REGIONAL DESCRIPTION

Glacier Bay National Park and Preserve is a rugged and sharply dissected mountainous area that contains a few gently contoured lowlands. It includes many deep fjords (over 500 m in places), and some of these have glaciers that reach tidewater. A recent land acquisition on the northwest has increased the size of the Park to 10,700 sq. km. It is up to 170 km in length and 125 km wide. The main body of water, Glacier Bay, is 120 km long and 15 km across at its widest point. The Bay trends southeast to northwest along the structural grain of the underlying bedrock.

Physiography and Vegetation

Warhaftig (1965) has described the physiographic divisions of Alaska, and he places the Monument within the Pacific Border Ranges province. He further subdivides this province into the Gulf of Alaska Coastal Section, the St. Elias Mountains (represented in the Park by the Fairweather Range), and the Chilkat-Baranof Mountains. Warhaftig's subdivision of the Park has been summarized by McKenzie (1968, p. 15-18).

A more local and perhaps a more pertinent description based on division of the Park into major drainage basins is given by Streveler and Paige (1971, p. 4-9). These authors recognize five provinces; four are discrete drainages, and the fifth is a region of many small contiguous ones (Figure 2).

Excursion Inlet. All of the area between the eastern boundary of the Park and the divide between the Bartlett-Salmon River drainages are included in this province. It is a straight U-shaped valley, the lower reach of which is partially below sea level. The highest peaks reach about 1,500 m elevation, and the lowest points are 150 m below sea level.

The vegetation cover is composed of "mature" spruce-hemlock forest that extends from sea level upwards to about 750 m. Above this is a zone of brush and then alpine meadow. Vegetation in low-lying areas has probably not been disturbed by glacier ice since the end of the Pleistocene.

Glacier Bay. This province is bounded on the north by the Chilkat and Takinsha Ranges, which reach a maximum elevation of about 2,100 m. To the northwest, the boundary is formed by the crest of the Fairweather Range. Maximum elevations along this portion of the crest reach 4,160 m. A series of discontinuous peaks less than 1,500 m high form the western boundary. The area defined consists
of a number of U-shaped valleys and fjords that merge to form the main Glacier Bay. Much of this region was occupied by ice (up to 1,200 m thick) during the Neoglacial advance. Parts of the province are still covered by glaciers, although rapid retreat has uncovered much land and sea. Some stands of "mature" forest survived on hillsides between the permanent snowline and the glacier surface. Newly uncovered land surfaces are in various stages of revegetation.

Water depths in the main fjord average 500 m in the upper reaches and decreases to 45 m near its mouth and out into Icy Strait. The Gustavus region was only partly covered by Neoglacial ice, but was extensively covered by outwash gravels, and was partly covered by marine waters.

Dundas Bay. This province is composed of the Dundas River Valley, a broad alluvial valley; the Dundas Bay fjord system; and parts of Icy Strait. Elevations range from 1,140 m above to 275 m below sea level. Spruce-hemlock-cedar forests intermingle with lategene pine dominated bogs. Timberline averages about 600 m. The floor of the river valley has a spruce-cottonwood forest, typical of a less stable floodplain environment.

Little ice or permanent snow-cover exists within this province, but large amounts of meltwater were shed into the Dundas River Valley through passes from Geikie Inlet and Berg Bay during the Neoglacial.

Brady Glacier. Much of this province is covered by the Brady Glacier. Ice locally dams adjacent valleys, ponding lakes and filling them with outwash. This ice is supplied from a basin east of the La Perouse-Crillon-Bertha segment of the Fairweather crest. Low mountains (less than 1,500 m) bound the eastern side of the province, and these do not contribute much snow to the glacier. The broad, shallow Taylor Bay is southeast of the glacier terminus.

The vegetation is similar to that near Dundas Bay, except that large lowland areas are covered by nearly barren active outwash.

Outer Coast. The physiographic features of this province are found in parallel northwest-southeast trending belts. There are from west to east, a 5 km wide band of shallow sea less than 170 m deep, a beach that is depositional in character, a series of raised beaches the highest of which is over 500 m above sea level and is superimposed on a ridge 1,500 m in elevation. Behind the ridge is a fault-controlled valley running the length of the province. The Fairweather crest rises from the valley to a maximum elevation of 4,665 m at Mt. Fairweather.

Treeline is at about 600 m. Below it, spruce-hemlock-cedar forest and pine-covered bogs are separated by scrub forests. Above it, brushland is found. Alpine meadows occupy the areas between the brush and the snowline.
Climate

Most portions of the Park that are near sea level fall into Koppen's Cfc climate type (Loewe, 1966). This is a warm, temperate, rainy climate with a cool, short summer. Loewe states that parts of the Monument might belong to type Cfc'c with a relatively dry summer, compared to winter precipitation, but still with a generally wet climate. He adds that some areas might also have type Dfs'c, a rainy climate with a cold winter and a cool summer.

Maritime influences dominate the climate of the Park. Annual and diurnal temperature variations are small, relative humidity and cloudiness are high, precipitation is heavy, and winds can be strong. Data from stations near the Park indicate that a rather steep gradient in both precipitation and temperature exists from coastal areas inland, but this is difficult to document within the Park itself owing to a lack of data. Wahrhaftig (1965) maps these gradients, and older local data have been discussed by Loewe (1966), Bengtson (1962), Haselton (1966) McKenzie (1968), Streveler and Paige (1971), Mackevett and others, (1971), and Derksen (1976). The data listed in Tables 1 and 2 are similar in content to those discussed by the above authors, but they are more recent. These are for four stations in or close to the Park and they were extracted from U.S. Climatological Data, published by the U.S. Department of Commerce.

The mean monthly temperatures (°F) and mean monthly precipitation (in.) in Table 1 are averaged for the period 1970-1978. Cape Spencer is on the outer Coast, Haines Terminal is near the head of Lynn Canal, and Linger Longer is north of the Park (59°26'N, 136°17'W) at about 200 m elevation. The Cape Spencer station was moved to Elfin Cove in 1974, but the new station is also located near the open Pacific Ocean. Measurements for Glacier Bay are from Bartlett Cove. It can be seen that mean temperatures are more moderate and precipitation greater near the coast. Table 2 shows the extremes of temperature for each month of 1972. In general, the further inland one goes, the greater is the range in extremes of temperature.

This effect could undoubtedly be seen within the Park if investigations were made, but local topographic effects in the fjords would probably render the overall pattern complex. Streveler and Paige (1971) do note that temperatures at sea level in the upper reaches of the Bay are cooler in the summer, both during the day and night, than they are at Bartlett Cove near the mouth of the Bay. Loewe (1966) suggests the same relationship for winter temperatures.

McKenzie (1968) states that the position and intensity of low pressure areas in the Gulf of Alaska control the weather in southeast Alaska. The winds generated by these lows cause prevailing winds to be from the south and southeast. They come in from the North Pacific and carry abundant moisture and precipitation to this
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From U.S. Climatological Data, the U.S. Dept. of Commerce.
TABLE 2

Monthly extremes in temperature (°F) and total monthly precipitation (inches) for four stations in southeast Alaska in 1972.

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part of Alaska. During late May and early June, the position of the Aleutian Low shifts to the north and west. This causes May to be the month with the most clear or partly cloudy days. May has an average of 12 such days (Streveler and Paige, 1971).

Streveler and Paige (1971) indicate that summer days often have overcast skies, mist, rain, and cool temperatures. In the fall, the Pacific High becomes important in weather patterns in southeast Alaska. As the High moves north in late summer, it strengthens the westerly onshore flow of air and causes September, October, and November to be the months of maximum precipitation.

During the winter, there are periods when southeast winds bring rain or snow and moderate temperatures. Occasionally, cold air masses from the north bring clear, harsh winter conditions. Snowfall is said to average about 4 m/yr, but owing to winter rains, snow rarely accumulates to more than 1.5 m at sites near sea level (Streveler and Paige, 1971).

Bedrock Geology

Southeast Alaska is composed of nine separate fault-bounded tectonostratigraphic terranes and several of these occur only in this region (Berg, 1979; Berg and others, 1978). The terranes trend north-northwest, and each is characterized by a distinct stratigraphic sequence and structural history. According to Berg (1979, p. B116), the gross differences between adjacent terranes "imply juxtaposition by large-scale tectonic transport. This juxtaposition produced a mosaic of discrete tectonic elements that record a long and complex history of amalgamation and accretion to the continental margin of North America."

The earliest rocks known to overlap any two terranes are Permian in age; thus, amalgamation had begun by Permian time. Rocks of Late Cretaceous age are the youngest to be regionally penetratively deformed. Apparently, the major episode of accretion occurred during or just prior to this time. Redistribution of terrane along major fault zones, Cenozoic intrusion and thermal metamorphism, and local deposition of continental volcanic and sedimentary rocks have occurred since amalgamation of the terranes (Berg, 1979).

Three of the nine terranes are known or suspected to be present within Glacier Bay National Park. From east to west, they are the Alexander terrane (Craig subterranne of Berg and others, 1978), Wrangellia (or Wrangell terrane), and the Chugach terrane. In addition, two more terranes, the Gravina-Nutzotin Belt and the Taku terrane, are present a few tens of kilometers to the east of the Park (Figure 3).
Figure 3. Tectono-stratigraphic terranes near the study area, after Berg, 1979.
Rocks of the Alexander terrane comprise most of the bedrock in the Park and all of the bedrock in the study area. The base of the section is not exposed in the region, but over 7,000 m of Middle Silurian to Upper Devonian limestone and argillite have been measured (Rossman, 1963; Seitz, 1959; Loney and others, 1975). Rocks to the north of this middle Paleozoic section (and assumed to be of similar age) contain far more volcanics (MacKevett and others, 1971). The terrane is interpreted to represent part of a lower and middle Paleozoic magmatic arc sequence (Berg and others, 1972).

Intrusive igneous rocks of Mesozoic and perhaps Tertiary age dominate the northern and western parts of the Alexander terrane in the Park. Most of these rocks are mesozonal foliated granitic bodies of diorite, and granodiorite. Pervasive metamorphism associated with the intrusives render stratigraphic interpretation of the sedimentary section difficult.

Wrangellia is a name derived from the Wrangell Mountains, and was given to an apparently allochthonous block of subcontinental dimensions by Jones and others, (1977). Brew and Morrel (1979a) suggest that a zone of structurally and stratigraphically complex geology 5-12 km wide and 100 km long extending from Tarr Inlet to Taylor Bay belongs to this terrane. The sequence consists of phyllite, slate, conglomerate, chert, greenstone, greenschist, and marble that are interpreted to be Permian and/or Triassic in age. Jurassic and Cretaceous granitic rocks invade both the Alexander and Wrangell terranes, suggesting that perhaps by Jurassic, and at least by middle Cretaceous time, the two terranes had been joined. The plutons do not intrude rocks of the Fairweather region; however, middle Tertiary granitic rocks occur in the Wrangell terrane and bordering terranes, and suggest that within the Park the western contact of the Wrangell terrane is post-Late Cretaceous and pre-middle Tertiary in age (Brew and Morrel, 1979a).

Rocks of the Fairweather Range and coastal area exposed west of the Border Ranges Fault comprise the Chugach terrane. The stratigraphic sequence is similar to and continuous with a sequence of graywackes, schists, and amphibolites north of the Park that are known to be Cretaceous in age. Similar rocks of Jurassic and Cretaceous age also occur to the south (Rossman, 1963; Brew and Morrel, 1979b). These rocks are interpreted to be a metamorphosed turbidite sequence that probably spans all of the Cretaceous.

Thick bodies of layered gabbro make up the high peaks of the Fairweather Range. They are underlain by the Cretaceous sequence mentioned above. Low strontium 87/86 ratios (Brew and others, 1977), internal layering, and stratigraphic position of the gabbros support the idea that they represent slices of oceanic crust tectonically implaced in the metamorphosed sequence as it was accreted to the continent during subduction. Regional mapping and internal structures suggest an intrusive origin (Brew, 1983, per comm.)
West of the Fairweather fault, nearly 4,000 m of Oligocene to Pliocene marine (including marine tillite of the Yakataga Fm.) and non-marine clastic and volcanic units unconformably overlie the Mesozoic basement (Miller, 1961; Plafker, 1967; Mackevett and others, 1971). These rocks are generally simply folded and often steeply dipping. Several north-northwest trending faults cut the folds.

Dominant north-northwest trending fault zones and subsidiary east-west trending zones within all terranes appear to control the position of the fjords of the Park (Gwenhynfel and Sainsbury, 1958).

Economic Geology

A number of mineral surveys have been carried out in the area occupied by Glacier Bay National Park (see Rossman, 1959 and Mackevett and others, 1971 for extensive summaries). Metallic mineral deposits are the most important geological economic resources of the Park. Brew and others (1979) provided locations of six important mineral deposits with identified resources. Nine general areas that have undiscovered speculative resources were also identified by these authors. Nickel, molybdenum, copper, zinc, and gold are the elements of greatest economic interest.

Areas marked (A) in Figure 4 indicate Pacific beach sands that have produced placer gold. Area (B), the Crillon-La Perouse area, includes the Brady Glacier magmatic nickel-copper deposit (1). The Mount Fairweather area (C) is inferred to have an ore environment and resources similar to area (B). Area (D) includes the Margerie Glacier porphyry-copper deposit (2) and the Orange Point volcanogenic sulfide deposit (3). Copper, silver, gold, and zinc are the important elements present. The Reid Inlet area (E) has produced gold from vein deposits (4), and the Rendu Glacier area (F) includes the "massive chalcopyrite" skarn deposit (5) with tungsten, copper, silver, and gold. Area (G) contains the Nunatak porphyry-molybdenum deposit (6) which has molybdenum and copper. Area (H) in the Takhinska Mountains contains speculative molybdenum and copper resources and also speculative copper-zinc resources in a volcanogenic environment.

A similar volcanogenic environment, the White Glacier area (I), contains unquantified speculative zinc and copper resources (Brew and others, 1979).

Glacial Geology

The glacial stratigraphy of the Muir Inlet region is the best described and the best known in the Park. All of the deposits are Wisconsin or Holocene in age. Goldthwait (1963; 1966) has
Figure 4. Economic mineral deposits (numbered triangles) and areas of known or suspected economic potential (shaded areas with letters) in Glacier Bay National Park, after Brew and others, 1979. See text for further explanation.
given a general discussion of the history represented by them, and
detailed field studies have been completed for Muir Inlet
(Haselton, 1966; 1967), Adams Inlet (McKenzie, 1968; 1970), and
Wachusett Inlet (Mickelson, 1971). A composite stratigraphy
(Figure 5) for this part of the Park is outlined below. Much of
the nomenclature derives from the work of Haselton and McKenzie.

The oldest glacial unit recognized is a compact dark gray,
loam till that contains abundant pebbles and boulders. It is
usually found on bedrock where it is exposed. McKenzie (1968)
proposed the name Granite Canyon Till for this unit. It is older
than 11,000 yrs B.P. The furthest advance of Late Wisconsin ice in
Glacier Bay probably occurred about 15,000 yrs B.P. (McKenzie and
Goldthwait, 1971). The Muir Formation (Haselton, 1966), a till in
Muir Inlet, is probably correlative to the Granite Canyon Till,
although pebble counts indicate that the two were derived from
separate flow systems (McKenzie, 1968).

Marine invasion of low-lying areas took place upon retreat of
Wisconsin ice, and the resulting marine deposits are named the
Forest Creek Formation (Haselton, 1966). These are blue-gray
laminated fossiliferous silts and clays.

The deposits are generally one to a few meters thick, and
contain shelly remains of shallow water pelecypods and gastropods.
McKenzie and Goldthwait (1971) suggest an age of 11,000 yrs B.P.
for this unit. It occurs from 3 to 59 m above sea level in the
Muir Inlet region. Dersken (1976) has found similar deposits west
of Taylor Bay. They also occur east of Gustavus (Brew, in McKenzie
and Goldthwait, 1971), and Post (unpublished data) has dated shells
from a deposit at Ptarmigan Point in the West Arm at about 13,000
yrs B.P. Wisconsin ice probably had withdrawn from the fjords of
the Park by 13,000 yrs B.P., and certainly by 11,000 yrs B.P.

Haselton (1966) postulated a Late Wisconsin readvance of ice
in Muir Inlet, based on the occurrence of a diamicton above the
Forest Creek Formation at the Forest Creek locality. McKenzie
(1970) has argued that evidence from one site is not enough to
prove that a readvance took place. He suggested that the strati-
graphic position of the Muir Formation at this site and the leach-
ing of carbonates from the till indicate weathering and mass move-
ment of nearby pre-Forest Creek till down onto the marine sedi-
ments.

Goldthwait (1963; 1966) suggests that the fjords of the Muir
Inlet region began to fill with outwash after the retreat of Wis-
consin ice. Minimal terminal positions were likely reached by the
glaciers between 7,000 and 4,000 yrs B.P. (the so-called Hypsither-
mal interval) when temperatures were probably slightly warmer than
at present.

The rate of deposition of Hypsithermal gravels has already
been discussed.
Figure 5. Generalized glacial stratigraphic section for the Muir Inlet region, see text for explanation.
Haselton (1966) termed these gravels the lower member of the Van Horn Formation. They are generally poorly sorted and poorly bedded yellow-brown gravels from a few to nearly 60 m in thickness. They built to present sea level about 7,000 yrs B.P. (McKenzie and Goldthwait, 1971). Isolated stumps and forest beds occur within these gravels and have been used both to date them and to define the composition of the Hypsithermal forests (c.f. Goldthwait, 1963; 1966; Cooper, 1937).

Above the oxidized gravels of the lower member are finer-grained sands, silts, and clays that are assigned to the middle member of the Van Horn Formation (Haselton, 1966). They are interpreted as having a lacustrine origin, and they are usually one to several meters thick. As previously mentioned, it has been suggested that lacustrine deposition began sometime between 4,400 and 3,400 yrs B.P. and continued to as late as 1,700 yrs B.P. The similar dates bracketing lake sediments at widely separated localities led Goldthwait (1963) to suggest that they had formed in one large ice-dammed lake when ice from the West Arm blocked Muir Inlet. The isolated occurrences and sometimes varied stratigraphic position of the sediments have suggested to some that they formed in lakes ponded locally by outwash fans (for example Haselton, 1966).

Within Muir Inlet, 3 to 75 m of fairly well-bedded, poorly-sorted, light-colored gravels overlie the lacustrine member. This is the upper member of the Van Horn Formation (Haselton, 1966). Haselton (1966) and Goldthwait (1966) both suggest that these gravels were deposited as ice advanced into the region, and outwash was spread in front of the glaciers. Haselton (1966) indicates that this upper outwash was building into Muir Inlet from about 2,300 yrs B.P. to 1,700 yrs B.P.

The stratigraphy in Wachusett Inlet is similar to that in Muir Inlet (Mickelson, 1971); but in Adams Inlet, McKenzie (1968) did not recognize the middle or upper members of the Van Horn Formation. The oldest dates on lower Van Horn gravels in Adams Inlet are about 3,800 yrs B.P. McKenzie (1968) suggests that the older parts of the lower member are below sea level. The oxidized gravels here are overlain by thick (up to 66 m) lacustrine deposits of the Adams Formation which are, in part, time equivalent to the upper Van Horn gravels of Muir Inlet. Lake damming began after 1,700 yrs B.P. McKenzie (1970) also indicated that the lake was probably initially dammed by outwash in Muir Inlet, although the source of the ice responsible for the outwash was not known to him. He assigned the outwash, deltaic and fluvial sediments stratigraphically above the Adams Formation to the Berg Formation. This unit is up to 100 m thick. Deposition is said to have begun about 840 yrs B.P.

All Neoglacial till and ice-contact deposits are assigned to the Glacier Bay Formation (Haselton, 1968; McKenzie, 1968). The till is a gray bouldery to pebbly sand loam as much as 30 m thick,
that shows little oxidation or leaching. It is, of course, diachronous and began to be deposited in Reid Inlet as long ago as 4,700 yrs B.P. (Goldthwait, 1966). In Wachusett Inlet, till began to form at 2,700 yrs B.P., and ice overrode a tree high on White Thunder Ridge at 2,100 yrs B.P. Ice from Casement Glacier reached within 2 km of Muir Inlet at 1,440 yrs B.P. (Haselton, 1966), and parts of Adams Inlet were glaciated as recently as 200-300 years ago (McKenzie and Goldthwait, 1971).

Outwash deposits in Adams Inlet that are still forming have been assigned to the Seal River Formation (Goldthwait, 1966). In some low areas, up to 30 percent of the land surface is covered by these deposits (McKenzie and Goldthwait, 1971).

The glacial geology of much of the rest of the Park is less well known. Derksen (1976) and Bengtson (1962) worked on the chronology of Brady Glacier. Neither formally named glacial units, although Derksen employed a number of radiocarbon dates to construct a detailed history of Holocene ice advance, which will be discussed later. Deposits in the inlets of the southwestern part of Glacier Bay, in the Beartrack Valley, and in the West Arm of the Bay are known only from reconnaissance studies (Goldthwait, 1963; 1966).
SEDIMENTARY SETTING

Sediments in Muir and Wachusett Inlets that have been assigned to the middle member of the Van Horn Formation (Haselton, 1966; Mickelson, 1971) generally overlie coarse oxidized gravels of the lower member of the Formation. A very sharp contact usually separates the two. The fine-grained clastic sediments of the middle member are interpreted as lacustrine in origin (Haselton, 1966), and they often contain primary sedimentary structures that suggest rapid deposition, often from suspension. Load casts, flame structures, graded beds, and climbing ripples were observed in many outcrops. Dropstones were observed in all exposures, and it is believed that all of the sediments were deposited in lakes formed behind ice-dams, or into which bergs were discharged.

Detailed observations of the finest-grained sediments were made in order to attempt to correlate the deposits from one outcrop to another. A number of problems complicated this endeavor. All of the sediments were overridden by glacier ice, and evidence of former shorelines has been eroded. In addition, exposures of lacustrine sediments are ephemeral, and as long ago as 1923, Cooper recognized the need to study the deposits before mass wasting and plant colonization covered them from view. In the course of this study, exposures were seen to change in character on a year-to-year basis as slumping and debris and/or mud flows removed material from outcrops or partially buried exposures. The availability of adequate exposures greatly influenced the distribution of study sites.

Besides the bias introduced by the availability of outcrops, an additional and related bias was introduced by glacial erosion of most of the sediment that had been deposited in Muir Inlet and its tributaries prior to Neoglacial ice advance. Lacustrine exposures are usually found at localities near bedrock knobs in areas that were protected from erosion by glacier ice. High energy mountain streams have incised the unconsolidated sediments at many places, and lacustrine deposits have been exposed at a few sites. Rhythmically bedded sediments consisting of silt-clay couplets comprise the most distinctive lacustrine sediments. At most sites (Figure 6), couplets usually have thick silt layers much greater in thickness than their overlying clay layer. Such a configuration has been cited by Ashley (1975) as evidence of deposition near the source of sediment input to a lake. Thus, most exposures studied were probably deposited near the mouths of streams, perhaps near deltas that prograded into the Neoglacial lakes. Distal sediments with thin silts and relatively thicker clays were rare in the Muir Inlet region. They were most likely deposited in the middle of the lakes (middle of the present fjords) or at least away from inflowing streams, and most distal sediments were removed by overriding Neoglacial ice.
Figure 6. Map showing sites where "varve" chronologies were obtained.
All of the lacustrine sediments of the Muir Inlet region are similar to one another in mineralogy and pollen content, as described in other chapters of this report. Sediments exposed in Adams Inlet (Adams Formation) are generally younger than the middle Van Horn Formation of Muir and Wachusett Inlets, but they are very similar in physical appearance, mineralogy, and pollen content.

Lacustrine sediments have been reported as far north in Muir Inlet as the McBride Glacier (Powell, pers. comm.). They are exposed as far south as Point George on the east side of the inlet. South of Point George on the west side of the Inlet, the lacustrine member of the Van Horn Fm. pinches out. The sediments that are present there are sheared and thrust to the north as if they had been overridden by northward moving ice. In addition, the upper Van Horn gravels appear to slope to the north. These factors suggest an ice dam and a sediment source across the mouth of Muir Inlet. Such a damming event has also been suggested by Ovenshine (1967) on the basis of the distribution of gabbroic boulders in Glacier Bay that were derived from the Fairweather Range.

Methods

Records of the thicknesses of successive silt-clay couplets were obtained using a method modified from that of De Geer (1912) and Antevs (1925). These authors recorded thicknesses directly from the outcrop onto paper strips which were later measured in the laboratory. The great amount of precipitation in Glacier Bay precludes the use of paper as it swells when wet, and will generally deteriorate under severe conditions. Instead of paper, long strips of plastic-fiber banding-tape were used for recording purposes. This material was 15 mm wide by 0.5 mm in thickness. The fibers were stiff and had been fused, making the tape non-stretching.

Outcrops that were sampled were cleared with an entrenching tool, and smoothed with a steel brick-trowel. Smoothing allowed observation of the fine details of sedimentary structures. Once an outcrop was cleared, a strip of plastic-tape of appropriate length was fastened to the top of the outcrop. This was accomplished by punching a hole in the top of the tape with an ordinary paperpunch, and then securing the tape in place by pushing a 16-penny spike through the hole and into the outcrop. Figure 7 shows two tapes in place on an outcrop.

A razor-blade was used to make a slit in the edge of the tape where a clay layer was in contact with an overlying silt layer. An indelible felt-tip marking pen was used to label the tapes and to record the angle of the outcrop from the horizontal (to calculate true thickness from the apparent thicknesses actually measured on the tapes). Other pertinent notes such as the position of distinctive sedimentary structures were also recorded directly on the tapes.
These records were measured in the laboratory, and plots of couplet thicknesses spaced uniformly along one axis (Figure 8) were constructed for 13 outcrops (Appendix A). The plots were then compared to one another visually following the method of De Geer (1912). It was hoped that by finding matching records of thick and thin couplets, it would be possible to correlate sediments from one outcrop to another.

In addition to the records of couplet thicknesses, detailed records of the stratigraphy of the entire fine-grained sequence, including sand bodies, were recorded at 11 sites (Appendix B) (most of the sites also yielded records of couplet thicknesses). These provide some insight into the nature of Neoglaciation lacustrine sedimentation in the Muir Inlet region.
Figure 8. Plot of the thicknesses of silt-clay couplets for site W5 and the lower part of the rhythmically bedded portion of site W1. Shading indicates segments where correlation between the two records is good.
Grain-size analyses were completed for some of the samples obtained for paleomagnetic analyses. Samples consisting of 10-15 gm (air-dried weight) were wet-sieved on a 40-sized mesh. They had been disaggregated and dispersed overnight in Calgon solution (sodium hexametaphosphate) prior to sieving. The less than 40 fraction was washed into a 1,000 ml graduated cylinder with distilled water, and the volume was then brought up to 1,000 ml. The final concentration of dispersant was 0.5 gm/liter of suspension. Analyses were carried out by the pipette method (Galehouse, 1971) using a pipette stand constructed by R.D. Powell.

A summary of the results of attempts to correlate deposits using thickness records, of results of grain-size analyses, and a description of stratification sequences in the Muir Inlet region will follow a brief review of sedimentation in glacial lakes.

Mechanisms of Glaciolacustrine Sedimentation

Interest in the nature of glaciolacustrine sedimentation was stimulated by the work of De Geer (1912) and Antevs (1925; 1951). These authors considered the distinctive silt-clay couplets characteristic of glaciolacustrine sediments to be annual in origin. They were able to suggest detailed chronologies for Late-Pleistocene ice retreat in Scandinavia and eastern North America prior to the advent of radiocarbon dating. These chronologies were based entirely on records of thicknesses of successive sediment couplets.

De Geer (1912) named the couplets "varves", and the word implies an annual origin for these features. Ashley (1975) pointed out that the annual connotation implied by the term "varve" cannot always be proven, and she used the term "rhythmite" interchangeable with "varve". Rhythmite is a descriptive term that denotes repetition of alternating sediment layers or laminae of different textures or structures. It implies nothing about the length of time required for deposition. Because the present author failed to prove conclusively that the silt-clay couplets in the Muir Inlet region are annual, the term rhythmite is favored in this report.

De Geer (1912) suggested that the silt layers in his varves (usually graded or containing multiple graded beds) were formed as the result of deposition from sediment-laden underflows that occurred during the summer melt-season. The clay layers were said to represent the settling of the finest particles from suspension during the winter when sediment input to a glacially influenced lake was small. This idea had been proposed earlier for glaciolacustrine sediments in New England (Emerson, 1898). However, Antevs (1925; 1951) and Johnston (1922) suggested that because water near 0°C is less dense than water near 4°C, overflows of cold meltwater out over warmer pro-glacial lake water should occur.

According to their ideas, predominantly silt-sized material rained from upper lake waters during the summer melt-season when stream competence was great. Clay deposition occurred in the
winter, as in the De Geer model, when stream flow slackened. This theory fails to take into account the density of the sediment with which inflowing water is charged.

Most workers now subscribe to the ideas of De Geer and Emerson. Using theoretical and physical models, Kuenen (1951) concluded that graded beds such as those found in varves were probably the result of deposition from turbidity currents. Stratigraphic and hydrodynamic studies (Latjai, 1967; Jopling and Walker, 1968; Agterberg and Banerjee, 1969; Banerjee, 1973; Stanley, 1974; Ashley, 1975; Harrison, 1975; Shaw, 1975; Shaw, 1977; Shaw and Archer, 1978) have emphasized the role of gravity-induced density underflows in glaciolacustrine deposition.

Physical limnologic studies of river-dominated lakes, including some lakes influenced by glacier-derived runoff (Mathews, 1956; Houbolt and Jonker, 1968; Fulton and Pullen, 1969; Gustavson, 1975; Gilbert, 1975; Sturm and Matter, 1978; Hamblin and Carmack, 1978; Pharo and Carmack, 1979; Killworth and Carmack, 1979; Carmack and others, 1979), suggest that density underflows are common in such lakes.

The important factor that determines whether inflowing water will enter a lake as an underflow, an interflow (flowing into the body of the lake as a spreading plume above the lake bottom and below its surface), or as a surface overflow is the density contrast between the two bodies of water.

Gustavson (1975) has described the physical limnology of Malaspina Lake, a large proglacial lake in southeastern Alaska. During the summer when observations were made, the lake was found to be density stratified with respect to suspended sediment content, but not with respect to temperature. Surface water was nearly 4°C, and water at the lake bottom was as cold as 0.3°C.

Inflowing streams highly charged with sediment entered Malaspina Lake as underflows (continuous turbidity currents). Streams entered as interflows if their suspended load fell between the maximum and minimum suspended load values for lake water, and overflows occurred where inflowing water had a suspended load less than that of the lake water. Gustavson observed all three types of flows in Malaspina Lake, but attributed the graded (normal and reverse) and rippled coarser-grained parts of "varves" that he observed in sediment cores from the lake bottom, as the result of deposition from density underflows. Clay layers were said to be deposited by the settling of fine particles from suspension throughout the winter (Gustavson, 1975, p. 261). Similar conclusions have been arrived at by other authors (Gilbert, 1975; Ashley, 1975); although Pharo and Carmack (1979) suggest that graded beds can also form as the result of deposition from interflows.
Hamblin and Carmack (1978) observed and modeled the passage of a strong river into a stratified fjord-lake (Kamloops Lake, British Columbia). River-induced currents influence and may dominate circulation patterns within such lakes. These authors concluded that predictions of stream behavior based on jet-theory (Batemen, 1953; Wright, 1977) are inadequate for stratified lakes. Specifically, the model of a round, turbulent jet entering a homogeneous body of water predicts too much entrainment of still water by the jet. Hamblin and Carmack modeled a "quasi-three dimensional, turbulent plume flowing into a rotating stratified body." Dense sediment-charged river water enters a lake and behaves like a sinking plume. Water turbidity decays as lake water is entrained (the rate of decay is proportional to the sediment concentration or turbidity), and if turbidity decreases to a value such that plume-water is equal in density to lake water, the plume separates from the bottom, spreads horizontally, and is subject to strong Coriolis effects. Entrainment of lake water becomes less important, and the flow is quickly deflected toward the right-hand shore of the lake.

In this situation, coarse material is deposited rapidly at the point of entry of the stream into the lake, and a delta progrades into the lake. Fine sediment is transported into the body of the lake by the horizontally spreading plume, and is deposited (rapidly) only after it exits the base of the turbulent plume and enters the relatively still hypolimnion (Pharo and Carmack, 1979).

A discontinuous influx of sediment could give rise to individual graded beds within summer silt layers. In fact, Sturm (1978) has suggested that discontinuous sediment influx into a lake while it is stratified (and assuming stratification is seasonal) is the only way to obtain "classical varve" couplets. Thus, both discontinuous density underflows (turbidity currents) and discontinuous interflows can explain multiple graded beds in rhythmically bedded lacustrine sediments.

Rhythmically bedded silt-clay couplets are the end of a continuous spectrum of sedimentary deposits that begin with the outwash stream, generally pass through the lacustrine delta, and end with lacustrine deposits (Fulton and Pullen, 1969; Gustavson and others, 1975; Shaw, 1975; Killworth and Carmack, 1979).

Where sub-ice channels empty into standing water, coarse esker deposits may grade laterally and distally into rhythmically bedded deposits (De Geer, 1912; Agterberg and Banerjee, 1969).

No sub-ice channel deposits were identified in the Neoglacial deposits of the Muir Inlet region. In many cases, not enough lateral exposure was available to evaluate the geometry of the depositional systems responsible for the lacustrine sequences, and in the one instance where the geometry was well preserved, deposition was from the outwash stream-delta-lake system.
Within such systems, the contrast in grain-size between the various deltaic environments will determine whether or not primary sedimentary structures and stratification sequences will differ greatly from environment to environment (Gustavson and others, 1975). Where grain-size differences are small, as in the case of fine-grained outwash prograding into a lake, stratification sequences will be similar in all environments. In extreme cases, a lack of coarse sediment can cause a lack of foreset bedding on deltas and the creation of stable vertically aggrading distributary channels (Smith, 1975).

Sequences in the Muir Inlet region do not appear to have lacked coarse-grained material, and the one complete sequence discussed below consists of a Gilbert-type delta (Gilbert, 1885; Stanley and Surdam, 1978) that prograded into an ice-marginal lake.

Sediment Deposition in the Neoglacial Lakes of the Muir Inlet Region

The response of local streams to the development of Neoglacial lakes appears to have depended upon whether or not sedimentation at a site was great enough to keep pace with rising lake waters. It is suggested that lacustrine sediments that consist of coarsening upwards sequences indicate that sedimentation was rapid enough to cause prograding delta systems to develop. Fining upwards sequences were deposited where sedimentation was too slow to keep pace with deepening lake water.

As discussed in the section on glacial geology, oxidized, poorly-sorted gravels usually form the substrate upon which lacustrine sediments were deposited. These gravels are interpreted as outwash deposits that prograded into the fjords of the Muir Inlet region during the Hypsithermal, and aggraded to the point of filling the Inlet and its tributaries. Where they have been dated in Muir and Wachusett Inlets, they range in age between about 7,000 to 3,000 yrs B.P. In Adams Inlet, similar gravels are as young as 1,700 yrs B.P. Whatever their age, lacustrine sediments are very much alike from site to site.

Specific descriptions of two sites believed to represent typical examples of Neoglacial lacustrine deposition are discussed next.
Adams Valley Delta (A Coarsening Upwards Sequence)

A small Gilbert-type delta that prograded away from glacier ice into an ice-marginal lake that was on the order of 100-150 m deep is partially preserved in Adams Valley on the south side of Adams Inlet (Figures 9 and 10). The geomorphic form of a portion of the delta escaped later glacier overriding.

Apparently, Neoglacial ice advanced from the southwest into Adams Inlet. This blocked local drainage and formed a lake. As ice advanced, it was pinned on the bedrock highs on either side of the mouth of Adams Valley (northwest edge of the facies map, Figure 9). Coarse to fine outwash rapidly prograded away from the ice and into the lake (McKenzie, 1968). No collapsed bedding that would suggest deposition of gravels against glacier ice was observed, but McKenzie (pers. comm.) reported seeing such features on the northwest side of the delta in 1966-1967.

The top of the delta is roughly 200 m in elevation. This certainly represents a rough limit to the elevation of former lake level in Adams Valley. The outlet to this lake is not preserved. No col low enough to have served as an outlet occurs in the upper reaches of Adams Valley. McKenzie (1968) suggested subglacial, englacial, supraglacial drainage, or some combination of these. However, it seems unlikely that a free ice-face was present along the northwest lake shore for any length of time. Deltaic sediments probably flanked the entire ice-margin, but lacustrine outflow may have kept an outlet open to the northeast.

The facies map (Figure 9) suggests thicker sediments on the west side of the delta. This may have been the result of Coriolis effects, of a drainage way to the northeast, or of later glacier erosion. Well-sorted sands that contain some gravel lenses and large trough and planar crossbeds (Berg Formation of McKenzie, 1968) occur in abundance along the south side of Adams Inlet and in Endicott Gap at the southeast end of the Inlet (c.f. McKenzie, 1968, Figure 2). These deposits generally cover lacustrine rhythmites of the Adams Formation. Cursory observations of transport directions indicated by crossbeds within Berg Formation sediments suggest that ice-marginal streams carried sediment to the east, and that Endicott Gap may have acted as the ultimate outlet for the lake or lakes in Adams Inlet.

Bedrock is not exposed in the Endicott Gap region. Neoglacial lacustrine, ice-contact stratified drift, and outwash that are all younger than the Berg Formation cover the drainage divide. The divide is now some 30 to 40 m higher than the top of the delta in Adams Valley, but this difference may be the result of the deposition of thick post-delta sediments. In part, the difference is probably a function of the removal of the topmost sediments of the delta by overriding glacier ice.
Figure 9. Map of the surficial geology near and the sedimentary facies of the Adams Valley delta.
Figure 10. Composite stratigraphic section through Adams Valley delta. The letters refer to sites shown in figure 9.
Figure 9 shows a simple facies map for the delta. Topset and foreset beds are present near the top of and in gullies that are actively dissecting the delta. Good exposures also occur along the bottom of Adams Valley where rhythmically-bedded lacustrine sediments are undercut by lateral migration of channels of Adams River. Large alluvial fans cover most of the flanks of the delta. Figure 10 shows, schematically, two measured sections that represent topset and foreset deposits, and bottomset and rhythmic lacustrine sediments respectively.

**Topset Beds**

Topset beds consist of stacked and overlapped horizontal lens- or wedge-shaped bodies of coarse sediment 0.5 - 2 m in maximum thickness (Figure 11). The lenses are tens of m wide, and often many tens of m long. A cobble layer one or two clasts thick often forms the base of such a lens. Clasts 20-40 m in longest dimension commonly make up the layers, and clasts often show rough imbrication. Individual beds generally fine upwards into coarse to fine gravel. This gives a general impression of normal graded bedding, although reverse grading does occur. Coarse clasts are in framework support, and a medium to coarse sand-matrix usually fills the voids between the clasts.

Finer-grained beds also occur. Many are composed of coarse sand to granule-sized material that is roughly horizontally stratified, and that contains abundant floating clasts of coarse gravel or fine cobble size (Figure 12).

Elongate, wedge-shaped bodies of medium-grained sand often occur on top of coarser-grained deposits. These usually contain small-scale, current-ripple, cross-laminations 2-4 cm in height. Epsilon crossbedding (Allen, 1970) is seen rarely and where observed, it contains reactivation surfaces, climbing bedding, and it laps onto fine to coarse gravels (Figure 13). Sand bodies are usually only 10-20 cm thick.

The overall aspect of the deposits suggests that they were deposited in a braided stream environment where flow was often in the upper flow regime and was capable of transporting large clasts. Coarse deposits represent deposition by traction and from suspension on bar tops, distal bar margins, small, shallow channels, or as overbank sediments (Fahnestock, 1963; Williams and Rust, 1969; Rust, 1972; Boothroyd and Ashley, 1975; Rust 1978).

**Foreset Beds**

Foresets underlie topset beds, and a marked erosional contact separates the two (Figure 14). The foresets dip at maximum angles on the order of 15-25° where they meet overlying topsets. Concave erosional surfaces with dip angles up to 30° truncate some foresets.
Figure 11. Horizontal coarse clastic topset beds of Adams Valley delta. The largest clasts are 20-40 cm in longest dimension.

Figure 12. Floating pebbles in normally graded topset beds.
Figure 13. Epsilon crossbedding with reactivation surfaces, found in topset beds. Flow was oblique to the right coming out of the photo. The sands lap onto gravel to the left.

Figure 14. View northeast of the contact between topset and the foreset beds of Adams Valley delta. The area shown is site A of Figure 9.
just below their contact with topset beds. These surfaces have been preserved by rapid sedimentation over them, and they are interpreted to be slump scarps.

The angle of dip of the foresets decreases down the delta face as does the grain size of the sediment comprising them. Foresets thin distally and sweep tangentially into bottomset beds. Figure 15 shows transport directions obtained from foreset beds and small scale ripples.

Stratification sequences in the foresets are similar to those described for topsets. In general, clast sizes are smaller, and sand is more common in foresets than in topsets. Individual beds consist of broad sheets of fine- to medium-grained gravel 0.5 to 1 m thick. Sheets of sand up to several tens of cm in thickness occur over most gravel units. These often contain simple, non-repeated sequences of type-A climbing ripples (Jopling and Walker, 1968) as shown in Figure 16. Within upper foresets, individual ripples are long and low, and have a low angle of climb. This suggests primarily traction transport and deposition of sediment, and relatively high flow velocities (Hopking and Walker, 1968; Harms, 1979).

Occasionally, elongate, lenticular bodies comprised of cobble-to coarse gravel-sized clasts were observed within the upper forest beds (Figure 17). They are 4 to 5 m in maximum thickness, and up to 40 m in width. As in topset beds, a finer clastic matrix fills the voids between framework supported clasts. The bodies dip down the face of the delta at the same angle as surrounding foreset beds, and they appear to fill channels scoured into pre-existing foresets.

Apparently, high-energy streams issuing from distributary channels on the delta fed coarse, clastic sediment to channels that continued down the delta front. Rapid aggradation probably caused bars to form and channels to fill. The increased importance of coarse and medium sand deposition on the delta front suggests a rapid loss of energy by streams, especially near turbulent flow margins, as they entered the lake. Gravity flows probably continued down the delta front in the manner described by Gustavson (1975) and Hamblin and Carmack (1978). The behavior of these flows was probably governed by turbidity contrasts with ambient lake water, bottom slope and friction, and entrainment of lake water (Middleton and Hampton, 1973).

Bottomset Beds

These consist of horizontal blankets of medium- to finegrained sand in 10 to 40 cm thick graded units (Figures 18 and 19). Climbing ripples comprised of fine sand often occur in sets 10 to 30 cm thick at the top of graded beds. Type-A climbing ripples are common, but type-B climbing ripples (Jopling and Walker, 1968) and draped laminations (Stanley, 1974; Gustavson and others, 1975) also
Figure 15. Rose-diagram of sediment transport directions on the Adams Valley Delta.
Figure 16. Low angle (type-A) climbing ripples within a sandy bed in foreset beds of Adams Valley delta. The ripples show up faintly on the cleared area.

Figure 17. Thick body of clast-supported channel-cobbles enclosed by finer grained forest beds.
Figure 18. Horizontally stacked sandy bottomset beds southeast of site A of Figure 9.

Figure 19. Climbing ripples and massive sand layers in proximal bottomset beds near site B of figure 9. Note the dropstone above and to the left of the lens cap.
occur. Less commonly, horizontally laminated sands and coarse silts are found above graded sands. Contacts between sedimentation units are sharp.

The angle of climb exhibited by trains of climbing ripples often increases gradually upwards, suggesting a gradual decay in the strength of the flow responsible for sediment deposition. In other cases, fluctuations in the angle of climb were observed that suggest periodic fluctuations in flow strength.

Load casts and flame structures are often developed in these beds, and their presence suggests rapid deposition of sand on finer-grained, water-saturated sediments.

In many instances, sand layers appear to have massive bedding or only slightly graded bedding. Ripples were often initiated directly on the upper surface of the massive to graded sand bodies, and load casts developed at the base of the ripples.

Clay layers 1 to 2 cm thick are found within bottomset sequences. They commonly bound 6 to 12 massive to graded sand climbing ripple sequences.

Whether the clays are annual or not is open to speculation. If they are, then only 6 to 12 major depositional flow events were recorded at any one site near the base of the delta front each year.

The sequences of massive to slightly graded sand topped by climbing ripples are similar to partial Bouma sequences described by many authors from deep-sea fan environments (Chipping, 1972; Hendry, 1973; Normark, 1978; Walker, 1979). Beds with structures commonly found in the lower portion of the ideal Bouma sequence (Bouma, 1962) are often ascribed to deposition in proximal positions within basins (Walker, 1967), and are often said to represent channel facies (Chipping, 1972). Mutti (1977) has described "thin turbidities" very similar to bottomset sequences of the Neoglacial Adams Valley delta that he found in Eocene deep-sea fan deposits in Spain. He suggests a channel-margin depositional environment for these beds.

Shaw (1975) suggested that stacked horizontally bedded sand units found in the Okanagan Valley in Canada had their origin as deposits in distributary channels of glacial melt-water streams on or near deltas. Not enough observations were made to determine whether or not the bottomset beds of the Adams Valley delta represent distributary channel deposits formed by gravity flows spreading away from the base of the delta. In terms of flow conditions, the massive to slightly graded sands appear to be the result of deposition from concentrated gravity flows, perhaps near the head or base of concentrated turbidity currents (Stauffer, 1967; Middleton and Hampton, 1973).
Climbing ripples immediately above the graded to massive sands indicate a rapid transition to less concentrated flow, perhaps the margin or tail of a turbidity current. Intervening horizontally bedded (Bouma B) units were not deposited. Concentrated turbidity currents that were generated by discontinuous river flow or by sediment slumping on the delta front probably decreased in velocity and/or spread laterally as they moved out onto the flat lake bottom. They deposited sandy sediment rapidly as their competence decreased.

Figure 20 shows a positive print of an x-radiograph of a thin epoxy-resin peel obtained from a sequence of climbing ripples and overlying massive sand. Micro-laminae can be seen in the stoss- and lee-side laminae of individual ripples, and slumped sediment is preserved on the lee-side of some ripples. These features may indicate rapid disequilibrium deposition of sediment (Stanley, 1974). The rapid transition from Bouma A to Bouma C beds in the deltaic bottomsets is consistent with disequilibrium flow conditions. Laminae above the 'ripples (and interbedded with them in places) that dip at low angles in the direction of flow may represent foreset laminae of microdeltas. The formation of such structures in sequences of climbing ripples also suggests disequilibrium flow.

Injection or load structures and rip-up structures can also be seen at the base of the massive sand unit near the top of the x-radiograph. They are overturned in the direction of flow as indicated by the underlying ripples (away from the delta front), and flows must have exerted a large amount of drag on underlying deposits.

Distal bottomsets are difficult to distinguish from proximal lacustrine deposits. The distinction is largely arbitrary, and is based on the occasional presence of climbing ripples in distal bottomset beds. In distal bottomsets, fine sand and coarse silt appear in graded units several to several tens of cm in thickness (Figure 21). Clay layers 1 to 2 cm thick appear every 0.5 to 1.5 m vertically. Clay rip-up clasts are present in both proximal and distal bottomsets and suggest a high degree of sediment reworking within these beds.

Load structures are common in distal bottomsets, again indicating deposition of sand on water-saturated, finer-grained sediment.

Lacustrine Sediments

These sediments are comprised of graded laminae (both normal and reverse) and some apparently structureless laminae of very fine sand and silt. Total thickness of the coarse part of a silt-clay couplet is usually several tens of cm. The coarse laminae probably represent deposition from distal, low-concentration turbidity currents. Clay layers are usually 2 cm or less in thickness, and
Figure 20. Positive print of an x-radiograph of a sediment peel obtained from the bottomset beds of Figure 19. Climbing ripples in the lower part of the print contain microlaminations; and some have avalanche deposits on their lee sides. Microdeltaic deposits succeed the ripples, and traction deformation and injection structures can be seen below the upper massive sand. Dark bands in the upper sand represent irregularities in peel thickness, not sedimentary structures.
Figure 21. Distal bottomset beds showing laminated sands and silts and load structures.

Figure 22. Laminated silts and thick clays (light colored bands) in proximal lacustrine rhythmites near site C of Figure 9.
30 to 50 graded or structureless laminae occur in the silt layers (Figure 22). This suggests that 30 to 50 flow events occurred at these sites between periods of clay deposition. If clay layers are annual features, the discussion on bottomset beds suggests that discrete flow events that deposited sediment on the delta front and on proximal bottomset depositional sites (6 to 12 events between clay layers) spread laterally over the lake bottom and overlapped one another where proximal lacustrine sediments accumulated. It is also likely that the greater number of events recorded in proximal lacustrine sediments is related to depositional events caused by interflows that were recorded at lacustrine sites but were not recorded (or preserved at any rate) in deltaic bottomsets.

Clay layers in bottomsets and proximal lacustrine beds often contain discrete, fine sand laminae 2 to 5 grains thick. Shaw and Archer (1978) have suggested that thicker coarse laminae found in "winter" clay layers of glacial rhythmites in the Okanagan Valley were formed as the result of turbidity flows (probably generated by slumping on the delta face) during winter months. This may also be the case for the laminae in Adams Valley, but this author is at a loss to explain how the fine, clean sand laminae enclosed in clay of apparently uniform size could be the result of a turbidity flow. Perhaps only sand of a uniform size and in very dilute turbidity currents was involved. One possible mechanism for generation of these "winter" turbidity currents could be small-scale slumping of sediment from restricted areas of the delta, perhaps from fine-grained foreset deposits.

Such sand layers are not present in thinner distal lacustrine sediments. Total couplet thicknesses within distal sediments are on the order of 1 to several cm (Figure 23). The silt portion of the couplets are comprised of many tens of very thin (tenths of a mm) graded to massive laminae of fine- to medium-grained silt. Clay laminae usually comprise one half or less of the total couplet thickness. The silt laminae probably represent deposition from dilute density underflows or interflows, as in more proximal deposits.

Occasionally, sand bodies with graded bedding or climbing ripples occur in thin lacustrine rhythmites. These are apparently deposits that were laid down by relatively strong gravity flows. In one case, Figure 23, a fine pebble layer about 1 cm thick that had a silty matrix was observed within the silt half of a rhythmite couplet. This too, is probably the result of a concentrated gravity flow.

One can see a gradual but progressive shift from dominantly traction transport and deposition of sediment in the topset beds of the delta to dominantly suspension transport and deposition in thin lacustrine rhythmites.
Figure 23. Distal lacustrine rhythmites, clay layers are dark colored. A prominent pebble-layer appears low in the shadow covered area and on the surface of the displaced blocks in the foreground. The photo is about 1.5 m across.

Figure 24. Dropstones in distal lacustrine rhythmites near site A5. The knife is about 20 cm long. Note also the thick clay (dark colored) layers at this site.
Other Features of Lacustrine Rhythmites

One of the most distinctive features of the lacustrine rhythmites (besides the repetition of silt-clay couplets) is the abundance of dropstones in the sediments (Figure 24). These occur in silt and clay laminae alike, and generally show the downward deformation of sediments beneath them that is so characteristic of ice-rafted clasts (Hardy and Leggett, 1960). Clasts occur singly and in groups arranged horizontally along bedding planes. Such groups were apparently deposited simultaneously on the lake bottom, perhaps from a single ice-berg as it overturned and dropped its surface debris into the lake.

Ice-rafted clasts from a few mm to 60 or 70 cm in longest dimension were observed in lacustrine sediments in the Muir Inlet region. The clasts varied from very well-rounded to extremely angular, and although no pebble-counts were obtained, most appeared to be derived from diorite and metasediments of the Muir Inlet region.

It is interesting to note that very few large ice-rafted clasts were observed in lacustrine sediments in the Muir Inlet region. The clasts varied from very well-rounded to extremely angular, and although no pebble-counts were obtained, most appeared to be derived from the diorite and metasediments of the Muir Inlet region.

It is interesting to note that very few large ice-rafted clasts were observed in Adams Valley. This may be a reflection of the fact that an outwash-apron separated glacier ice from the lake. Perhaps only clasts in ice masses that were rolled to the lake by melt-water streams fell into the lake, and the size of the ice-blocks that could be rolled, limited the size of the clasts delivered to the lake. This rolling process was observed in the Seal River in front of Casement Glacier during the summer of 1977.

Deformation features are also quite common in most exposures of glaciolacustrine rhythmites. In many cases, these are normal and reverse faults with 1 cm or less of net slip. These are similar to ones described by Smith (1959). Several more spectacular faults were observed in Adams Valley, and one is shown in Figure 25. This is a thrust fault in thick lacustrine rhythmites, and it has well developed drag folds. The fault was observed near site C of Figure 9. The lower extent of the fault was covered by slumped debris, and its upper edge was covered by post-faulting lake-sediment deposited soon after the feature formed. This suggests that the very fine sands and silts behaved in a coherent, brittle manner, even shortly after deposition.

Clastic dikes were not observed in Adams Valley, but they were seen in other exposures, particularly where thick beds of coarse silt- to medium sand-sized sediment occurred.
Figure 25. Thrust fault in proximal lacustrine rhythmites near site C of Figure 9. The uppermost sediments are not cut by the fault and therefore post-date it.

Figure 26. Severely folded lacustrine rhythmites exposed near river level. Undisturbed sediments blanket the folded ones. The vergence of the folds is southeast away from the delta.
Numerous penecontemporaneous folds were observed at nearly all sites. These interfered considerably with "varve" thickness studies. A series of folded couplets is shown in Figure 26. Beds are complexly folded and overturned. The vergence of the folds is to the southeast away from the Neoglacial delta. McKenzie (1968) reported several diamictons within the Adams Formation (c.f. McKenzie, 1968, Figure 10, column B) and basal gravels that were sheared up into the lacustrine rhythmites in Adams Valley. He suggested that grounded, glacier ice had periodically shoved the sediments and deposited up to four diamicton units. An alternative explanation is that the diamictons are the result of large slumps of lacustrine sediments toward the center of the lake basin. This author observed severely distorted, but recognizable bedding in the diamicton that he measured in Adams Valley, and other authors have noted slump features near the base of deltas in modern lakes receiving heavy loads of sediment (Fulton and Pullen, 1969; Gilbert, 1975).

Another feature not observed in Adams Valley, that is seen at other sites, is channel cut-and-fill. Channels up to 0.5 m deep and 1 to 1.5 m wide were observed, but not often. The bottom of such channels were often filled with sediment that is coarser-grained than surrounding deposits into which the channels were incised. Usually, the top of a channel was filled with rhythmically bedded sediments similar to those around the channel. These features may well have been cut by erosive gravity flows. There is no evidence to suggest subaerial erosion of the channels.

Fining Upwards Sequence at Site W1

A Neoglacial lacustrine sequence approximately 18 m in thickness occurs at site W1 on the north shore of Wachusett Inlet (Figure 27). Oxidized Hypsithermal gravels and Neoglacial lacustrine deposits fill gullies cut into local bedrock. The elevation of bedrock spurs between the gullies appears to have controlled the depth to which pre-Neoglacial ice advance sediments were eroded by Neoglacial ice.

As in Adams Valley (and elsewhere in the Muir Inlet region), the contact between lacustrine deposits and the underlying poorly-sorted, oxidized gravels is very sharp. However, instead of thin, rhythmically-bedded lacustrine sediment, a thick-bedded sequence of massive silts rests on the gravels at site W1. The remains of Equisetum (horsetail rushes) were found in growth position in these silts, and the rushes were apparently buried rapidly by the aggrading silts. The silts are believed to represent deposition in very shallow water near the lacustrine shore or perhaps between distributary channels of a stream entering the lake.
Figure 27. Stratigraphic section through the middle member of the Van Horn Formation at site WI, Wachusett Inlet. The section is described in Appendix B.
Upwards in the sequence, bodies of well-sorted, medium-grained sand up to 2 m thick were observed. These contained planar and trough crossbeds with several tens of cm of relief. Coarse sand and fine gravel laminae occur within these beds. The crossbedded units are interbedded with thick roughly horizontally bedded coarse sands and fine gravels. Clay layers several cm thick occur between major sedimentation units. Sand and gravel bodies thin and fine toward the top of the sequence, and clay rip-up clasts are common in the thinner sand bodies.

The sequence is interpreted to reflect deposition in distributary channels (gravels and sands) and perhaps on distributary mouth bars (crossbedded sands). Similar sequences are known from Recent and Pleistocene glaciolacustrine deltas (Shaw, 1975; Gustavson and others, 1975), and they are represented in many modern deltas (Coleman and Wright, 1975).

A three m thick unit of thin lacustrine rhythmites caps the Neoglacial lacustrine sequence. These are very similar to the basal lacustrine sediments in Adams Valley. They have silt layers that are thicker than overlying clay layers, dropstones, and penecontemporaneous deformation structures. The rhythmites have been erosionally truncated and channels filled with coarse sand and fine gravel are incised into lake sediment. It is not clear whether these are pre- or post-ice advance sediments. The lack of glacio-tectonic deformation structures in the rhythmites suggests that the gravels preceded ice advance.

The entire sequence is interpreted as either the response of a stream to progressive deepening of the lake with progressively deeper water sedimentation, or as the lateral migration of distributary channels near a delta mouth. Channel migration would have caused alternating channel, channel mouth bar, and interdistributary deposition. Perhaps it is more likely that both processes were occurring at the same time.

Most of the stratigraphic sequences of Neoglacial lacustrine sediments that were measured for this study (Appendix B) are similar to the two just described. Other more distal sites consist of thin, rhythmically bedded sediments sandwiched between coarse outwash gravels. Such sites were probably not as close to points of sediment input to lakes as the sites just described.

The valley in which site W1 is located provided a laterally continuous exposure of thin lacustrine rhythmites. Couplet thicknesses were measured at site W1, and they were also measured 50 to 60 m up the valley at site W5.

**Varve Correlation Studies**

Detailed records of the thickness of successive silt-clay couplets were obtained from a number of outcrops (records for 18 outcrops are listed in Appendix A). No success in correlation of
the records was achieved. Figure 8 suggests why this is so. The diagram shows thickness records for the two parallel transects measured in Wachussett Inlet. Overall, the records match well, but there are a number of couplets at either site that are not present at the other. Usually two thin couplets at one site are represented by one thick one at the other site. Several explanations seem plausible. Deposition of clay layers may have been discontinuous over the lake bottom. Also, remembering the evidence for strong gravity flows in Adams Valley (and the presence of clay rip-up clasts in deltaic bottomset and lacustrine sediments), it seems likely that erosion by gravity flows may have removed some clay layers. A third possibility is that the rhythmites represent deposition from individual (although perhaps discontinuous) flow events generated by slumping on a delta front or sediment input from separate streams or distributary channels. This would lead to overlapping couplets of limited lateral continuity and of unique thickness.

The problems encountered in attempting to correlate couplets over distances of tens of meters are multiplied by orders of magnitude with distances between outcrops are several to tens of kilometers. Couplets may be annual deposits, but erosional removal or nondeposition of couplets or parts of couplets render longdistance correlation of thickness records impossible. This problem is especially significant in the Muir Inlet region where most of the preserved lacustrine rhythmites appear to have been deposited near the source of sediment input where strong gravity currents appear to have been active.

In addition, as stated in the section on regional climate, precipitation (which would strongly influence summer runoff and sedimentation) can be quite local and variable within the study region. The sequence of thickness variations of annual couplets at two sites tens of kilometers apart might be quite different owing to different precipitation histories.

There is also evidence that clay deposition is not necessarily an annual event. Figure 28 shows a sequence of rhythmite that occur above thick beds of massive and graded silts and fine sands. Each rhythmite consists of a thick, basal, laminated silt 5 to 10 cm thick. Above this are thin, alternating, graded silt-clay couplets (usually 5 to 8) that are similar in all respects to other thin lacustrine rhythmites. It is possible that each thick silt and its overlying silt-clay couplets represent one annual "varve". But if this is the case, then 5 to 8 clay layers were deposited each year. This leaves open the possibility that none of the clay layers in the region record annual events.
Figure 28. Rhythmites of site M5 that consist of thick silts and multiple silt-clay couplets. Thick bedded laminated sands underlie the rhythmites.
In either event (annual or nonannual couplet deposition), the fine-grained sediment at all sites studied represents a short period of deposition. Usually only 150 to 300 couplets are present at a site, so 150 to 300 years or less are represented by the couplet records at any one site.

Grain-size Analyses

Some of the samples obtained for paleomagnetic analyses were later analyzed to obtain information about their grain-size distributions. The grain-size data was helpful in assessing the significance of some of the paleomagnetic data (particularly magnetic bulk susceptibility data). However, the hydraulic significance of the grain-size data is limited because more than one sedimentation unit was analyzed in every case.

Even with this limitation in mind, it can be said that the grain-size distributions (Figure 29) are similar to distributions obtained by other authors for rhythmically bedded glaciolacustrine sediments (Peach and Perrie, 1975). Peach and Perrie (1975) looked at the grain-size distributions of samples from “varved” sediments scraped from horizontal swaths only 0.5 mm wide in order to study the internal grain-size variations within single “varves”. They found that the shapes of cumulative distribution curves for these subsamples were all quite similar. They also found that there were slight fluctuations in median grain-size values within the silt portion of individual “varves”. These variations were attributed to flow variations inferred to be on the order of several days in duration. Flow variations can also be inferred for the rhymites in the Muir Inlet region based on the presence of multiple graded beds within the silts.

Grain-size analyses completed for this report were terminated at the 11 reading. This was done as a matter of necessity in order to finish the large number of analyses that were run. At the temperatures at which the analyses were completed, the 11 reading is not taken until 56 to 60 hours after an analysis is begun. If the analyses had been taken to 13 or 14, many days would have been required to obtain data from a small number of samples. As a result of not carrying the analyses further, the only meaningful statistic that was obtained for most samples was the median diameter (size of the 50th percentile on the cumulative distribution curve). The statistics of Folk and Ward (1958) could not be
Figure 29. Grain-size distribution plot for twelve samples obtained at 25 cm vertical intervals from site WL.
calculated for dispersion, skewness, or kurtosis because all of these statistics require that the size of the 95th percentile on the curves be known. In most cases, even the 84th percentile was not known, so only the Trask mean, $P_{25} + P_{75}/2$, could be determined. Mean values are not reported, but in the cases where they were obtained, they were always about 0.2 units finer than median grain-size values. This indicates that all of the cumulative distribution curves for the glaciolacustrine rhythmites are probably slightly negatively skewed. This may be a relict effect caused by the inclusion of clay layers in the samples analyzed.

A number of the samples shown in Figure 29 have a marked coarse tail. Upon sieving the samples prior to pipette analysis, coarse angular sand and some small pebbles (from well rounded to very angular) were left on the sieve. This material is responsible for the coarse tails, and it is interpreted as the product of ice-rafting.

The median diameters obtained from the distribution curves are very similar to the values obtained by Peach and Perrie (1975) for Pleistocene "varves" similar to the Muir Inlet sediments. This may be a fortuitous circumstance, but it may also reflect the similar hydraulic sorting history of sediments derived from a similar source, glacially ground bedrock.

Summary

In conclusion, it appears that the lacustrine sediments of the Muir Inlet region have a glaciolacustrine origin; however, silt-clay couplets are impossible to use to correlate spatially separated outcrops of lacustrine sediment. Couplets must be used with caution to determine deposition rates or the number of years of lacustrine deposition at a site. Without some independent means to prove the annual nature of the couplets, it seems unwise to call them varves and use them as unqualified annual increments.

It must also be stated that the author has not proved that all of the couplets of the Muir Inlet region are not annual in nature. Some may be annual and others may not be. But it does seem certain that their use to obtain absolute time estimates must be exercised with caution.
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