this elevation change. Similar minor features like this can be found throughout the data.

Final depth sections for the cross lines are shown in fig. 6.5. The cross lines at 4 and 29 m along the lines have a well defined capillary fringe reflection which matches well with the main lines, as shown in Table 6.4. The difference between the capillary fringe reflection elevations at the tie points range between 0.01 and 0.17 m. The largest differences in elevation at the tie points are in areas where the onset of the capillary fringe reflection is not clearly defined.

In contrast to the cross lines at 4 and 29 m, the cross line at 59 m does not have a clearly identifiable capillary fringe reflection. As mentioned previously, the capillary fringe reflection on the main lines also becomes unidentifiable after 50 m. Examining the later part of line 10 (fig. 6.4) and cross line 59 (fig. 6.5), we see a reflection at between 55 and 60 m along line 10 with an elevation of 164.2 m a.s.l. which has similar characteristics to the capillary fringe reflection in the early part of the sections. This is also seen on the cross line at 59 m as a strong reflection between 8 and 10 m and as a weaker reflection between 3 and 6 m along the line. However, this reflection is approximately 1 m below the capillary fringe reflection identified in the early part of the main lines. This reflection is also present on line 8, but is not identifiable or has a greatly decreased amplitude on the other main lines. Curiously, when the elevation of this reflection is compared to the wells near the southeast end of the line, there is a difference of only 0.10 m between the capillary fringe reflection and the water table elevation in the wells. It is possible that this reflection is a continuation of the capillary fringe reflection after a change in the subsurface, such as a change in grain size. Another possible cause for an apparent drop in the capillary fringe elevation is a change in velocity. As no CMP gathers were collected between 50 and 70 m along the main GPR lines, we have no way to determine if a velocity change occurs in the area. However, there is a

Figure 6.5 Depth and elevation corrected cross lines 4, 29 and 59 from Strathroy. The ground surface and capillary fringe reflection are indicated on cross lines 4 and 29.

change in the character of the GPR section, which may indicate a change in subsurface conditions. In Chapter 3, we saw that changes in the capillary fringe height combined with the use of incorrect velocities compounded the error in depth to the capillary fringe and could result in an error of 1 m or more. There is too little information to determine with any certainty the location of the capillary fringe reflection or the cause of the change in character

Table 6.4 Comparison of GPR tie points for cross lines 4 and 29.

	line CF_{elev} (m)	cross line 4 CF_{elev} (m)	difference (m)	line CF_{elev} (m)	cross line 29 CF_{elev} (m)	difference (m)
line 2	165.35	165.32	0.03	164.90	165.00	0.10
line 4	165.38	165.41	0.03	165.05	164.96	0.09
line 6	165.39	165.40	0.01	165.01	165.08	0.07
line 8	165.30	165.37	0.07	165.00	165.17	0.17
line 10	165.26	165.24	0.02	165.03	165.07	0.04

Table 6.5 Uncertainties which contribute to the uncertainty in the capillary fringe and water table elevation estimates from ground penetrating radar sections.

	Δd_{pick} (m)	Δd_{vel} (m)	Δd_{cap} (m)	Δh_{cap} (m)	ΔZ (m)
wells					0.005
line 2	0.04	0.08	0.09	0.15	0.04
line 4	0.04	0.07	0.08	0.15	0.04
line 6	0.04	0.07	0.08	0.15	0.04
line 8	0.04	0.08	0.09	0.15	0.04
line 10	0.04	0.07	0.08	0.15	0.04
xline 4	0.04	0.08	0.09	0.15	0.04
xline 29	0.04	0.06	0.07	0.15	0.04

on the NW end of the main GPR lines. A final possibility is that the circled event is a multiple of an overlying reflection at 165.7 m a.s.l.. Both events have similar structures, especially the dip in both events just before 60 m along the section. As well, the time difference between the surface and the first event is very similar to the difference between the first and second events.

In Chapter 3, we determined the uncertainty in the water table elevations measured in wells and estimated from ground penetrating radar. Table 6.5 summarizes the uncertainties in the depth pick, Δd_{pick} , the depth due to velocity uncertainties, Δd_{vel} , the depth to the capillary fringe from GPR sections, Δd_{cap} , the surface elevations, ΔZ , and the height of the capillary fringe above the water table, Δh_{cap} for each of the lines and the wells. These values will be used to determine the uncertainty in the capillary fringe and water table elevation estimates from GPR data and the water table elevation measurements from the wells. The capillary rise, h_{cap} used to convert capillary fringe elevations into water table elevations, must be different from the value used with the data from the Gilby Waterflood site. The Strathroy subsurface is composed of fine sand, which has a theoretical capillary fringe of between 30 and 60 cm. Therefore, the capillary rise used to convert $CF_{\text{elev}}(\text{GPR})$ into $WT_{\text{elev}}(\text{GPR})$ is h_{cap}

Table 6.6 Comparison of water table elevation measurements from the wells with the nearest water table elevation estimates from the GPR depth sections.

Lines and Wells to compare	GPR estimates		Well	difference (m)
	CF _{elev} (m)	WT _{elev} (m)	WT _{elev} (m)	
line 6 @ P7	165.49 ± 0.09	165.04 ± 0.17	164.29 ± 0.007	0.75
line 6 @ P8	165.49 ± 0.09	165.04 ± 0.17	164.26 ± 0.007	0.78
line 8 @ P9	165.29 ± 0.10	164.84 ± 0.18	164.25 ± 0.007	0.59
line 2 @ W6	165.41 ± 0.10	164.96 ± 0.18	164.20 ± 0.007	0.76

= 0.45 m ± 0.15 m. The uncertainties in CF_{elev}(GPR), WT_{elev}(GPR), and WT_{elev}(well) are calculated using eq. (3.7), (3.8) and (3.25).

Figure 6.1 identifies four monitoring wells located near the SE end of the main lines. Table 6.6 provides a comparison of the water table elevations measured in the wells to their closest capillary fringe and water table elevation estimates from the GPR sections. For all four monitoring wells, the difference between the GPR water table elevation estimates and the monitoring well water table elevation measurements are much greater than the uncertainty in the GPR water table elevation estimates. Figure 6.6 is a contour map of the water table elevations estimated from the GPR data. It shows a surface which is dipping to the northwest along the direction of the lines. As mentioned above, the ground water flow direction for the site is to the southeast, so the water table surface should also dip to the southeast. As the GPR and monitoring well water table elevations are not similar and the ground water flow direction from the GPR data is exactly opposite to the flow direction determined from the well data, we must draw the conclusion that the GPR water table elevations do not represent the actual water table elevations. The GPR data is in fact misleading and if used on its own would result in a flow direction which is reversed and of greater magnitude than the actual direction an magnitude determined from the monitoring wells. In order to understand these results, we can re-examine the sources of error in determining water table elevation estimates from GPR data. These include surface elevations, radar wave velocities and the estimated

Figure 6.6 Surface contour map of the water table elevation estimates from the GPR depth sections, using a capillary rise of $h_{\text{cap}} = 0.45$ m. Plus (+) signs indicate the location of the GPR water table elevation estimates used to contour the surface.

height of the capillary fringe above the water table.

As mentioned previously, the surface elevations along the GPR lines were not surveyed. Instead, surveyed points from throughout the field were used in a kriging algorithm to provide surface elevations along the GPR lines. While this method can not reproduce minor

topographic changes, the general topography is correct. Any minor errors in the surface elevations would not have a large effect on the water table elevation results.

As mentioned in section 6.4.1, the velocity variation at the site is also relatively unknown. The three CMP gathers which were collected do not have a great deal of variation between them, as can be seen in fig. 6.3 and Table 6.3. Additional CMP gathers between 7.5 and 45 m along the lines would have removed any concern about velocity variations through this section of the data. While the RMS velocities at the top of the capillary fringe are very similar for all three CMP gathers (fig. 7.3 and Table 7.2), there are differences between the RMS velocity profile for the CMP gather collected at the south end of the lines (line 10 at 7.5 m) and the two CMP gathers collected between 45 and 50 m along the lines. In the unsaturated zone, the RMS velocities are lower in the south than in the north. At and below the capillary fringe, the pattern reverses, with higher RMS velocities in the south than in the north. This may be an indication of velocity variations between the south and north ends of the line. In order to correct the dip of the estimated water table elevations, a 10 % change in the velocities would have to occur between 7.5 m and 45 m along the lines. This corresponds to a difference of approximately 0.01 m/ns between the south and north ends of the lines, with the south end higher than the north. Additional velocity data for the last portion of the lines (50 to 70 m) may also have helped the interpretation of the GPR data in this area. The change in reflection characteristics at the northwest end of the lines may indicate a change in velocity, as the radar velocity and reflection coefficient are both dependent on the dielectric constant of the subsurface. We do not have enough information to determine if velocity variations are a factor in the differences between the water table elevation estimates from ground penetrating radar and the actual water table elevations measured in the wells.

Finally, there is the effect of the capillary rise itself, which is dependent on grain size, a rising or falling water table, and the wetting and drying history of the site. Since we appear to have a relatively stable water table at the time of data collection, grain size is essentially the only important factor in the height of the capillary rise above the water table. If we assume that

the capillary fringe elevations estimated from the GPR depth sections are an accurate representation of the subsurface, the only way to reproduce the water table elevations and ground water flow directions determined from the monitoring wells is to assume that there is a change in the grainsize distribution along the line. At the southeast end of the lines, the grainsize would have to be in the range of very fine sand instead of fine sand, which would account for the additional 0.59 to 0.78 m error in the water table elevations at this end of the line. The grainsize would then have to increase to a fine to medium sand by the 50 m point along the lines. There is no indication from the grainsize analysis completed on the Strathroy site (Ryan, personal communication; Sayani and Stenson, unpublished data) that a change in grainsize is occurring.

6.5 Summary

While the results from the Strathroy field site have been disappointing, some interesting points can be taken from them. When surveying the GPR lines is not feasible and surface elevations are available for the area in general, kriging has proven to be a useful technique for determining the surface elevations along the GPR lines. However, artifacts in the GPR data show that kriging can not reproduce the minor topographic changes which occur in the surface topography.

Our understanding of the velocity profile and variation at the Strathroy field site is minimal. While the RMS velocity profiles from the three CMP gathers collected at the site are quite similar, the differences may be an indication of larger velocity variations along the lines. Therefore, we can not rule out velocity variations as the cause of the errors in the water table elevation estimates or the dip of the water table surface determined from the GPR depth sections. As well, a CMP gather at the northwest end of the GPR lines may have assisted the interpretation of that end of the common-offset lines.

Large errors of 0.59 to 0.78 m were found between the GPR and monitoring well water table

elevations. As well, the water table contour map developed from the GPR estimates has a dip which is exactly opposite to the ground water flow direction determined from the monitoring well data. We do not have coincident GPR data and water table elevation measurements, leading to the potential for the actual water table elevations on GPR data collection days to differ from those in fig. 7.2 by up to 10 cm. This does not account for the difference between the measured water table elevations and the GPR estimates, nor can it account for the error in dip. Therefore the water table elevations estimated from the GPR depth sections do not represent the actual water table conditions. While it can be shown that changes in grainsize along the GPR lines could account for the discrepancies between the monitoring well and GPR data, there is currently no evidence of grainsize changes from the analysis which has been done. With the information available at the present time, we cannot explain the errors in the GPR water table elevation results.

7 North Caicos Island, Turks and Caicos Islands, British West Indies

7.1 Study Site

The Turks and Caicos Islands are located on the Caicos Platform at the southern end of the Bahamian archipelago (Jackson, 1995). The northern half of the platform is bordered by a semi-continuous chain of islands with North Caicos Island near the centre of the string. The majority of the near surface on North Caicos Island is composed of Pleistocene aeolianites which appear as a hard, well cemented limestone often exhibiting a karstified surface (Jackson, 1995). The limestone forms a series of ridges which were originally deposited as wind blown sand dunes, and rise up to 16 m above the present day sea level. In the swales between the ridges, a series of elliptical depressions can sometimes be found. These depressions are one to two metres below the surrounding land, with sandy material within the depression and karstified limestone forming the outer edge. Figure 7.1 is a map of North Caicos Island, with the approximate locations of the five study areas indicated. Area B is on the south slope and top of a ridge at the south end of the island. Areas A and C are in the swales to the south and north respectively of the ridge at area B. Area C is in a deeper swale than area A and contains a number of the elliptical depressions. In area D, data was collected over one of the highest ridges on the island, as well as in the low to the north of the ridge. Area E to the northwest is within the transitional zone between the Caicos platform interior and platform margin (Jackson, 1995). The near surface is composed of Holocene beach ridges deposited against the Pleistocene limestone.

The inland ridges form the major source of fresh ground water on the island. Fresh water collected from precipitation events can pool on top of the salt water within the beach ridges. The extent of this ground water resource is presently unknown, but is generally considered to be minimal. Five wells were available during this study. The wells were hand dug, with a diameter of approximately 2 m. They penetrated the water table to no more than 60 cm, in order to avoid contaminating the fresh water with mixing zone water. Three of the five wells

Figure 7.1 North Caicos Island field sites. The five areas are indicated with the letters A through E. Areas A and C have been expanded to show the locations of the GPR and EM-31 lines, CMP gathers, wells, and depressions in these areas. Easting, Northing and elevation values were measured from a survey marker in area B which was assigned the arbitrary values of (N,E,Z) = (150, 150, 20).

were abandoned, one well was used for irrigation, and the fifth was used as a source of 'dirty water'. This last well was located in area E which is within the transitional zone between the Caicos platform interior and platform margin, and has a large variation in water quality depending on the location of the salt water wedge.

7.2 Dataset

The North Caicos Island dataset contains ground penetrating radar common offset lines and common midpoint gathers, water table depths, and electromagnetic induction data from five different locations on the island. Areas A, B and C were also surveyed. The data was collected between July 11 and 21, 1997. The GPR dataset contains common offset lines and CMP gathers collected with the 50, 100, and 200 MHz antennae. CMP gathers were collected at least once per common offset line, with additional CMP gathers in promising areas. Data collection parameters varied from site to site, depending on the target depth and the results of antenna tests. The locations of the GPR lines for areas A and C are inset into fig. 7.1. The maximum depth of penetration for the GPR data will be the point where brackish or slightly saline water is encountered. Rapid attenuation of the GPR wave occurs in saline water, so an abrupt drop in returned energy may indicate the location of saline water. Soldal et al. (1994) showed that a salinity of 3-4 ‰, or approximately 6-8 mS/cm, corresponded to the maximum penetration of the GPR signal. In general, the GPR sections from North Caicos Island were dominated by discontinuous reflections and diffractions, and it was not possible to associate the top of the capillary fringe with a reflection. Some very near surface reflections can be identified and will be discussed in the context of the geology of the island. Electromagnetic induction surveys were also performed on all of the GPR common offset lines using a Geonics EM-31 system in vertical dipole mode.

The GPR lines were surveyed at sites A, B and C. The other areas were not surveyed due to time constraints and poor GPR data quality. Actual surface elevations are not available, but the surveyed areas were close enough to tie together to give relative elevations. As we do not

know the exact elevation above mean sea level for any of the surveyed areas, the relative values are higher than the actual elevations would be so that there will not be any confusion between real and relative elevation.

Water table depths were collected from water wells in 4 of the 5 areas when ground penetrating radar or electrical conductivity data was being collected nearby. The water wells were also sampled for electrical conductivity (EC) and temperature. Table 7.1 shows the water depths, electrical conductivity measurements and temperature from the five wells. The electrical conductivity values ranged from 0.35 to 18.3 mS/cm at the various wells, with the lower values from relatively fresh water and the 18.3 mS/cm water coming from a well within the mixing zone between fresh and salt water. The slightly higher electrical conductivity in the well at Don's farm (B1) is presumably due to pumping earlier in the day. The water temperature ranged from 25.9 °C to 29.4 °C. Water table depths were measured

Table 7.1 Electrical conductivity (EC), temperature, water table depth below datum (WTD), relative water table elevation (WT_{elev}), and relative datum elevation for wells and water bodies on North Caicos Island.

	EC (mS/cm)	Temp. (°C)	WTD (m)	WT_{elev} (m)	Datum elev.† (m)
A1 GOV-1 well	0.36	25.9	8.400	13.825	22.225
A2 GOV-2 well	0.96	26.5	6.590	13.789	20.379
			6.570	13.809	20.379
			6.605	13.774	20.379
			6.592	13.787	20.379
B1 Don's well‡	1.38	27.0	7.620	13.695	21.315
D1 Road cut well	0.68	28.9	2.100		
E1 Hopkin's well	18.30	29.4	2.600		
Sea	off scale	29.0			
Cottage Pond	4.20				

† datum was generally the cement block around the well, 0.2 to 0.3 m above the ground

‡ well was pumped for two minutes prior to sampling to remove warm water from line

below the surrounding cement border of the wells, which was generally 10 to 30 cm above the ground surface. Don's well (B1), and the wells GOV-1 (A1) and GOV-2 (A2) were surveyed, allowing a relative water table elevation to be reported in Table 7.1. It is also important to note that July 1997 was a very dry month on North Caicos Island, with very little rain. Water levels were extremely low across the island.

7.3 Processing Characteristics

The processing steps applied to the North Caicos Island GPR common offset profiles and common midpoint gathers differs slightly from the previous datasets. In the next section, we will see that there was very little useful data on the common midpoint gathers collected on many of the lines. Of the thirty-nine CMP gathers collected on the island, only six had identifiable reflection hyperbolae which could be analysed for RMS velocities. These CMP gathers were along area A (agricultural farm), line 2 with identifiable reflections on two of the five CMP gather, and area C (swale and depressions), lines 1 and 2 with identifiable reflections on four of the six CMP gathers. The CMP gathers collected in area B (Don's Farm) had only one or two near surface reflections, and the common-offset data had very few identifiable reflections. Most of the GPR sections were not processed due to a lack of reflected energy. However, line 2 at the agricultural farm (area A) and the data from area C had sufficient reflections to warrant further processing. Constant Q compensation was not included in the processing stream as the GPR sections contain a large amount noise, making it difficult to achieve acceptable results. Time migration was also not successful on most of the lines, due to the lack of velocity information. Line 2 in area A and lines 1 and 2 in area C are the exceptions, where time migration has enhanced the near surface reflections and focussed diffractions. Normal moveout correction, spherical gain, and time-to-depth conversion all require an RMS velocity function. When velocity data was not available for lines in areas A and C, the average velocity functions computed from the known velocities in the area were used.

Many of the common offset sections contain numerous diffractions from sources above and below the surface. Sun and Young (1995) show that scattering from surface objects can be identified by migrating the sections at the airwave velocity (0.3 m/ns). Migration at the airwave velocity will focus surface scattered events, while geologic events will be overmigrated. Each of the sections was migrated at 0.3 m/ns to distinguish between surface scattered events and geologic events.

7.4 Results

7.4.1 Common Midpoint Gathers

Thirty-nine common midpoint gathers were collected on North Caicos Island using all three antenna frequencies. However, all six CMP gathers which had more than one identifiable reflection were collected using the 100 MHz antennae. Direct wave velocities ranged from 0.083 to 0.206 m/ns where the direct wave was strong enough to be identifiable. These numbers are quite different from the RMS velocities determined from the reflection hyperbola, which ranged from 0.074 to 0.104 m/ns with a mean and standard deviation of 0.091 ± 0.008 m/ns. Because of the difference between the direct wave velocities and the RMS velocities, we chose not to use the direct wave velocities when processing the data, as the occasional large difference caused problems when converting the RMS velocities to interval velocities. Figure 7.2 shows the RMS velocity profiles for the six CMP gathers which had identifiable reflections. At the Agricultural farm (area A), only two of the fifteen CMP gathers contained reflection hyperbola (fig. 7.2A). These two CMP gathers were two of five collected along line 2, and their RMS velocity functions were used in the processing stream for that line. The average of these velocity functions was computed to use on line 1 in area A, as none of the CMP gathers on line 1 contained reflections. Figure 7.2B shows the RMS velocity functions for lines 1 and 2 in the swale between ridges (area C). Four of the six CMP gathers collected on lines 1 and 2 contained reflection hyperbola whose velocity functions were used to process these lines. The other six CMP gathers collected in the swale

Figure 7.2 RMS velocity profiles from A area A - line 2, B area C - lines 1 and 2. The bold lines marked with the triangles ($\blacktriangle$) on each graph are the average RMS velocity functions for each set of velocities.

area did not have any interpretable reflections, and so the average function from lines 1 and 2 was used to process lines 3 to 6 in the swale area. The RMS velocity functions on the whole show a decreasing trend with depth, which may be due to an increase in pore water content or an increase in the salinity of the pore water. The RMS velocity profiles for the two areas are relatively similar, with the CMP gathers from the swale area (C) achieving a greater penetration depth than at the Agricultural farm (A).

7.4.2 Common Offset Lines

The common offset GPR lines from areas B (Don's Farm), D (Road Cut) and E (Hopkin's) had very poor data quality, with very few interpretable reflections. Therefore, the GPR common offset lines from these areas will only be discussed briefly in the context of comparisons between the apparent conductivity data and the GPR data quality. In areas A and C, a reflection from the top of the capillary fringe was not detected, but some of the near surface reflections can be related to the geology of the area.

7.4.2.1 Area A - Agricultural Farm

The agricultural farm (area A) was chosen for its proximity to two wells which could be sampled, GOV-1 and GOV-2. The wells were dug as part of a government test farm which has since been abandoned. Inset A in fig. 7.1 shows the locations of lines 1 and 2 at the agricultural farm. GPR common-offset lines were collected along line 1 with the 50, 100, and 200 MHz antennae. Test lines were collected along line 2 in various frequencies and offsets, after which the 100 MHz antennae with an antenna separation of 1.5 m was chosen. The time and depth sections for line 2 are shown in fig. 7.3, as well as the apparent conductivity data. At 8 m along the line, diffractions from a large tree dominate the sections. As well, a building with a tin roof and metal mesh siding was located off the end of line 2, creating the dipping noise on the 90 m end of the line. The low amplitude, high frequency dipping events seen below 60 ns on the time section are all caused by scattering from surface objects. The

Figure 7.3 North Caicos Island, area A - line 2 (frequency - 100 MHz, antenna separation - 1.5 m). **A** Time section. **B** Depth and elevation corrected section. **C** Apparent conductivity data.

prominent feature on this line is a reflection from 25 to 60 m, with a relative elevation of 19 m. The GOV-2 well which was located from 2.5 to 5 m along the line, showed a contact between aeolian and subtidal zones at an elevation of 16.48 m, the capillary fringe/transition zone at 14.63 m and the fresh water table at 13.79 m. These contacts are too deep to be associated with the reflection at 19 m relative elevation. From observations of solid rock outcrops at 73 m along the line and a bedrock contact at 20.94 m relative elevation in the GOV-1 well 45 m to the west-southwest of the end of the line, we interpret the reflection at 19 m to be from the top of the bedrock surface. Beneath this reflection, migration has focussed a small diffraction at 35 m along the line and 18 m relative elevation (circled). The original diffraction may be been caused by a dissolution cavity or brecciated carbonate rock.

Line 1 in the agricultural farm area was collected along a relatively well travelled road perpendicular to line 2. This line was collected with all three frequencies to test their effectiveness, none of which showed more than a few weak reflections in the very near surface.

7.4.2.2 Area C - Swale and elliptical depressions

Area C is located in the swale to the north of area B, where a number of elliptical depressions were located. These depressions are surrounded by karstified limestone, filled with sandy material and roots, and are dry during the summer months, but fill with water during the rainy season. While the vegetation surrounding these depressions is some of the largest on the island, no vegetation grows within the depressions. The location of the GPR lines with respect to the depressions are shown in inset C on fig. 7.1. Line 2 is an extension of line 1 along a seldom travelled dirt road through the vegetation. Lines 3, 4 and 5 all have the same geographical location, but were collected with different frequencies. At 7 m along lines 3, 4 and 5, they intersect with line 1 at 28 m and then pass across the road into the first depression. Line 6 is perpendicular to lines 3, 4, and 5, and passes through the first depression, across the rise between two depressions, and through the next depression.

Figure 7.4 shows the time and depth sections, as well as the apparent conductivity data for lines 1 and 2. This is the only GPR section which was successfully migrated. The near surface to a relative elevation of approximately 13 m contains a number of small reflections which tend to dip toward the northeast for the first 40 m along the section. From 60 to 100 m, the reflections are somewhat chaotic between 14.5 and 15 m relative elevation, with additional northeast dipping reflections beneath. This pattern indicates that these reflections are likely part of the aeolian dune structure. At approximately 13 m relative elevation, the reflected energy drops sharply, allowing underlying reverberations to be seen. On the other side of the ridge to the south, the wells in areas A and B place the water table at approximately 13.7 to 13.8 m. Therefore, the abrupt end to the reflections may be the point where water is encountered with sufficient salinity to attenuate the GPR wave. Finally, lines 1 and 2 show two areas of decreased penetration of the radar wave associated with decreased velocity. Between 5 to 10 m and 40 to 60 m along the line, the decrease in penetration is distinguished by truncated reflection and lower frequencies. These areas may be connections between the depressions to either side of the road on which the line was collected, serving as flow channels when the depressions fill with water during the rainy season.

Lines 3, 4, 5 and 6 are all primarily within the depressions to the northwest of lines 1 and 2. We do not have a good understanding of the geology of this area, but it has been postulated (Hopkins, personal communication) that the sides of these depressions are solid Pleistocene carbonates with sand and the roots of the surrounding vegetation inside the depression to the approximate depth of the water table (fig. 7.5). From a pit in the centre of the first depression, we determined that the first 0.6 m below the surface was composed of breccia and sand, followed by roots and dry sandy material with few rocks of any size to 1.5 m below the surface. We have no direct data from deeper in the depressions. Figure 7.6 shows the time and depth GPR sections and the apparent conductivity data from line 3, which was collected using the 100 MHz antennae. Figure 7.7 shows the depth section from line 5, which was collected along the same traverse as line 3 using the 200 MHz antennae. The CMP gathers collected within the depression did not contain interpretable reflections, so the average RMS

Figure 7.4 North Caicos Island, area C - lines 1 and 2 (frequency - 100 MHz, antenna separation - 1.5 m). **A** Time section. **B** Depth and elevation corrected section. **C** Apparent conductivity data.

Figure 7.5 Proposed cross-section through one of the depressions in area C (Hopkins, personal communication). From the pit near the centre of one depression, we found rocks and sand to 0.6 m below the surface, followed by sand and roots to 1.5 m. Below 1.5 m within the depression, we have no direct evidence of the geology. The Pleistocene carbonates surrounding the depressions have aeolian deposits overlying a subtidal zone. The depth of the contact between these two zones is unknown.

velocity function from lines 1 and 2 was used to convert the GPR data from lines 3 and 5 to depth. The surface features along lines 3, 4 and 5 are as follows. The line passes through a cleared path in the trees from 0 to 7 m, onto the road where lines 1 and 2 were collected from 7 to 11 m, and then back into a treed area from 11 to 15 m. At 15 m, the line begins to drop into the depression, where the antenna were occasionally suspended above the ground to avoid damage due to broken rock. From 22 to 35 m, the lines are within the depression, on relatively flat, sandy ground. On fig. 7.6 and 7.7, there is a strong reflection at a relative elevation of 14.5 m from 0 to 7 m along the sections. This reflection has a polarity which suggests that its source could be from the transition from unsaturated to saturated zones. We have two sources of information concerning the relative elevation of the water table. The first is the wells to the southeast. In Don's well (B1), which is over 600 m away from line 3, the water table was found at 13.70 m, with a capillary fringe / transition zone at 13.90 m. In the GOV-2 well at the agricultural farm, the water table was at 13.78 m with the capillary fringe / transition zone at 14.63 m. This information seems to indicate that it would not be unreasonable to associate the reflection at 14.5 m with the capillary fringe. However, we also know that within the depression, the subsurface was dry to a relative elevation of 13.2 m.

Figure 7.6 North Caicos Island, area C - line 3 (frequency - 100 MHz, antenna separation - 1.5 m). **A** Time section. **B** Depth and elevation corrected section. **C** Apparent conductivity data.

Figure 7.7 North Caicos Island, area C - line 5 (frequency - 200 MHz, antenna separation - 0.75 m). Depth and elevation corrected section. Line 5 has the same geographic location as line 3.

Therefore, unless there is a significant drop in water table between the surrounding land and the depression, it is unlikely that this reflection is actually from the capillary fringe.

As the line dropped into the depression, the ground became very broken and rough. This is seen in the GPR sections as a broken and chaotic direct wave return between 15 and 21 m. At approximately 23 m on the 200 MHz section (fig. 7.7), the direct wave appears to split into two separate events. We interpret the dipping reflection to be a continuation of the Pleistocene limestone beneath the sand at the edge of the depression. The reflection appears to fracture and then disappear by 27 m along the line. A similar but much less distinct reflection is seen on the 100 MHz section. Other than this dipping reflection, the only energy returning from the subsurface to the receiver within the depression is the diffraction from the open pit in the centre of the depression. The other large diffractions were identified as surface scattering events from the edges of the depression by migrating the section at 0.3 m/ns.

Figure 7.8 shows the 100 MHz GPR time section, depth section and apparent conductivity data from line 6, which passes through two of the depressions. At 7 m along line 6, it intersects the 31 m mark on lines 3, 4, and 5. Along line 6, the edges of the depressions are not as clearly defined as on line 3. As well, line 6 has no apparent reflections from within the depressions. As the ground surface rises between the two depressions, we see a weak reflection between 15 and 20 m along the line, at a relative elevation of 13 m. This is at the same elevation as the disappearance of identifiable reflections along lines 1 and 2, which may indicate a similar source. As on line 3, migration of the section at the airwave velocity indicates that the source of the diffraction below 13 m relative elevation is scattering from surface objects such as the side of the depression.

7.4.3 Apparent Conductivity data

The apparent conductivity data from North Caicos Island was measured using a Geonics EM-31 in the vertical dipole position, with the coils approximately 0.90 m above the ground surface. Table 7.2 summarizes the apparent conductivity data collected on North Caicos Island. At the agricultural farm (area A), the conductivities are between 5.62 and 12.52 mS/m, with the highest values along the main road where line 1 was collected and line 2 started. Along line 2 (fig. 7.3C), the conductivity begins to decrease at approximately 30 m, which correlates to the beginning of the subsurface reflection labelled A in fig. 7.3A and 7.3B. From 40 to 90 m, the apparent conductivities fluctuate around 7 mS/m, with the higher values between 58 and 72 m. On the GPR sections for line 2, the area between 60 and 70 m contains the coherent noise which occurs when reflected energy is very low or absent.

The apparent conductivities from Don's farm (area B) are very low for the island, with values ranging from 5.16 to 6.68 mS/m. This range neglects the apparent conductivity values measured between 60 and 80 m, which were affected by scrap metal, a wind powered generator and the metal anchor lines for the windmill located to the side of the line. The apparent conductivity data implies that this area should have GPR results similar to line 2 at

Figure 7.8 North Caicos Island, area C - line 6 (frequency 100 MHz, antenna separation - 1.0 m). **A** Time section. **B** Depth and elevation corrected section. **C** Apparent conductivity data.

Table 7.2 Mean, standard deviation and range of apparent conductivities measured using a Geonics EM-31 electromagnetic induction system.

Area	line	Apparent Conductivity (mS/m)		
		mean	standard deviation	range
A	line 1	11.01	0.78	9.76 - 12.52
A	line 2	8.22	1.52	5.62 - 10.93
B	line 1	6.05	0.37	5.16 - 6.68
C	lines 1, 2	13.61	1.11	11.42 - 15.89
C	lines 3, 4, 5	21.40	7.34	12.97 - 31.59
C	line 6	22.56	3.80	17.67 - 29.30
D	line 1	17.48	3.53	11.23 - 24.75
D	line 2	30.75	4.27	26.40 - 39.34
E	line 1	142.11	20.81	102.54 - 172.12

the agricultural farm. Unfortunately, the various metallic parts between 60 and 80 m created a large amount of noise in the GPR data, obscuring subsurface reflections more than 50 ns below the surface. Reflections similar to those seen in the very near surface on lines 1 and 2 in area C are identifiable in parts of the data, but are not as strong as on the area C data. The CMP gathers collected in area B show evidence of hyperbolic reflections, but are also affected by the noise from the metallic objects to the side of the line.

In the swale to the north of Don's farm (area C), the apparent conductivities had a large range of values. Along lines 1 and 2, the apparent conductivities ranged from 11.42 to 15.89 mS/m (fig. 7.4C), which are in the range of values found in poor data areas on line 2 at the agricultural farm (area A). The rise and fall of conductivities along this line seem to be inversely related to the topography, implying that the apparent conductivity values in this area are greatly influenced by the distance from the saline water beneath the island. The most interesting data is from the depressions (fig. 7.6C and 7.8C) which show a doubling of the apparent conductivities within the depressions. For lines 3, 4, and 5 (fig. 7.6C), the edge of

the first depression was at 15 m, which is where the conductivity begins to increase. Along line 6 (fig. 7.8C), the line is within the first depression from 0 to 18 m, crosses a rise between two depressions from 18 to 39 m, and is within a second depression from 39 to 55 m. The edges of the second depression were much less distinct than the first, as can be seen by the topographic change in the GPR depth section (fig. 7.8B).

Area D is near a large road cut, where the road between the towns of Kew and Whitby runs through one of the larger aeolian dunes. We collected GPR and EM data over the dune (line 1), with the expectation of comparing the GPR data to the dune stratigraphy exposed in the road cut. No interpretable reflections were seen on line 1, and the GPR system did not deal with the heat of the exposed dune very well. The apparent conductivities were between 11.23 and 24.75 mS/m, which are comparable to the values in the swale (area C). We also collected GPR and EM data near well D1 to the north of the dune and road cut. The water table was only approximately 2.1 m below the ground surface, which itself was slightly lower than the 0 m mark on line 1. The GPR data was collected on a sealed road near the well, and shows only one reflection which we believe comes from the bottom of the road fill. No geologic reflections were seen on the data.

The GPR and EM data collected in area E runs over a Holocene beach ridge towards the ocean. The GPR data has a few reflections at the top of the beach ridge, which corresponds to the low in the apparent conductivity values in fig. 7.9. The apparent conductivity values measured in this area are the highest of all the sites, which is understandable given the salinity in the well in area E and the proximity to the beach.

Almost all of the apparent conductivity data collected on North Caicos Island show an inverse relationship to the topography. In fig. 7.10, the estimated water table depth versus apparent conductivity has been graphed for areas A, B, and C. Along the lines in these areas, we have relative elevation data, and areas A and B have water table elevation data. As well, Table 7.1 shows that the water collected in areas A, B, and D all have similar conductivities.

Figure 7.9 North Caicos Island, area E - line 1, apparent conductivity data. This line runs over a Holocene beach ridge which peaks at approximately 15 m.

For area C, we used the water table elevations from area B (Don's well - B1) to determine the approximate water table depth below the surface, and have assumed that the water has a similar electrical conductivity. From the pit in the first depression, we know that the water table depth determined in this way are slightly too shallow. Figure 7.10 also includes a single point from area D, which is the apparent conductivity measured 30 m from the well. No data was collected closer to the well, as overhead powerlines affected the measurements. The graph in fig. 7.10 shows a nonlinear relationship between apparent conductivity and water table depth. The data from area B (Don's farm) seems to indicate that once the water table depth is greater than approximately 7 m, the water conductivity no longer influences the apparent conductivities measured at the surface. The effective depth of penetration for the Geonics EM-31 system in vertical dipole mode is 6 m (Geonics Ltd., 1994), which concurs with the data in fig. 7.10. The apparent conductivities measured in area B are probably the best representation of the unsaturated zone conductivities throughout the island. This has important implications to the comparison of GPR data quality to the apparent conductivity data. As the water table is within 6 m of the surface for most of the data collected on North Caicos Island, the measured apparent conductivities are largely influenced by the water table depth rather than the conductivity of the unsaturated zone. This makes a comparison between GPR data quality and apparent conductivity difficult as the majority of the changes in

Figure 7.10 Estimated water table depth versus apparent conductivity for various lines on North Caicos Island. Apparent conductivity data measured using a Geonics EM-31 in vertical dipole mode.

apparent conductivity are caused by the topography. The apparent conductivities associated with a lack of reflected energy on the GPR sections vary considerably across the island, from approximately 8.0 mS/m between 60 and 70 m on line 2 in area A, to 11.23 to 39.34 mS/m in area D. However, while area E had conductivities ranging from 102.54 to 172.12 mS/m, a few identifiable reflections were also seen which relate to the top of a Holocene ridge and the low point in the conductivity values (fig. 7.9). From this information, we cannot correlate a specific apparent conductivity value at which the GPR data quality decreases.

7.5 Summary

When the objectives of this thesis are considered, the GPR data collected on North Caicos Island is disappointing. We were unable to identify reflections from the top of the capillary fringe on any of the GPR sections. The ground penetrating radar sections contain chaotic reflections and numerous diffractions, with very few continuous reflections. Of the thirty-nine common midpoint gathers collected on the island, only six contained significant reflection hyperbola to warrant further processing and discussion. The RMS velocities were all measured from reflections within the unsaturated zone, and ranged in value from 0.0740 to 0.1037 m/ns. These values are significantly lower than RMS velocities measured in the unsaturated zone at the Gilby Waterflood and Strathroy field sites. They are also lower than the value of 0.12 m/ns given in Table 2.1 from Davis and Annan (1989) for limestone. The reasons for this discrepancy is not currently known. All of the RMS velocity functions show a decreasing trend with depth, though the CMP gathers collected in area C have a greater penetration depth than the CMP gathers collected in area A.

Many of the common offset lines collected on the island had very few continuous reflections. The noted exceptions have been interpreted to be the top of the bedrock in area A, and depositional features within the aeolian dunes in area C. The abrupt loss of these reflections at approximately 13 m relative elevation may be the point where water is encountered with sufficient salinity to attenuate the GPR wave. GPR data collected within the depressions in area C appeared to contain only surface scattering in the form of diffractions from the depression edges. This may be due to either an increase in conductivity, or a lack of dielectric contrast from the sand within the depressions.

The apparent conductivities measured on North Caicos Island range from 5.16 to 172.12 mS/m. When the water table depth is less than 6 m below the surface, there is a nonlinear, inverse relationship between water table depth and apparent conductivity. This is expected with an underlying conductive layer such as saline water. The apparent conductivities

measured in area B, where the water table depth was greater than 6.5 m, are a good representation of the conductivity of the unsaturated zone on the island. These apparent conductivities ranged from 5.16 to 6.68 mS/m with a mean of 6.05 mS/m. Therefore, it is unlikely that the lack of reflections from the unsaturated zone is caused by the conductivity of the unsaturated zone. The poor data quality is likely due to a lack of spatially continuous reflectors and the presence of numerous diffractors in the form of dissolution cavities and brecciated carbonate rock.

8 Conclusions

The objective of this thesis was to investigate the accuracy of water table elevation estimates determined from ground penetrating radar. The accuracy of an estimate is dependent on the magnitude of the uncertainties introduced by various sources of error. In Chapter 3, we identified and quantified the sources of error in water table elevation estimates from ground penetrating radar data. These included

- uncertainty in the capillary fringe height estimate
- uncertainty in the depth to a reflector due to
 - the time sampling interval
 - velocity uncertainty on a single CMP gather
 - velocity uncertainty due to heterogeneity and spatially sparse CMP locations
- surface elevation uncertainties
 - at each survey station due to the measurement uncertainty of the survey equipment
 - due to interpolation of surface elevations between survey stations

For the GPR data collected at the Gilby Waterflood and Strathroy field sites, the magnitudes of these uncertainties, and the combined effect of these uncertainties on the water table elevation estimates were determined. Table 5.7 and 6.5 respectively show that the measurable uncertainty which had the greatest contribution to the uncertainty in the water table elevation estimates was the uncertainty in depth to the capillary fringe reflection due to velocity uncertainty. The most difficult value to estimate was the height of the capillary rise above the water table which could be expected. Without more information about the subsurface than was available at any of the three sites, the capillary rise used to convert capillary fringe elevation estimates to water table elevation estimates can only be estimated

to the nearest known grainsize. The smaller the grainsize, the larger the possible range of capillary rise values which could apply.

At the Gilby Waterflood site, the differences between the water table elevation estimates determined from the ground penetrating radar data ($WT_{\text{elev}}(\text{GPR})$) and the water table elevation measurements from the wells ($WT_{\text{elev}}(\text{well})$) were generally less than the uncertainty in the GPR elevation estimates (Tables 5.8 and 5.9). The GPR water table elevation estimates had uncertainties ranging from ± 0.14 to ± 0.21 m, with differences between $WT_{\text{elev}}(\text{GPR})$ and $WT_{\text{elev}}(\text{well})$ ranging from 0.02 m for the closest values to 0.23 m for the worst points. In comparison, the water table elevation estimates from the GPR data collected at the Strathroy field site did not compare well with the water table elevation measurements from the wells. The differences between the GPR estimates and the well measurements were between 0.59 and 0.78 m. As well, the water table contour map developed from the GPR water table elevation estimates had a dip which was opposite to the ground water flow direction determined from the monitoring well data. While changes in grainsize along the GPR lines could account for the discrepancies between monitoring well and GPR data, there is currently no evidence that this is occurring.

Determining the velocity of the radar wave in the subsurface has proven to be one of the more challenging and important tasks when determining the accuracy of water table elevation estimates from ground penetrating radar. For the Gilby Waterflood and Strathroy field sites, the mean RMS velocity at the capillary fringe was 0.129 and 0.108 m/ns with a variation of 9.5 % and 1 % respectively. However, we were able to determine that the velocities measured at the Gilby Waterflood field sites are correlated to a maximum distance of 20 m. Since the CMP gathers were collected every 10 m along the common offset profiles at the Gilby site, it is likely that the velocity variations at the site have been captured. On the other hand, potential velocity variations exist at the Strathroy site, as only three CMP gathers were collected for more than 400 m of common-offset data. In either case, the velocity variations are much less than the 36 % variation in the RMS velocity at the top of the capillary fringe

which was calculated from theoretical variations in the saturated hydraulic head, ψ_s , and the pore size distribution index, b . They are also less than the 30 % variation in velocities reported by Greaves et al. (1996). The RMS velocities determined from CMP gathers collected on North Caicos Island were lower than at either of the other two sites, ranging from 0.074 to 0.104 m/ns with a mean of 0.091 m/ns. These values are also much lower than the 0.12 or 0.15 m/ns reported in Table 2.1 for either limestone or dry sand, both of which were encountered.

Finally, we have the apparent conductivity data measured using a Geonics EM-31 electromagnetic induction system. The Gilby Waterflood data shows a direct correlation between a decrease in reflected energy, an increase in apparent conductivity and increase in fine material and vegetation at the surface. The data indicates that at Gilby, when the apparent conductivity exceeds 11 mS/m, the reflected energy decreases significantly. The apparent conductivity data collected on North Caicos Island was not so forthcoming. When the water table was less than 6 m below the surface, the apparent conductivity values were strongly influenced by the water table. We found that there was a nonlinear, inverse relationship between estimated water table depth and apparent conductivity for water table depths less than 6 m caused by the conductivity of the brackish water. When the water table depths exceeded 6 m, the apparent conductivities were approximately 6 mS/m. From this, we can deduce that the conductivity of the unsaturated zone on North Caicos Island is typically around 6 mS/m. It is therefore unlikely that the difficulties experienced with the GPR data quality are due to the conductivity of the unsaturated zone. The poor GPR data quality is more likely due to a lack of spatially continuous reflectors and the presence of diffractions from above and below the surface.

A number of conclusions can be drawn from the results presented in this thesis. Accurate water table elevation estimates rely heavily on three pieces of information, surface elevations along the GPR lines, accurate knowledge of the subsurface velocities, and an understanding of the geologic material present at the water table in sufficient detail to estimate the capillary

fringe height above the water table. Surface elevations not only provide information about the topography which can be useful in itself, but they also provide a common value for comparing GPR depth sections to the well data. Surveying the GPR stations provides the most accurate surface elevations, reducing this component of the uncertainty in the final estimates. When determining the survey station spacing to use, it is important to consider the interpolation error which could result when the survey station spacing is greater than the GPR station spacing. If surveying of the GPR stations isn't feasible, the data from the Strathroy field site has shown that using kriging to interpolate survey points from the site in general can provide an acceptable alternative. The Strathroy data also emphasizes the need for surface elevations. Without surface elevations, it would have been much more difficult to identify that the dip of the water table elevations estimated from the GPR data was in exactly the opposite direction to the dip of the water table determined from the monitoring wells.

Accurate knowledge of the subsurface velocities along the GPR common offset lines is essential for determining accurate elevation estimates, no matter what the target. Errors in the velocity profile along a line can cause errors in the depth section, as well as problems when processing the data. Unfortunately, there are no simple rules for determining the frequency of CMP gathers needed at a particular site. Using geostatistical analysis of the velocities collected at the Gilby Waterflood field site, we were able to determine that by collecting CMP gathers every 10 m along the common offset profiles we had captured the major velocity variations in the area. On the other hand, only three CMP gathers along 400 m of common offset data at the Strathroy site has left us very uncertain about potential velocity variations which have not been captured by the CMP gathers. During data collection, we would suggest collecting the common offset GPR data initially, and examining the sections to determine where CMP gathers are needed. Changes in reflector characteristics such as strength, continuity and dip may indicate changes in the subsurface velocity which should be addressed with a CMP gather. As well, changes in surface topography will cause changes in the two-way travel time and depth to a reflection, which may change the velocity

profile.

In order to determine the height of the capillary rise above the water table which will be used to convert capillary fringe elevation estimates to water table elevation estimates, the geologic material at the water table must be known or estimated. The greater the certainty in the sediment type, the smaller the uncertainty in the estimated capillary rise, and the greater the accuracy of the final water table elevation estimate. As well, if the subsurface is heterogeneous, the capillary rise can be vary spatially. This was seen in the Gilby Waterflood data, which was converted from capillary fringe elevations to water table elevations using a single capillary rise for the site. The water table elevation contour maps have a number of anomalous peaks and valleys which were interpreted as variations in the height of the capillary rise above the water table due to grainsize variations.

When the GPR lines have been surveyed, the velocities are well known and the height of the capillary fringe above the water table can be estimated fairly accurately, the uncertainty in the water table elevations determined from ground penetrating radar data can be as low as ± 0.10 to 0.15 m. In areas where velocity variations and changing capillary fringe height occur, this minimum uncertainty will rise. The Gilby Waterflood data generally matched the monitoring well measurements and showed uncertainties ranging from ± 0.14 to 0.21 m with the uncertainty in depth due to velocity uncertainties and the uncertainty in the height of the capillary fringe as the largest contributing factors. On the other hand, the Strathroy data showed the pitfalls in this technique, with a large mismatch between the GPR and monitoring well water table elevations.

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**Appendix A: Ground Penetrating Radar Data
collected at Gilby Waterflood, Alberta**

Figure A.1 Depth and elevation corrected section from line 1, 100 MHz antennae. Station spacing = 0.25 m, antenna offset = 1.00 m.

Figure A.2 Depth and elevation corrected sections from line 1, 200 MHz antennae. Station spacing = 0.25 m, antenna offset = 0.75 m. **A** Section collected June 17, 1997. **B** Section collected June 18, 1997.

Figure A.3 Depth and elevation corrected sections from line 2. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.4 Depth and elevation corrected sections from line 3. A 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. B 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.5 Depth and elevation corrected sections from line 4. A 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. B 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.6 Depth and elevation corrected sections from cross line 10. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.7 Depth and elevation corrected sections from cross line 20. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.8 Depth and elevation corrected sections from cross line 30. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.9 Depth and elevation corrected sections from cross line 40. A 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. B 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.10 Depth and elevation corrected sections from cross line 50. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.11 Depth and elevation corrected sections from cross line 70. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.12 Depth and elevation corrected sections from cross line 85. **A** 100 MHz section, station spacing = 0.25 m, antenna separation = 1.00 m. **B** 200 MHz section, station spacing = 0.25 m, antenna separation = 0.75 m.

Figure A.13 Common Midpoint Gathers from line 1, 100 MHz antenna. Step size = 0.20 m.

Figure A.14 Common Midpoint Gathers from line 1, 200 MHz. Step size = 0.20 m. **A** CMP gathers collected June 17, 1997. **B** CMP gathers collected June 18, 1997.

Figure A.15 Common Midpoint Gathers from line 2. A 100 MHz section, step size = 0.20 m.
 B 200 MHz section, step size = 0.20 m.

Figure A.16 Common Midpoint Gathers from line 3. **A** 100 MHz section, step size = 0.20 m.
B 200 MHz section, step size = 0.20 m.

Figure A.17 Common Midpoint Gathers from line 4. **A** 100 MHz section, step size = 0.20 m.
B 200 MHz section, step size = 0.20 m.

Figure A.18 Common Midpoint Gathers from cross lines 10, 20, and 30. **A** 100 MHz section, step size = 0.20 m. **B** 200 MHz section, step size = 0.20 m.

Figure A.19 Common Midpoint Gathers from cross lines 40, 50, and 70. **A** 100 MHz section, step size = 0.20 m. **B** 200 MHz section, step size = 0.20 m.

Figure A.20 Common Midpoint Gathers from cross line 85. **A** 100 MHz section, step size = 0.20 m. **B** 200 MHz section, step size = 0.20 m.

**Appendix B: Ground Penetrating Radar Data
collected at Strathroy, Ontario**

Figure B.1 Depth and elevation corrected sections from lines 2 and 4 at Strathroy. Antenna frequency = 200 MHz, station spacing = 0.10 m, antenna offset = 0.75 m. **A** Line 2. **B** Line 4.

Figure B.2 Depth and elevation corrected sections from lines 6 and 8 at Strathroy. Antenna frequency = 200 MHz, station spacing = 0.10 m, antenna offset = 0.75 m. **A** Line 6. **B** Line 8.

Figure B.3 Depth and elevation corrected sections from line 10 and the cross lines at 4, 29 and 59 m along the main lines at Strathroy. Antenna frequency = 200 MHz, station spacing = 0.10 m, antenna offset = 0.75 m. A Line 10. B Cross lines at 4, 29 and 59 m.

Figure B.4 Common midpoint gathers from Strathroy, Ontario. Antenna frequency = 200 MHz. **A** CMP gathers acquired along line 10. Step size for CMP @ 7.5 = 0.1 m, step size for CMP @ 45 = 0.2 m. **B** CMP gather acquired along line 6. Step size = 0.2 m.

**Appendix C: Ground Penetrating Radar Data
collected on North Caicos Island,
Turks and Caicos Islands, British West Indies**

Figure C.1 Sections from lines 1 and 2 acquired at area A, North Caicos Island. Antenna frequency = 100 MHz, station spacing = 0.25 m. **A** Time section from line 1, antenna separation = 1.0 m. **B** Depth and elevation corrected section from line 2, antenna separation = 1.5 m.

Figure C.2 Common midpoint gathers from area A, North Caicos Island. Antenna frequency = 100 MHz. A CMP gather acquired along line 1. Step size = 0.2 m. B CMP gathers acquired along line 2. Step size = 0.2 m.

Figure C.3 Time sections from line 1 acquired at area B, North Caicos Island. Station spacing = 0.25 m. **A** 100 MHz section, antenna separation = 1.0 m. **B** 50 MHz section, antenna separation = 2.0 m.

Figure C.4 Common midpoint gathers from area B, North Caicos Island. **A** CMP gathers acquired using the 100 MHz antennae. Step size = 0.2 m. **B** CMP gathers acquired using the 50 MHz antennae. Step size = 0.2 m.

Figure C.5 Depth and elevation corrected section, lines 1 and 2, area C, North Caicos Island. Antenna frequency = 100 MHz, station spacing = 0.25 m, antenna separation = 1.5 m.

Figure C.6 Depth and elevation corrected section, lines 3 and 4, area C, North Caicos Island. **A** Line 3, antenna frequency = 100 MHz, station spacing = 0.25 m, antenna separation = 1.5 m. **B** Line 4, antenna frequency = 50 MHz, station spacing = 0.25 m, antenna separation = 2.0 m.

Figure C.7 Depth and elevation corrected section, lines 5 and 6, area C, North Caicos Island.
A Line 5, antenna frequency = 200 MHz, station spacing = 0.25 m, antenna separation = 0.75 m. **B** Line 6, antenna frequency = 100 MHz, station spacing = 0.25 m, antenna separation = 1.0 m.

Figure C.8 Common midpoint gathers from area C, North Caicos Island. **A** CMP gathers acquired along lines 1 and 2 using the 100 MHz antennae. Step size = 0.2 m. **B** CMP gathers acquired along lines 3, 4, 5 and 6. Lines 3 and 6 were acquired using the 100 MHz antennae. Line 4 was acquired using the 50 MHz antennae and line 6 was acquired using the 20 MHz antennae. Step size = 0.2 m.

Figure C.9 Time sections from area D, North Caicos Island. Station spacing = 0.25 m, antenna frequency = 100 MHz. **A** Line over road cut. Antenna separation = 0.75 m. **B** Line along road near well. Antenna separation = 1.0 m, well at 0 m, overhead power lines at 10 m.

Figure C.10 Common midpoint gather from line along road near well, area D, North Caicos Island. Antenna frequency = 100 MHz. Step size = 0.2 m.

Figure C.11 Time sections acquired from area E (Holocene beach ridge), North Caicos Island. Station spacing = 0.25 m. **A** 100 MHz section, antenna separation = 0.75 m. **B** 200 MHz section, antenna separation = 0.75 m.

Figure C.12 Common midpoint gathers, area E (Holocene beach ridge), North Caicos Island. Antenna frequency = 200 MHz. Step size = 0.2 m.