Basin Morphology, Sedimentology, and History of a Small Proglacial Lake, Matanuska Glacier, Alaska

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**TABLE OF CONTENTS**

<table>
<thead>
<tr>
<th>Section</th>
<th>Page #</th>
</tr>
</thead>
<tbody>
<tr>
<td>Introduction</td>
<td>4</td>
</tr>
<tr>
<td>Field Site</td>
<td>7</td>
</tr>
<tr>
<td>Methods</td>
<td>11</td>
</tr>
<tr>
<td>Results</td>
<td>16</td>
</tr>
<tr>
<td>Discussion</td>
<td>25</td>
</tr>
<tr>
<td>Facies</td>
<td>25</td>
</tr>
<tr>
<td>Ground-Penetrating Radar</td>
<td>29</td>
</tr>
<tr>
<td>Basin Morphology</td>
<td>33</td>
</tr>
<tr>
<td>Rhythmites</td>
<td>33</td>
</tr>
<tr>
<td>Sedimentation in Similar Lakes</td>
<td>36</td>
</tr>
<tr>
<td>Lake Drainage</td>
<td>37</td>
</tr>
<tr>
<td>Lake History</td>
<td>39</td>
</tr>
<tr>
<td>Conclusion</td>
<td>41</td>
</tr>
<tr>
<td>Acknowledgments</td>
<td>42</td>
</tr>
<tr>
<td>Works Cited</td>
<td>43</td>
</tr>
<tr>
<td>Figures</td>
<td></td>
</tr>
<tr>
<td>Fig. 1: Sediment dispersal mechanisms</td>
<td>6</td>
</tr>
<tr>
<td>Fig. 2: Location map</td>
<td>8</td>
</tr>
<tr>
<td>Fig. 3: 2003 Aerial photograph</td>
<td>8</td>
</tr>
<tr>
<td>Fig. 4: 1962 Aerial photograph of the glacier margin</td>
<td>9</td>
</tr>
<tr>
<td>Fig. 5: 1993 Aerial photograph of the glacier margin</td>
<td>9</td>
</tr>
<tr>
<td>Fig. 6: 1966 Aerial photograph zoomed in on the lake studied</td>
<td>10</td>
</tr>
<tr>
<td>Fig. 7: 1993 Aerial photograph zoomed in on the lake studied</td>
<td>10</td>
</tr>
<tr>
<td>Fig. 8: Laminae visibility with water content</td>
<td>13</td>
</tr>
<tr>
<td>Fig. 9: Location of pits, cores and GPR transects</td>
<td>14</td>
</tr>
<tr>
<td>Fig. 10: How to read a GPR profile</td>
<td>15</td>
</tr>
<tr>
<td>Fig. 11: Photo of the upper section of Pit C</td>
<td>17</td>
</tr>
<tr>
<td>Fig. 12: Photo of Pit A</td>
<td>17</td>
</tr>
<tr>
<td>Fig. 13: Graph of sediment thickness and laminae counts</td>
<td>18</td>
</tr>
<tr>
<td>Fig. 14: Photo of a graded sand bed</td>
<td>19</td>
</tr>
<tr>
<td>Fig. 15: Photo of a deformed layer</td>
<td>19</td>
</tr>
<tr>
<td>Fig. 16: Photo of a load structure</td>
<td>20</td>
</tr>
<tr>
<td>Fig. 17: Photo of convoluted bedding</td>
<td>20</td>
</tr>
<tr>
<td>Fig. 18: Stratigraphic columns</td>
<td>21</td>
</tr>
<tr>
<td>Fig. 19: Topographic map</td>
<td>22</td>
</tr>
<tr>
<td>Fig. 20: GPR profiles</td>
<td>23-24</td>
</tr>
<tr>
<td>Fig. 21: Facies locations</td>
<td>27</td>
</tr>
<tr>
<td>Fig. 22: Photo of a modern push moraine</td>
<td>28</td>
</tr>
<tr>
<td>Fig. 23: Interpretations of GPR profiles</td>
<td>30-32</td>
</tr>
<tr>
<td>Fig. 24: Photo of possible varves and daily rhythmites</td>
<td>35</td>
</tr>
<tr>
<td>Fig. 25: Photo of a frost-cracked wedge</td>
<td>38</td>
</tr>
<tr>
<td>Fig. 26: Schematic history</td>
<td>40</td>
</tr>
</tbody>
</table>
Abstract

This study examines the basin morphology, sedimentology, and geologic evolution of a small proglacial lake, which is now almost completely drained, located less than a kilometer north of the terminus of the Matanuska Glacier, Alaska. The methods of analysis include pits dug into drained lake deposits, Livingston cores taken from two small lake remnants, and ground-penetrating-radar data (GPR) collected during the summer of 2003. The proglacial lake was approximately 120 m long by 85 m wide and had a maximum depth of 8 m as determined by a topographic survey of the modern lake bed. The lake developed in a topographic low as the Matanuska Glacier receded from a recessional moraine that probably formed 50-100 years ago; analysis of aerial photography indicates the lake drained most of its water sometime after 1966. The basin morphology was complicated by the development of a push moraine during a small re-advance early in lake development. Due to proximity of the lake to the glacier, it is assumed that the majority of water input was derived from glacial melt. Analysis of GPR transects, basin topography, and stratigraphy reveals three distinct facies: (1) a deltaic facies consisting of prograding forsets and interlayered laminated silt and gravelly-sand deformed by loading structures; (2) a basinal facies consisting of finely-laminated fine sand, silt and clay; and (3) a push-moraine facies consisting of diamicton found in a subtle ridge. Horizontal laminations, graded beds, and leveled topography, along with some inclined laminations and draping of the basinal facies onto topographic highs, indicate basinal sedimentation was controlled by a combination of density underflows and suspension settling from interflows/overflows. The influence of underflows and lack of consistent clay partings make annual varves indistinguishable. Since the draining of the lake basin, small-scale channel development and polygonal freezing features have appeared on the surface.

Keywords: Matanuska Glacier, lake, sedimentation, proglacial, ground penetrating radar
INTRODUCTION

The Matanuska Glacier has been the study site for many researchers, and a wealth of literature has been produced in the fields of geology and glaciology from studies at the Matanuska. The topics of these studies have varied widely, from the origin of debris-rich bands in basal ice (Lawson et al., 1998; Ensminger et al., 1999; Ensminger et al., 2001) and the characteristics and pebble fabrics of diamictons (Lawson, 1979a; Lawson, 1981), to seismic and radar studies of glacial ice (Arcone et al., 1995; Baker et al., 2003) and others (Williams and Ferrians, 1961; Lawson and Kulla, 1978; Lawson, 1979b; Lawson, 1979a, 1982; Strasser et al., 1996; Denner et al., 1999; Alley et al., 2003; Pearce et al., 2003). However, only one of these studies has dealt with any of the many small semi-permanent or short-lived lakes that litter the glacier margin (Lawson, 1979b). This study is designed to add to the wealth of information that has come out of the Matanuska Glacier by providing an in-depth sedimentological and morphological history of a small proglacial lake.

Many sedimentological studies have been done on large modern proglacial lakes (Gustavson, 1975; May, 1977; Shaw et al., 1978; Lambert and Hsu, 1979; Smith, 1981; Pickrill and Irwin, 1983; Desloges and Gilbert, 1998; O'Brien and Pietraszek-Mattner, 1998) and their Pleistocene counterparts (Ashley, 1975; Shaw, 1975; Gibbard and Dreimanis, 1978). Fewer researchers have investigated the smaller and often short-lived lakes that are common on glacial margins (Donnelly and Harris, 1989; Desloges, 1994; Syverson, 1998). The majority of these studies have focused on the identification of varves in rhythmically laminated basinal sediment. The term “varve” was originally defined by DeGeer as a couplet of light and dark laminae that represents a time span of
one year (DeGeer, 1912). There are several limnological criteria, such as the existence of suspended matter in the water column and water stratification for part of the year, that must be present in order for rhythmic laminations to be true varves (Sturm, 1979). Rhythmic light and dark laminations are not necessarily varves (Lambert and Hsu, 1979; Syverson, 1998) and there may be more than one light-dark couplet in an annual deposit (Shaw et al., 1978; Pickrill and Irwin, 1983; Desloges, 1994).

The physical character of the lake and the input water dictates the character of the sedimentary deposits that will result. Perhaps the most important physical parameter controlling sedimentation is the mechanism by which sediment is dispersed through the lake; sediment may be dispersed through overflows/interflows, underflows, equal-density mixing, or some combination of the three (Ashley et al., 1985) (Fig. 1). Overflows and interflows tend to produce thin deposits of silt and clay that thin distally, and are found at all elevations in the lake basin (Ashley et al., 1985; Syverson, 1998). Underflow deposits are essentially extensions of a deltaic bottomset, are found exclusively in topographic lows of the lake, and may contain sand, silt and clay (Smith, 1981; Ashley et al., 1985; Weirich, 1986). Deposits produced by equal-density mixing will be similar to overflow/interflow deposits but will lack any stratification. Other variables that affect sediment characteristics are suspended sediment concentration, lake stratification, temperature of the lake and input water, source of input (glacial vs. non-glacial), distance of the lake from the glacier, distance from the shore, depth, lake-bottom topography and amount of supra-glacial debris (Lambert and Hsu, 1979; Smith, 1981; Ashley et al., 1985; Weirich, 1986). Temporal variations in the dispersal mechanisms and input characteristics produce variations in sedimentation, which are responsible for the
rhythmic lamination that is often found in glacial lacustrine deposits (Ashley et al., 1985). These variations operate on several different time-scales, from slump-related density currents lasting minutes to daily fluctuations in discharge and suspended sediment concentration to seasonal summer/winter alternations to changes in glacial position or climate that may take years or decades (Lambert and Hsu, 1979; Sturm, 1979; Smith, 1981; Ashley et al., 1985; Desloges, 1994; Johnson, 1997b; Syverson, 1998).

![Diagram of different dispersal mechanisms](image)

Fig. 1: Variations in deposits of glacier-fed lakes caused by different dispersal mechanisms. Figure from (Ashley et al., 1985).

Most sedimentological studies of lakes have relied on cores to obtain data; however, in cases where water levels were lowered or the lake drained completely, outcrops have provided insights into three-dimensional facies variations (Donnelly and Harris, 1989; Syverson, 1998). Such is the case in this study; the lake had drained almost completely, making it possible to utilize pit digging and ground-penetrating radar. Ground-penetrating radar has been shown to be an effective method of viewing sedimentary structures and stratigraphy when continuous outcrops are not possible (Davis and Annan, 1989; Lonne and Lauritsen, 1996; Smith and Jol, 1997; Van Dam and Schlager, 2000).
Field Site

The Matanuska Glacier (61°47′N, 147°45′W) has its origin in the ice fields of the Chugach Mountains of south-central Alaska and extends north approximately 40 km to its terminus at the head of the Matanuska River valley (Lawson, 1981) (Fig. 2). The proglacial lake that was studied is located less than a kilometer north of the terminus of the glacier (Fig. 3). The lake was approximately 120 m long by 85 m wide and had a maximum depth of 8 m as determined from a topographic survey of the modern lake bed. The lake resided in a topographic low between two recessional moraines. The outer ridges of the moraine that formed the northern boundary of the lake have been dated to approximately 200 to 250 years old using tree ring data (Williams and Ferrians, 1961). However the vegetation on the inner ridges of this moraine group consisted of only willow and balsam poplar seedlings in 1961 (Williams and Ferrians, 1961). The inner ridges would have formed sometime 50 to 100 years ago, constraining the maximum age of the lake to between 90 and 100 years. The lake was in direct contact with the glacier for at least part of its lifetime as evident from aerial photography (Figs. 4 & 6). As seen in the aerial photographs, the extent of the lake during the 1960s appears to be larger than the current lake basin. This study only deals with the deposits and features within the current dry basin (Fig. 5 & 7). The moraine that currently forms the southern boundary of the lake basin is fairly recent, forming less than 40 years ago. Aerial photography shows that the lake drained most of its water some time after 1966 (Fig. 7). The subaerial exposure of the dry lake bottom provided an ideal setting to study the basin morphology and sedimentary characteristics.
Fig. 2: Location map. Adapted from (Lawson, 1979b)

Fig. 3: Oblique aerial photograph of the dry lake taken in July 2003. The lake is approximately 120 m in length. The gray area at the top is the debris covered glacier margin.
Fig. 4: Aerial photograph of the margin of the Matanuska Glacier taken May 31, 1962. The lake studied is at the glacier margin, indicated by the arrow.

Fig. 5: Aerial photograph of the glacier margin taken June 14, 1993. The lake studied is indicated by the arrow. Note the lake is now mostly drained; the dry basin is smaller than the extent of the lake in Fig. 4. The orientations of figures 4 and 5 are slightly different.
Fig. 6: Aerial photograph of the lake studied taken September 5, 1966. The lake is in the center of the photo; it is noticeably larger than the dry basin shown in Fig. 7. Note the locations of stream inlets and outlets as indicated.

Fig. 7: Aerial photograph of the dry lake taken June 14, 1993. The lake studied is almost completely drained at this time and is indicated by the arrow. Figures 7 has a slightly larger scale than figure 6 as they were taken at different altitudes. While it is difficult to know the scale precisely, the dry lake is roughly 120 meters long.
METHODS

Fieldwork at the dry lake was performed during June and July of 2003. Methods of analysis included pit digging, coring, ground-penetrating radar, and topographic surveying. Seven pits were dug into the drained lacustrine deposits; locations of the pits were arranged so that they were spread fairly evenly across the surface of the lake bed and included the important morphological features. The pits were dug until gravel was encountered; this was assumed to be the lacustrine sediment/diamicton boundary. The sections viewed in the pits were documented by detailed digital photography. These photographs were used to examine sedimentological features and to count light and dark laminae. The photographs were taken shortly after the completion of digging each pit, which in retrospect was a mistake. It was found, upon later examination of the cores, that the visibility of laminae varies with the water saturation of the sediment. Since the water saturation varied widely from the top to the bottom of some of the pits, much of the sediment was either too wet or too dry for laminae to be clearly visible (Fig. 8). So what appear to be changes in grain-size or depositional processes in the photographs are only a result of varying water content. Due to this problem in visibility, all counts of laminae from the pits are considered to be minima. A trench was also dug into the lake bed, in order to examine a surficial polygon feature in cross-section.

A Livingston core was taken from each of the two small lake remnants within the basin (Fig. 9). The cores were taken to the depth of a sand or gravel layer that halted the penetration of the coring device; it is fairly certain that the cores reached the base of the lacustrine sediment. Core 1 was a test core and was discarded. Core 2, taken from the southern lake remnant, was extracted in four sections and had a total length of 1.62m.
However, the hole depth of Core 2 was 2.09 m; this inconsistency is due to the loss of some sediment at the ends of each section. Core 3, taken from the northern lake remnant, was completed in one section and was only 0.26 m long.

Eight ground-penetrating radar (GPR) transects were taken across the dry lake bed; the transects were situated to maximize the coverage of the lake, to investigate the major morphological features, and to correlate between the pits (Fig. 9). Ground penetrating radar uses the reflection of electromagnetic waves to image the subsurface at the centimeter to meter scale (Davis and Annan, 1989; Baker and Martinez, 2004 in prep). GPR offers better resolution but less depth of penetration than seismic reflection (Baker and Martinez, 2004 in prep). Electromagnetic waves reflect at changes in dielectric permittivity or magnetic permeability (Van Dam and Schlager, 2000; Baker and Martinez, 2004 in prep). Changes in dielectric permittivity result primarily from changes in water content, which can be linked to grain size and porosity (Van Dam and Schlager, 2000). The minimum vertical resolvability of GPR with a bandwidth of 100 MHz is on the order of 0.5 meters (Davis and Annan, 1989); this means that the GPR system will not be able to distinguish between reflections caused by changes in dielectric permittivity or magnetic permeability that are less than half a meter apart. Figure 10 gives a brief explanation of how to read a GPR profile. For more information on the theory of GPR see publications by (Davis and Annan, 1989; Conyers and Goodman, 1997). This study used a Sensors & Software PulsEkko 100 GPR system. Transects Dry_B, Delta, Dry_Wet, Dry_01, Dry_C and Dry_D were run with 100 MHz antennas, 1 m spacing and 0.25 m step size; transects Dry_B_1, Dry_Uppr, and Dry_Send were run with 200 MHz antennas, 0.5 m spacing and 0.1 m step size (Fig. 9). 100 MHz antennas
generally offer the best compromise between range, resolution and portability (Davis and Annan, 1989). Processing of the GPR data was done using Win Ekko and Ekko Mapper; however, in general GPR data require very little processing. Topographic corrections and a dewow were performed on the profiles and the shading was adjusted.

A topographic survey of the dry lake bed was performed using a Leica Total Station. Along with selected points for the general topography, the location of the seven pits, the two cores and the nine GPR transects were surveyed in. The data collected from this surveying were used to perform a topographic correction on the GPR profiles and to create a topographic map using Golden Software Surfer 7.0.

Fig. 8: This photograph shows three images of the same top section of Core 2 illustrating the change in laminae visibility with varying water content; the top section is the driest and the bottom is the wettest.
Fig. 9: Aerial photograph of the dry lake, showing locations of pits, cores and ground-penetrating radar profiles. The arrows on the ends of the GPR lines indicate the direction in which they were run.
Imagine that you are running GPR antennae across an area that contains three stratigraphic units that have different dielectric or magnetic properties, such as the picture above on the right. The EM wave transmitted will reflect off of each of the interfaces between the units due to the change in properties. The picture on the left shows how the electronic read out of the reflections will appear on the computer screen. Each reflection will consist of a white line in between two black lines. The first reflection is not a true reflection; it is called the direct wave and is caused by a part of the EM wave that travels along the ground directly from the transmitter to the receiver. When considering the depth to reflections, the bottom of the direct wave should be used as the ground surface.

This is an example of what a GPR profile would look like. The black and white lines are the direct wave and reflections as explained above. The gray area at the bottom is indicative that the electromagnetic signal has been attenuated, or weakened and scattered enough so that it no longer has enough strength to make it back to the receiver. Attenuation happens at different depths depending on the wavelength used and the dielectric properties of the material it is travelling through. Weak reflections such as those indicated in the example are unreliable; they may only be a result of waves that bounce around multiple times between layers. (Conyers and Goodman, 1997)
RESULTS

The lacustrine sediment found in the pits and cores is typically horizontally laminated rhythmites composed of fine sand, silt and clay (Figs. 11 & 12). Laminae are usually about 1-2 mm thick, but some are as thick as several centimeters or as thin as fractions of a millimeter. The thickness of sediment varies from pit to pit, as does the number of visible laminae (Fig. 13 and Table 1). Small graded sand beds (Fig. 14), deformed laminae (Fig. 15), loading structures (Fig. 16), and convoluted bedding (Fig. 17) were also found in the pits. Figure 18 shows stratigraphic columns from each of the pits and cores and their locations in the lake. The basin shows little topography but does have an interesting bench that curves around the northern and western sides (Fig. 19).

The GPR transects yielded strong reflectors and penetration of about 3 m (Fig. 20). The majority of reflectors in the basin are fairly flat and horizontally continuous; while some are slightly inclined, onlapping or pinching out to the sides (Fig. 20 A, B & E). Reflectors are vertically spaced by approximately half a meter. In the area where there is a bench-like structure on the surface, there is a strong, slightly parabolic reflector about a half-meter to a meter below the surface that dips steeply to the south and gently to the north (Fig. 20 E, F & H). Below this strong reflector, reflections are weak and discontinuous. In the area where a stream is thought to have entered the lake, reflectors dip steeply lakeward and onlap onto each other (Fig. 20 A). An EM wave velocity of 0.1 m/ns was determined by performing a common midpoint survey (CMP). CMP data are collected by varying the offset between the transmitter and receiver. This velocity is consistent with the published velocity range for silts (0.05 – 0.13 m/ns) (Davis and
Annan, 1989). This wave velocity was used to translate time into elevation on the GPR profiles.

Fig. 11: Photograph of the top 65 cm of Pit C. Rhythmically laminated lacustrine sediment. Scale is in centimeters.

Fig. 12: Photograph of the entire section of lacustrine sediment in Pit A. Scale is in centimeters.
Table 1: Table of the number of rhythmites that were counted in each of the pits and cores. The rhythmites in pits were counted from digital photos, while the rhythmites of the cores were counted from direct observation. Numbers of rhythmites from the pits are minima.

<table>
<thead>
<tr>
<th>Pit</th>
<th>Number of rhythmites counted</th>
</tr>
</thead>
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<tr>
<td>A</td>
<td>293</td>
</tr>
<tr>
<td>B</td>
<td>560</td>
</tr>
<tr>
<td>C</td>
<td>722</td>
</tr>
<tr>
<td>D</td>
<td>101</td>
</tr>
<tr>
<td>E</td>
<td>106</td>
</tr>
<tr>
<td>F</td>
<td>101</td>
</tr>
<tr>
<td>G</td>
<td>147</td>
</tr>
<tr>
<td>Core2</td>
<td>1550</td>
</tr>
<tr>
<td>Core3</td>
<td>215</td>
</tr>
</tbody>
</table>

Fig. 13: Comparison of the thickness of lacustrine sediment and the number of visible dark laminae that were counted for each pit and core. Laminae counts are absolute minima. The two cores are the only ones in which the laminae are visible for the entire section; much of the sediment in the pits was too dry or wet to see the laminae clearly.
Fig 14: Photograph of a normally graded sand to silt bed found in Pit C. Graded sandy beds such as this are evidence of turbidity underflows. Scale is in centimeters.

Fig. 15: Layer containing deformed laminations and gravel from Pit C, most likely representing a large slumping event. This layer is present in Pit B and Core 2 as well. Scale is in centimeters.
Fig. 16: Loading structure from Pit E. Formed by the rapid deposition of sand and gravel on top of less dense, water saturated silt and clay. Scale is centimeters.

Fig. 17: Photograph of convoluted bedding taken from Pit B. Structures such as this indicate rapid deposition by density underflow currents.
Fig. 16: Stratigraphic Columns and Locations of Pits and Cores

Stream Entrance

Location of push moraine ridge

~ 10 m

Legend:
- Laminated fine sand, silt and clay
- Massive diamicton
- Silt with deformed laminations and gravel
- Gravelly sand
Fig. 19: Contour map of the dry lake basin. The area shaded light blue indicates the area covered by the lake when it was filled. The area shaded by the dark blue indicates the current level of the lake. Contour intervals are 0.5 meters.
Fig. 20: Ground Penetrating Radar Profiles

A) Profile Delta (100 MHz)

B) Profile Dry_Wet (100 MHz)

C) Profile Dry_Uppr (200 MHz)

D) Profile Dry_SEnd (200 MHz)
Fig. 20 (A-H): GPR profiles of the dry lake. While the original y-axis of these figures were in time (ns), they have been converted to elevation (meters) using the wave velocity of 0.1 m/ns determined from the CMP in order to facilitate interpretation. For locations of each profile see Fig. 9.
**DISCUSSION**

**Facies**

Examination of the sedimentary characteristics in pits and the reflection characteristics of the GPR profiles reveals three major facies: (1) a deltaic facies, (2) a basinal facies, and (3) a push moraine facies (Fig. 21).

(1) The deltaic facies consists of prograding foresets approximately 2 m thick and interlayered gravelly sand and laminated silt/clay deformed by loading structures. This facies is located in the NE corner of the lake and extends into the lake to the moraine ridge. The delta is also detectable in the surface morphology; the boundary moraines are cut by a channel leading into the lake, and a small (<1m in height) triangular mound discloses where the stream entered the lake. The stream that entered the lake in this area was the primary source of water and sediment input; secondary sources would be direct input from the glacial margin and minor meltwater channels. The greater maturity of the vegetation in the channel compared to that of the lake basin suggests that the stream became extinct before the lake drained. As the glacier receded from the margin of the lake the stream may have been blocked or diverted to a different area. This scenario leaves the lake without significant water or sediment input.

(2) The basinal facies consists of rhythmically laminated very fine sand, silt and clay and covers almost the entire lake basin. Minor constituents of this facies are graded sandy beds and layers with deformed laminae, some including pebbles. Basinal sediment covers all areas of the lake bottom topography, but is concentrated in the deeper areas. Sedimentation in this facies is controlled by combined density underflows and overflow/interflows.
The push moraine facies consists of a subtle topographic ridge composed of diamicton. This facies and its associated structure were formed by re-sedimentation of the underlying diamicton. It was influential in determining the characteristics and distribution of the lacustrine sediment and is an important morphological feature, so it is included in the three major facies in this lake. GPR profiles (Fig. 20 E, F & H) show that the southern side of the ridge dives steeply under the lacustrine sediment. This structure has been interpreted as a push moraine formed by a re-advance of the glacier shortly prior to or during the beginning of lake development. Reasons for this interpretation include the GPR structure and the U-shape of the ridge, which may have mirrored an advancing lobe of the glacier. The shape of the ridge visible in the GPR profiles is similar to that of a ridge that was being formed during the summer of 2003 by the recent advance of the glacier (Fig. 22). Alternatively, the ridge may have been a product of a stagnation of the glacier during the general retreat causing an increased amount of melt-out till and supraglacial debris to be deposited at that location during overall retreat followed by a rapid retreat to the margin of the modern lake basin.

Along with the three major sedimentary facies, two minor facies were identified: an outwash facies and an ice contact facies. The outwash facies consists of a thin interval of coarse sand and gravel located stratigraphically in-between the massive diamicton and the lacustrine sediments. This unit is only about 10 cm thick and is found in pits A and B. During the recession of the glacier, this area would have experienced a brief period of fluvial deposition prior to lake development. The ice contact facies is located along the southern margin where the glacier was in direct contact with the lake and was observed in Pit D. This facies is a variation of the basinal facies; however it contains increased
amounts of sand and gravel, sourced from supra-glacial debris sliding off the glacier margin directly into the lake. The intervals of gravel mixed into the silt may represent small glacial transgressions, bringing supraglacial debris farther into the lake. Alternatively each gravel rich interval may represent the beginning of a melt season when the glacier is right next to the location of Pit D, then later in the season the glacier has melted away from that location and the supraglacial gravel is being deposited elsewhere.

Fig. 21: Schematic diagram of the dry lake showing the general locations of the three major facies and the directions in which density underflows most likely flowed.
Fig. 22: An example of a push moraine that was being formed during the summer of 2003 at the margin of the Matanuska Glacier. The slope of this moraine on its glacier side is similar to that of the ridge in the dry lake. The moraine is about 5m tall.
Ground-Penetrating Radar

One of the goals of this project was to judge the effectiveness of ground-penetrating radar in examining lacustrine sediment and basin morphology. The GPR profiles were helpful in identifying large structures such as deltaic foresets that were not found in any of the pits, as well as interpreting the origins of morphological features such as the push moraine. Figure 23 shows the interpretations of selected GPR profiles. The major reflectors are outlined and the GPR facies are highlighted: the deltaic facies is blue, the basinal facies is red, and the push moraine facies is green. Major reflectors are found at the boundary between the lacustrine sediment and the diamicton and between individual delta foresets. Since the lacustrine laminations are thinner than the vertical resolution of a 100 MHz GPR wave, individual laminations or couplets are not visible. However, in the basinal GPR facies, there are regular horizontal or slightly inclined reflectors at roughly the vertical resolution intervals (~0.5 m), indicating that there are changes in electromagnetic properties at intervals that are less than or equal to 0.5 m. It is probable that the grain-size changes between the individual laminae are producing reflections, and those reflections are coalesced into the reflectors at every 0.5 m. The gradual increase in water saturation with depth was not apparent.
Fig. 23: GPR Facies and Interpreted Reflectors for Selected Profiles

A) Profile Delta (100 MHz)

B) Profile Dry_Wet (100 MHz)
Fig. 23 A-E: Interpretations of selected ground-penetrating radar profiles. The original profiles are placed above or beside the interpretations. Strong reflectors were traced from the originals and the three major GPR facies were highlighted.
**Basin Morphology**

The morphology of the dry lake basin is relatively simple; it is almost ovoid in shape, and only has about 2 m of relief on the lake bed (Fig. 19). Evidence from sediment thickness in pits and GPR profiles indicates that sedimentation has leveled topography, and prior to sedimentation there was about three additional meters of relief in the basin. The most significant topographic feature is the bench-like structure that curves around the northern and western part of the lake bed. This feature has been interpreted as a push moraine (facies 3). This ridge affected the dispersal of sediment into the lake in two ways. First, the lake was divided into two basins that were connected but had differentiated sedimentation: the deeper basin in which sedimentation was dominated by underflows, and the shallower basin, which was shielded from the majority of underflows by the ridge. Second, the delta forsets filled in behind the ridge, preventing coarser material from entering the lake. Turbidity currents would have had to flow over the ridge and down into the deeper basin, reducing their velocity, which may explain the lack of cross bedding and other current structures in the basinal sediments.

**Rhythmites**

The time scale represented by the rhythmites is ambiguous. Each couplet does not represent one year; at absolute minimum counts there are far too many for a lake that only existed for less than 100 years. It is possible that there may be annual varves that contain multiple light/dark couplets; this has been shown to be the case in other proglacial lakes (Shaw et al., 1978; Pickrill and Irwin, 1983; Desloges, 1994). The extra laminae are caused by strong underflows or slump related flows. In the varved deposits of other lakes there is always a well-defined winter clay layer (Gustavson, 1975; Ashley et
al., 1985; Desloges and Gilbert, 1998). However, the winter clay layers are either nonexistent or at least not obvious in the deposits of the dry lake. In an underflow-dominated system, the winter clay layer may have been deposited and subsequently eroded away when the underflows started back up the following melt season. Erosion of the winter clay would leave one year’s rhythmites indistinguishable from the next year’s. Alternatively, much of the clay-sized sediment may have been transported out of the lake through an outlet stream, leaving little material left to form the winter layer.

In Core 2, several rhythmite packets (7-8 cm each) divided by thin clay laminae (Fig. 24) were identified; these clay laminae are few and inconsistent, making annual varves impossible to confidently identify. However, three of the 8 cm thick units contained 110, 95 and 112 rhythmites. If each of these units is interpreted as one melt-season, then each rhythmite could represent one day; ~110 is reasonable for the number of days in the melt season in south-central Alaska (pers. comm. Lawson 2004). Several studies have suggested that rhythmites may represent daily deposits (Johnson, 1997a; Lamoureux, 1999). Stream-water discharge and suspended sediment loads in glacial meltwater streams have been shown to fluctuate diurnally; peak suspended sediment concentrations lagged slightly behind peak daily temperatures, followed shortly thereafter by peak discharges (Church and Gilbert, 1975; Ostrem, 1975; Ashley et al., 1985). Fluctuations in suspended sediment could produce underflows that were active during peak periods of the day then became inactive during non-peak times.

Furthermore, if it is assumed that each year, roughly 8 cm of sediment was deposited in the location of Core 2, then sedimentation would have lasted for about 26 years, given a hole depth of 209 cm. An annual sedimentation rate of 8 cm may even be
an underestimate. The large amounts of supraglacial debris found on the Matanuska Glacier would yield high suspended sediment concentrations in meltwater streams. Weekly sedimentation rates in several small ponds (<10m across) around the terminus of the Matanuska have ranged from >0.1 to 8.1 cm (Lawson, 1979b). If the sedimentation rate was really on the order of 8+ cm/yr, the lake may have been under active sedimentation for less time than previously suspected.

While the thinnest rhythmites may represent diurnal fluctuations, thicker laminae may be the results of late-season rainfall events. A study by Denner et al. (1999) at the Matanuska Glacier demonstrated that a strong storm event late in the melt season caused daily mean discharge at a stream gauging station to exceed the summer daily maximum by 36% and the daily mean suspended-sediment concentration to exceed the summer daily maximum by 430%. This was a very large storm with precipitation totaling 56 mm, but since it was able to increase the suspended sediment concentrations by over 400%, smaller events could still result in strong underflows. Other processes, which probably caused rhythmites, are subaqueous slumping and debris flows entering the lake from either the glacial margin itself or from the sides of surrounding moraines. Slope instability and debris flows caused by buried ice are common along the margin of the Matanuska Glacier (Lawson, 1979b).

Fig. 24: Photograph of the third section of Core 2 with up being to the right. The arrows indicate the tops of potential varves. Each of these units contain 95-115 rhythmites; the laminae are thinnest at the tops and bottoms of the units and thickest in the center. If each of these rhythmites were caused by a diurnal fluctuation in stream discharge or suspended sediment concentration then the units may be varves.
Sedimentation in Similar Lakes

Two small ice-contact lakes at the Burroughs Glacier, Alaska, were studied in the 1980’s and 90’s by Syverson (Syverson, 1998). The Burroughs Glacier lakes were in a similar location relative to the glacier to the Matanuska dry lake and were just slightly larger. A few rhythmites were found and were determined not to be varves, because there were too many for the time span of the lakes. Despite the similarities between the Burroughs Glacier lakes and the Matanuska dry lake, sedimentation was very different. The majority of the deposits were massive silt, weakly laminated silt that draped the lake-floor topography and massive gravelly sand in shoreline areas. The lakes existed for about nine years and accumulated only about 12 cm of offshore sediment. The most apparent reason for the differences is the control of overflows and interflows over the distribution of sedimentation in the Burroughs Glacier lakes, while the Matanuska dry lake was dominated by underflow sedimentation. It is also important to note that one of the lakes was a river-lake. The short-residence time of the water in the river-lake may help explain the lack of varves; the fine suspended matter would be carried out of the lake along with the water, and there would be very little clay left to form the winter layer.

Three small lakes in the Southern Canadian Cordillera were studied in the early 90’s by Desloges (1994). In contrast to the deposits studied by Syverson, Desloges determined that each rhythmite is a varve. In the three Canadian lakes, bottom currents created by large sediment loads distributed sediment during the summer, and suspension settling produced winter sedimentation. The sedimentary processes in these lakes are similar to those of the dry lake; however their interpretations of the rhythmites are very different. These differences are most likely indicative of the greater depth of the
Canadian Lakes; the three lakes had maximum depths of 47 m, 35 m and 30 m, while the
dry lake had a maximum depth of 8 m. There seems to be a balance between surface
area, depth and sediment load necessary for the formation of recognizable varves. It is
possible that a certain depth is needed in a lake to ensure that enough suspended matter is
present in the water column to deposit an identifiable winter lamination.

Lake Drainage

At some time after 1949, most of the water drained through a breach in the
moraines on the eastern side of the lake. Following the draining of the lake, small
channels (<1/2 meter wide) developed on the lake bed surface as snowmelt and runoff
flowed into the lower basin, which still contains a small lake remnant. Aerial
photographs of the dry lake basin revealed polygons ~1 to 2 m in diameter (Figs. 3 & 9).
The polygons are most clearly visible in the area of the lake that has drained since 1993
but are faintly visible in other areas of the lake bed as well. On the lake bed surface these
polygons have a slightly depressed border. A cross-sectional view shows a wedge of
deformed sediment approximately 30 cm wide at the top and 70 cm deep with the
laminations curving down slightly into the wedge (Fig. 25). While the origin of these
polygons is unclear, it is most likely that they were formed by the contraction/expansion
of freezing and thawing while the sediments were fully saturated shortly after the lake
drained. The sediment would have cracked from contraction during freezing, and then
collapsed back into the crack following the spring thaw. Seasonal frost cracks form in
continental climates with wide ground temperature fluctuations where the mean annual
temperature approaches 0 degrees Celsius (Karlstrom, 2001). The mean annual
temperature for nearby Palmer, Alaska between 1949 and 1998 was 1.78 degrees Celsius
Seasonal frost cracks tend to have shallower depths and smaller polygonal patterns than ice or sand wedges; the host material may also be down-turned at wedge boundaries (Karlstrom, 2001). Ice wedges may be ruled out, as they require years of permafrost to grow, and desiccation cracking is unlikely because the area is still fairly wet. Smaller scale desiccation cracks and the footprints of moose have also deformed the lake bed surface.

Fig. 25: Cross-sectional view of one of the polygons that has appeared on the lake bed surface since lake drainage. Scale is in cm.
Lake History

From the data gathered through observations of the sedimentary deposits, GPR profiles, aerial photos, basin morphology and surficial features, it is possible to reconstruct an interpretation of the history of the small proglacial lake (Fig. 26). The Matanuska Glacier has remained relatively in the same location with some horizontal retreat while thinning considerably over the past 100 to 150 years (Williams and Ferrians, 1961). Following the deposition of a recessional moraine 75 to 100 years ago, the glacier retreated 50 to 100 meters, uncovering the area of the future lake. The area experienced a brief period of fluvial deposition. A small advance by the glacier formed a push moraine. The glacier then retreated and a lake developed at its margin. While the glacier was located at the lake margin, a meltwater stream delivered sediment laden water to the lake. Density underflows and suspension settling distributed and deposited sediment throughout the lake. This sedimentation probably lasted for 20 to 30 years. Sedimentation may have ceased after a glacial retreat causing a shift in meltwater drainage, depriving the lake of water and sediment input. Alternatively, a small glacial advance, which would have created the moraine that is now the southern boundary of the lake basin, may have blocked the meltwater stream. The lake remained for some time without input before draining at some point between 1966 and 1993. Post-draining of the lake, freezing and thawing action created surficial polygons and deformed wedges of sediment, runoff eroded channels into the lake bed, and desiccation cracking and moose prints deformed the surface.
1) The glacier forms a recessional moraine.

2) The glacier recedes from the moraine creating a topographic low in between itself and the moraine. Minor melt water streams develop.

3) A minor advance by the glacier creates a small moraine ridge.

4) A small lake forms in the topographic low. The lake is in direct contact with the glacier margin and is supplied with glacial melt water and sediment by a stream entering on the east side. A stream breaches the recessional moraine creating an overflow outlet from the lake. Density underflows concentrate sedimentation in the deepest areas; some interflow/overflows distribute sediment throughout the lake.

5) As the glacier recedes from the lake margin, melt water drainage channels shift so that the lake no longer receives any major water or sediment input.

6) The lake drains through a small channel cut through the moraines on the left. Post-drainage, small down-cutting channels and polygonal freezing features appear on the lake bed.
Conclusions

(1) Sedimentation in this small proglacial lake was dominated by underflow processes but was influenced by overflows and interflows as well. Suspended sediment concentrations of the melt water input were the most important factor influencing the amount and characteristics of sediment deposited.

(2) Three major facies were identified: a deltaic facies, a basinal facies, and a push moraine facies. Two additional minor facies, the fluvial and ice-contact facies, played limited roles in sedimentation.

(3) Individual rhythmites are not varves. Varves containing multiple rhythmites may have been deposited but cannot be confidently identified. Reasons for the lack of identifiable varves are erosion of the winter clay by underflows or inadequate depth to build up enough suspended sediment in the water column to deposit a winter clay lamination. Individual rhythmites may represent diurnal fluctuations in suspended sediment concentrations and discharge and/or underflows caused by slumping or meteorological events.

(4) Ground-penetrating radar is an effective tool for determining subsurface basin morphology, examining large sedimentary structures, and evaluating the horizontal continuity of sedimentary facies or structures. Vertical resolution of the radar waves limits the analysis of small sedimentary structures such as lacustrine laminae. Future GPR studies on lacustrine sediments should experiment with higher radar frequencies.

(5) Post-drainage processes such as expansion/contraction due to freezing, fluvial erosion, desiccation and bioturbation threaten the preservation small proglacial lake deposits.
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