

THE QUATERNARY HISTORY OF THE CORDILLERAN ICE SHEET FRINGE,
ASHLEY LAKE AREA, MONTANA

by

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A thesis submitted in partial fulfillment
of the requirements for the degree

of

Master of Science

in

Earth Science

MONTANA STATE UNIVERSITY
Bozeman, Montana

January 2004

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Denny Lane Capps
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ACKNOWLEDGEMENTS

1. Dr. William Locke, advisor, who was always there, introduced me to the problem and oversaw the project work
2. Dr. Kenneth Pierce, committee member, who shared his extensive field knowledge and encouragement throughout the project
3. Dr. John Montagne, committee member, who selflessly gave his time to the betterment of this thesis
4. This study was primarily funded by a generous donation of the Donald Smith Memorial Scholarship Fund
5. Jeff Deems – friend and colleague, continuous encouragement and positive criticism
6. Dr. Larry Smith – Montana Bureau of Mines and Geology – regional geologic information and encouragement
7. Gamma Family – land access, hospitality and local information
8. Scott Larson – field assistance
9. Plum Creek Timber Company – land access and aerial photos
10. Stoltz Timber Company – land access and maps
11. Flathead National Forest – especially Jane Kollmeyer, District Ranger – painless permitting and campground use
12. Dr. Sharon Eversman – MSU Department of Ecology – plant identification
13. Brist Family – land access
14. Montana DNRC - lake bathymetry

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ABSTRACT

The Quaternary glacial history of the Ashley Lake area, approximately 20 km west of Kalispell, MT, was deciphered using classic techniques of glacial geology. Unlike the Laurentide Ice Sheet, the multiple advances of the Cordilleran Ice Sheet have not been well defined. Ashley Lake nestles within the central Salish Mountains, an area interpreted by some previous workers (e.g., Alden, 1953; Richmond, 1986; cf. Waitt and Thorson, 1983) as surrounded by, but not covered by the fringes of the lastglacial Cordilleran Ice Sheet. This study attempts to establish a foundation for the Quaternary history in this region for the Cordilleran Ice Sheet.

Specific interests were to define the spatial and temporal attributes of the pre-lastglacial maximum, lastglacial maximum, and lastglacial recession. The techniques employed include: 1) mapping of glacial features to reconstruct glacial margins and ice flow directions; 2) mapping of the distribution of erratic boulders to determine ice sheet maxima; 3) radiocarbon dating; 4) tephrochronology to numerically date deposits; and 5) measurement of weathering rinds to determine the relative age of deposits.

Abundant glacial landforms combined with erratics allowed a reconstruction of ice margins. Moraines, direct overflow channels, lateral meltwater channels, kames and glacial depressions are among the most important landforms mapped. Radiocarbon and weathering rind dating proved to be very limited in their usefulness. No reliably dateable, closely limiting carbon samples were found. Weathering rinds were too thin and inconsistent to provide for meaningful relative dating. Tephras known to exist in the region were found in ice marginal bogs and provided a minimum date for deglaciation.

The spatial retreat of the Cordilleran Ice Sheet here was determined and it was found that local topography greatly affected the lastglacial recession and therefore the lastglacial advance. The lastglacial maximum inundated the entire Ashley Lake area and its tributaries. The pre-lastglacial maximum was even more extensive and covered everything in the area except the highest ridges. The information gathered in this field investigation can now be applied to our knowledge of the Cordilleran Ice Sheet to fine-tune its spatial attributes.

INTRODUCTION

Problem

Unlike the Laurentide Ice Sheet, the multiple advances of the Cordilleran Ice Sheet have not been well defined spatially or temporally. The precise extents of multiple glacial episodes of the Cordilleran Ice Sheet in this area have not been determined due to a lack of detailed fieldwork. The precise timing has been difficult to derive because of the lack of materials suitable for radiocarbon dating and the necessity of applying a variety of relative weathering techniques (Andrews, 2000). There is a need for a field setting that would allow appropriate applications to be used to define more accurately the spatial and temporal attributes of the Cordilleran Ice Sheet along its margins; the Ashley Lake area of northwestern Montana appeared to be well located for this purpose.

Ashley Lake nestles within the central Salish Range of the northern Rocky Mountains, approximately 20 km west of Kalispell, Montana (Fig. 1). Most workers in this area generally agree on terminal ice margins of the lastglacial Cordilleran Ice Sheet. However, the Ashley basin, which is interpreted by all as contained within the ice margin, is interpreted by some previous workers (e.g. Richmond, 1986) as surrounded by, but not covered by ice, i.e. a nunatak. Other workers interpret the basin as having been completely covered by ice (e.g. Waitt and Thorson, 1983) (Fig. 2). With detailed use of appropriate techniques, this study site was used to constrain the late Pleistocene history of this region of the Cordilleran Ice Sheet.

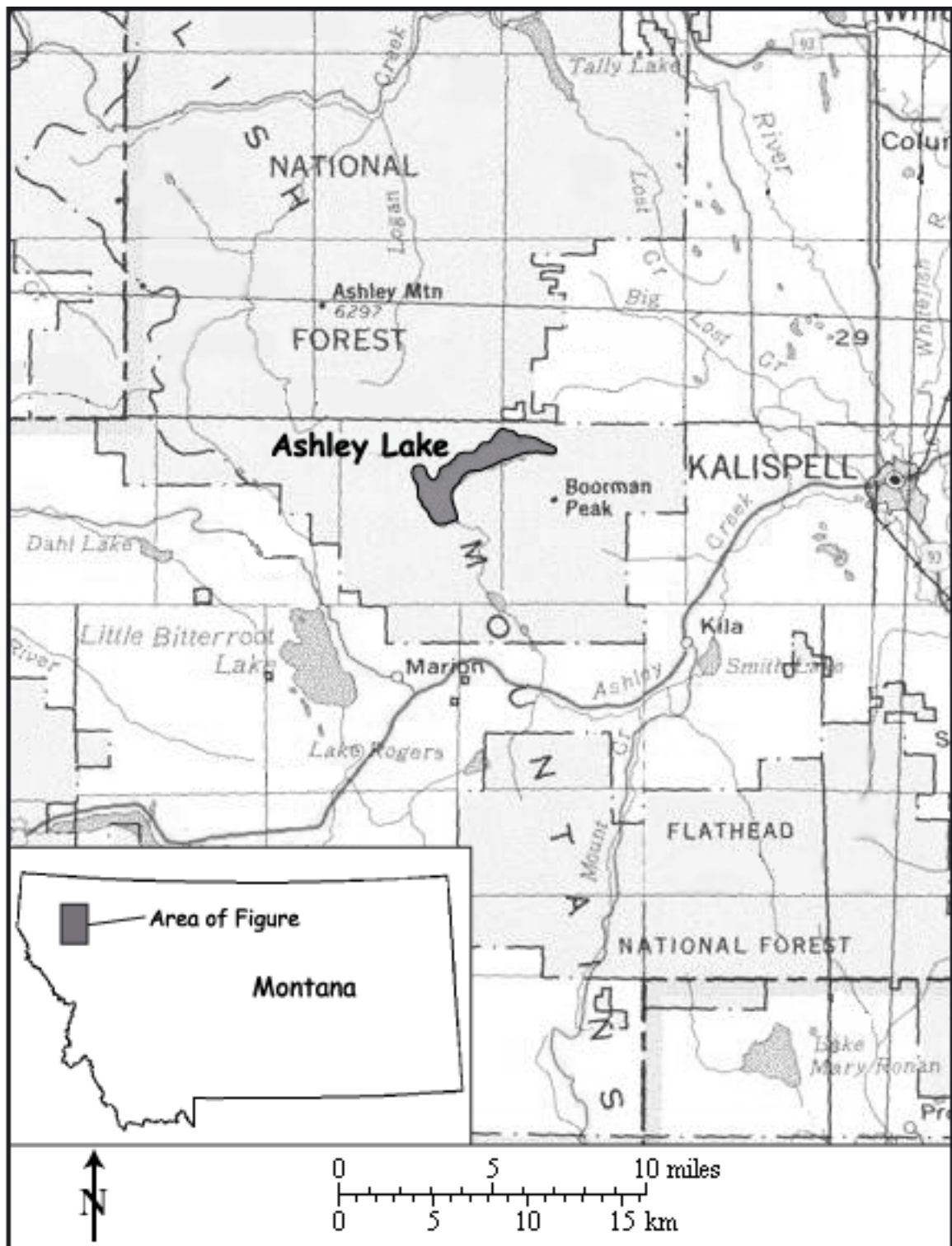


Figure 1 – Study area

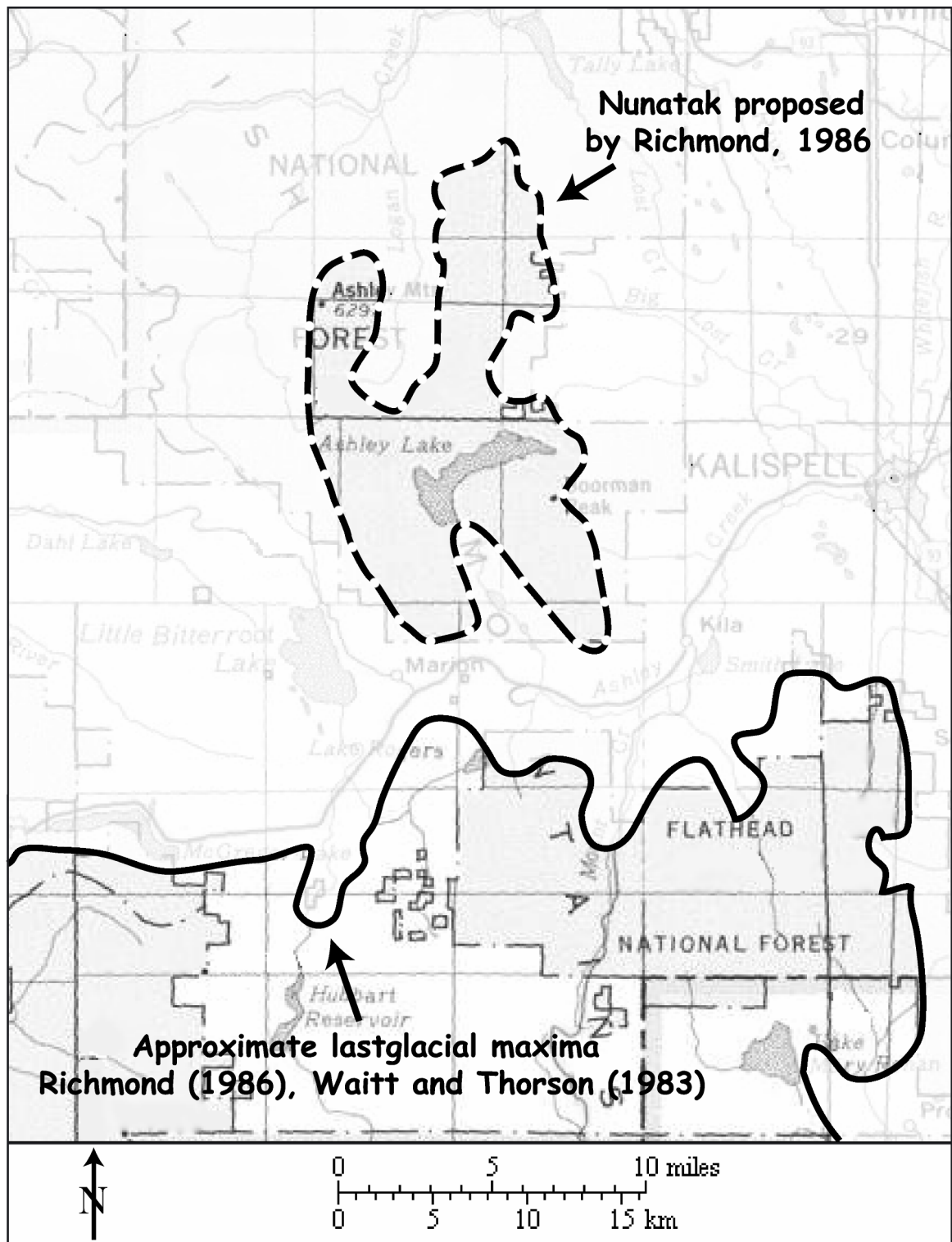


Figure 2 – Past worker's lastglacial Cordilleran Ice Sheet margins

Because of its position at the fringe of the Cordilleran Ice Sheet and the geometry of the basin itself, the Ashley basin is well suited to record progressive maxima and retreats of former ice sheets. Fringes are important areas of ice sheets; dateable materials tend to be concentrated and glacial landforms from multiple advances have a higher chance of preservation here. The basin is ideal to record retreat of the former ice sheet because of the multiple ways that ice could enter, and therefore exit, the basin. The lake, elevation 1203 m, is surrounded in the north, east and west by a relatively high mountain divide no less than 1372 m and as high as 1899 m. To the north and east, there are relatively high passes that the ice could have overtopped, thus spilling down on to the basin floor (Fig. 3). As the ice retreated through these passes, the piedmont lobes would eventually get pinched off as the ice elevation declined and the ice that remained in the basin by these routes would stagnate. Ice could enter the basin from the south where no barriers exist other than a gradual increase in elevation. The basin outlet is on the south side down an 11 km valley, opposite to the general approach of the ice out of Canada. After deciphering the local glacial history, this information was then be applied in a regional context.

This project sought to develop a local stratigraphy with subsequent correlation to regional stratigraphies. Richmond (1986) reported that in seeking a standard for correlation, the United States Report Committee concluded that in no region in the U.S. is a stratigraphic succession known to include deposits of all of the glaciations recorded in the country, and the deposits represent not more than 20% of the Quaternary glaciations in any single stratigraphic section. In addition, physical, biological and chronometric criteria for extension of time boundaries from a standard section across the country are

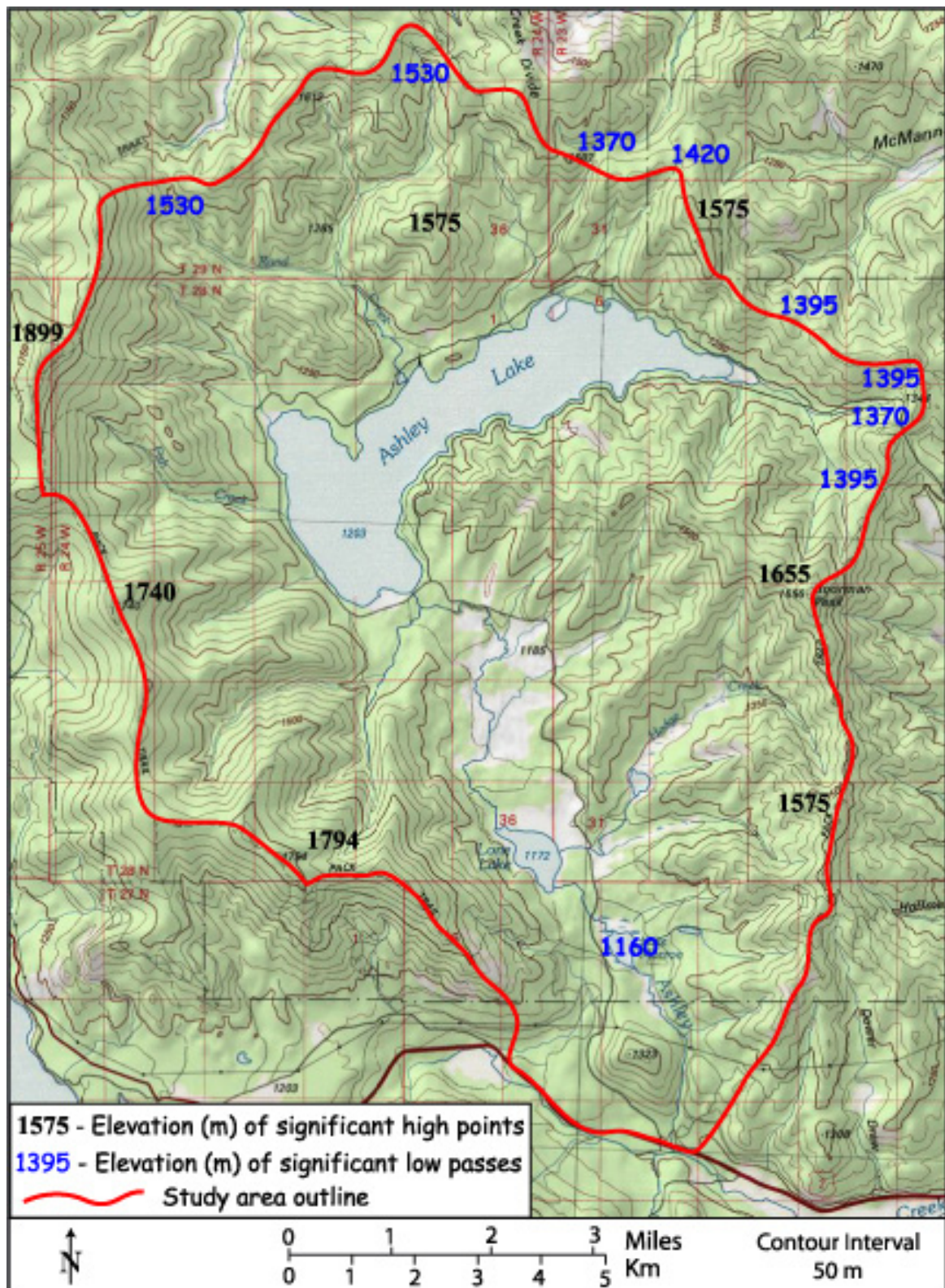


Figure 3 – Study area boundary and important elevations

lacking. Consequently, establishment of a standard stratigraphic section for Quaternary glacial deposits is not practical, although the marine oxygen-isotope stages provide the most complete proxy. Nonetheless, existing age controls permit grouping of the deposits throughout the country into arbitrary time divisions. The deposits relegated to each time division are separated from those of other divisions by one or more markers such as a paleosol, a nonglacial deposit, an identified tephra, a major unconformity, or an identified paleomagnetic polarity reversal. Stratigraphic assignments have differed also, and neither the terminology from the Rocky Mountains nor that from the Pacific Northwest is applicable to the Ashley basin (Richmond, 1986). Past workers in similar settings have used Laurentide Ice Sheet terminology, alpine glacier terminology and other local terminology (Table 1).

Setting	Lastglacial	Pre-lastglacial
Laurentide Ice Sheet	Wisconsin	Illinoian
Alpine glacier	Pinedale	Bull Lake
Cordilleran Ice Sheet	Mixed	Mixed

Table 1- Terminology Used by Past Workers

According to the Draft North American Stratigraphic Code, 1981, conventional approaches of lithostratigraphic and chronostratigraphic classification have proved impractical in stratigraphic studies of many upper Cenozoic rocks. The lithic heterogeneity and discontinuous geographic distribution of many deposits present special problems. In addition, units in vertical sequence are commonly separated by unconformities and consequently are not the products of continuous deposition. Moreover, upper Cenozoic stratigraphic units are classified in greater detail and as

smaller units than are those of older parts of the stratigraphic column (Draft North American Stratigraphic Code, 1981). Therefore, a more appropriate stratigraphic approach needs to be used such as allostratigraphy.

The North American Stratigraphic Code, 1983, states that in an allostratigraphic unit, the bounding discontinuities define and identify a mappable stratiform body of sedimentary rock. Formal allostratigraphic units may be defined to distinguish between different units: 1) superposed discontinuity-bounded deposits of similar lithology; 2) contiguous discontinuity-bounded deposits of similar lithology; 3) geographically separated discontinuity-bounded units of similar lithology; or 4) discontinuity-bounded deposits characterized by lithic heterogeneity as a single unit. Internal characteristics (physical, chemical and paleontological) may vary laterally and vertically throughout a unit. Boundaries of allostratigraphic units are laterally traceable discontinuities. A formal allostratigraphic unit must be mappable at the scale practiced in the region where the unit is defined. A type locality and type area must be designated; a composite stratotype or a type section and several reference sections are desirable. Genetic interpretation is an inappropriate basis for defining an allostratigraphic unit. However, genetic interpretation may influence the choice of its boundaries. Inferred geologic history is not used to define an allostratigraphic unit. However, well documented geologic history may influence the choice of the unit's boundaries. An allostratigraphic unit is extended from its type area by tracing the boundary discontinuities or by tracing or matching the deposits between the discontinuities (North American Stratigraphic Code, 1983).

Objective

The objective of this study was to determine the spatial and temporal attributes of a site along the lateral margin of the Flathead Lobe of the Cordilleran Ice Sheet by classic techniques of glacial geology. More specifically, this study explored these attributes for multiple advances of an ice sheet that had yet to be well defined. This was accomplished by first identifying features and collecting data, then reconstructing ice margins to match the features and data, and finally, defining a glacial history.

Aerial photograph and topographic map reconnaissance performed by Dr. William Locke and myself, combined with other previous author's work, was used to construct the hypothesis that the Ashley basin was not glaciated during the last advance of the Cordilleran Ice Sheet. It was further hypothesized that if ice had intruded into the basin, its extent was probably minimal (Fig. 4). This hypothesis is largely in agreement with work done by Alden (1953) and Richmond (1986) (Fig. 2), but contrasts with a map from Waitt and Thorson (1983) who did not define a nunatak at Ashley Lake.

Study Area

The glacial history of the Ashley basin is largely an outcome of the geographical setting of the basin, including its position, geology and climate. This will be discussed first in terms of its regional setting then of the study area itself.

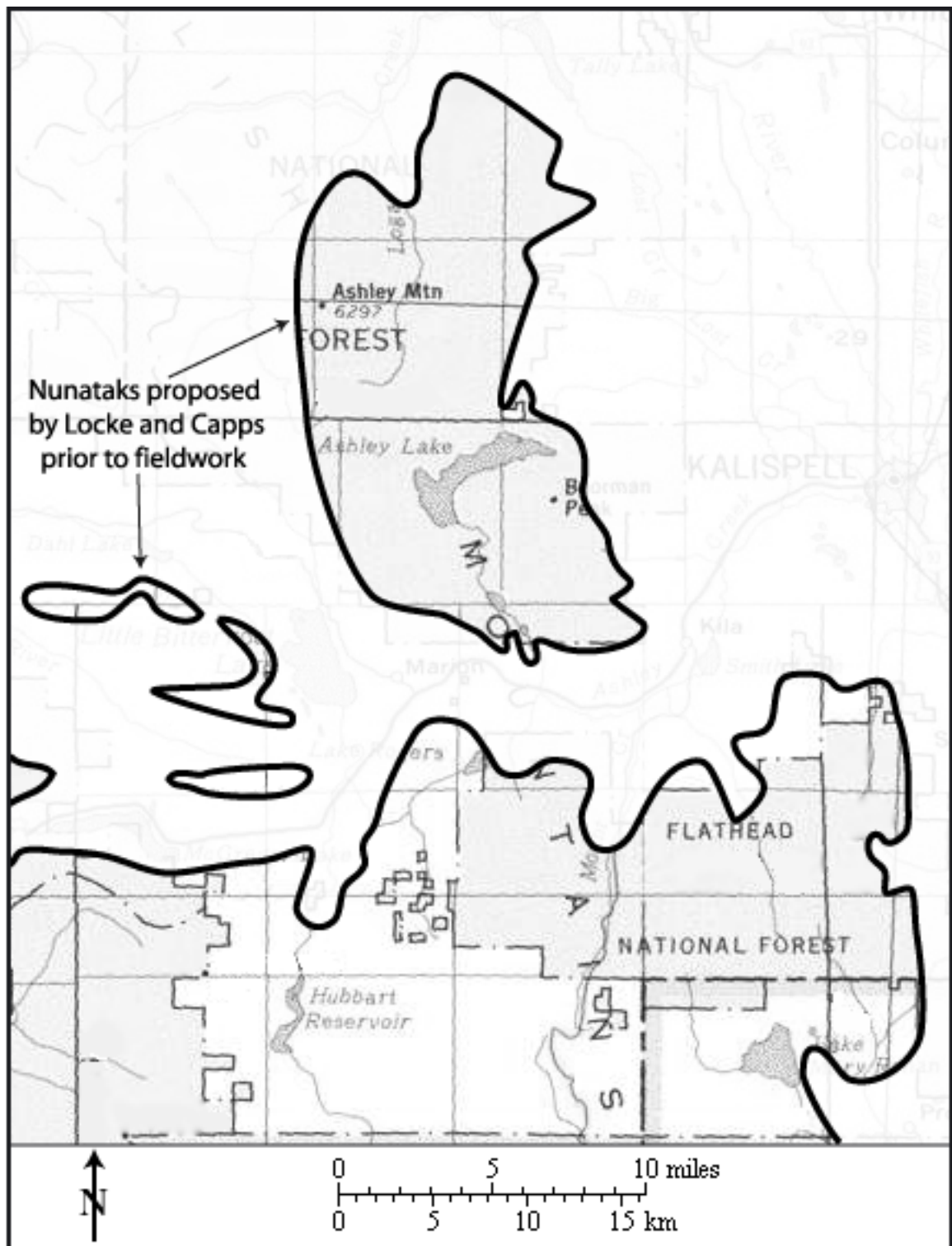


Figure 4- Locke and Capps' proposed lastglacial Cordilleran Ice Sheet margins

Regional Setting

Topography. The study area is located in the Northern Rocky Mountain Province and includes successive ranges trending north to northwest and separated by long, straight valleys that show mature drainage patterns (Fig. 5). Between the Montana-Idaho border and the Great Plains are a dozen ranges (Johns, 1970). The study area lies within the Salish Mountains whose peaks stand 1500 – 2150 m above sea level, with a total relief between peaks and adjacent valleys that seldom exceeds 700 m.

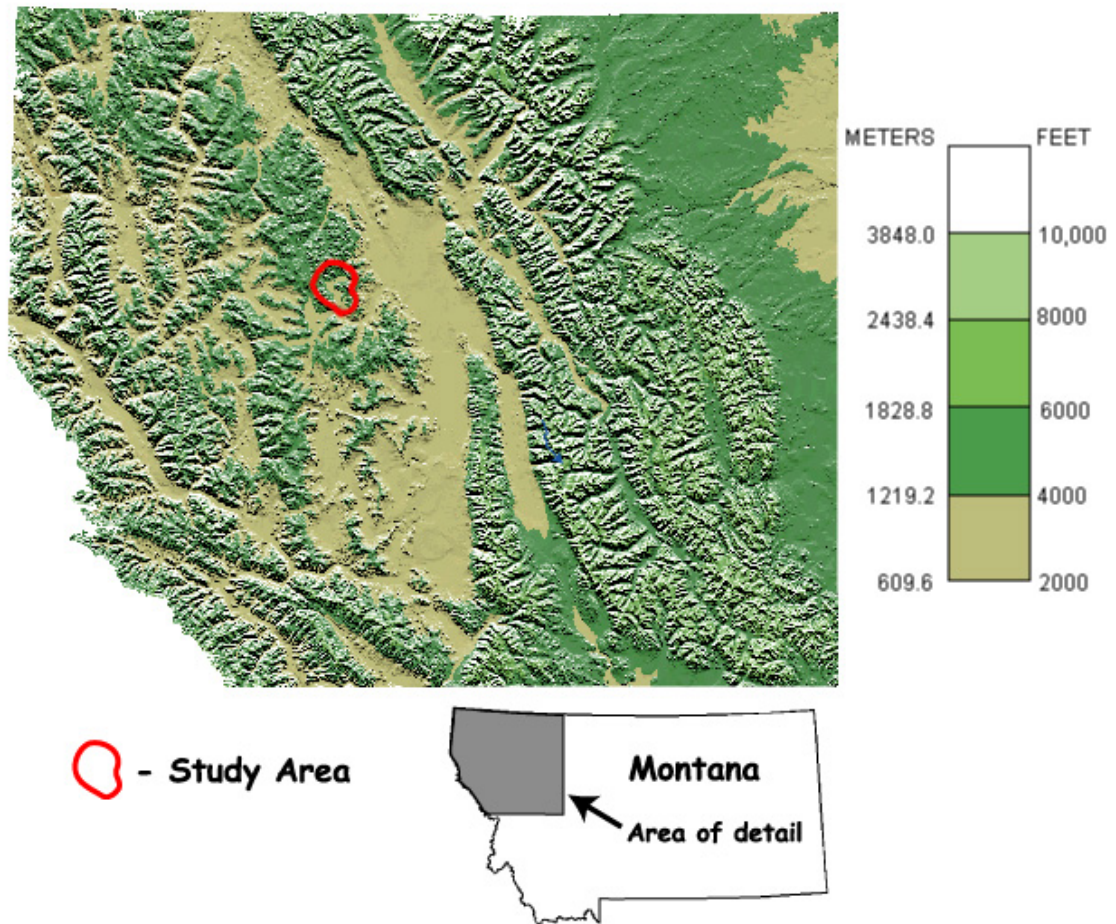


Figure 5 - Northwest Montana digital elevation model

Topography is strongly influenced by processes acting during the Quaternary. Alden (1953) noted abundant glacial moraines, outwash plains, knolls, kettles, smoothed ledges, drumlins, abandoned meltwater channels and other glacial features in the region. The basin is drained by Ashley Creek that flows to the Pacific Ocean via the Flathead, Clark Fork, and finally Columbia rivers. Topography is also strongly controlled by the bedrock geology of the region.

Geology. Harrison et al. (1992) summarized the bedrock geology and structure of the region. The region is largely underlain by Belt Supergroup rocks. The Proterozoic Belt Supergroup consists of fine-grained clastic and calcareous rocks 5200 – 12,200 m thick, which have undergone regional low-grade metamorphism. The rocks crop out in northwest trending patterns (Fig. 6). Dikes, sills and stocks ranging in composition from granitic to ultrabasic intrude Belt strata in the area surrounding the study area.

Major faults strike between north and northwest and are parallel or subparallel to folds. Major west-dipping or vertical reverse and normal faults include the Leonia, Pinkham, Mission, Swan-Whitefish, and the Flathead-Tuchuck faults. Subsidiary east and northeast striking structures displace earlier northwest faults. Tight to broad symmetrical, asymmetrical, and overturned folds plunge north or south; some are doubly plunging. Traces of some axial planes have been mapped for distances of 64 km, but most folds are much shorter.

The major physiographic features formed between 2 and 60 Ma. In places, this was concurrent with valley filling by Tertiary sediments. There was westerly extension on multiple high angle listric normal faults and backsliding on some thrust faults. Dikes intruded along some faults (Harrison, 1992).

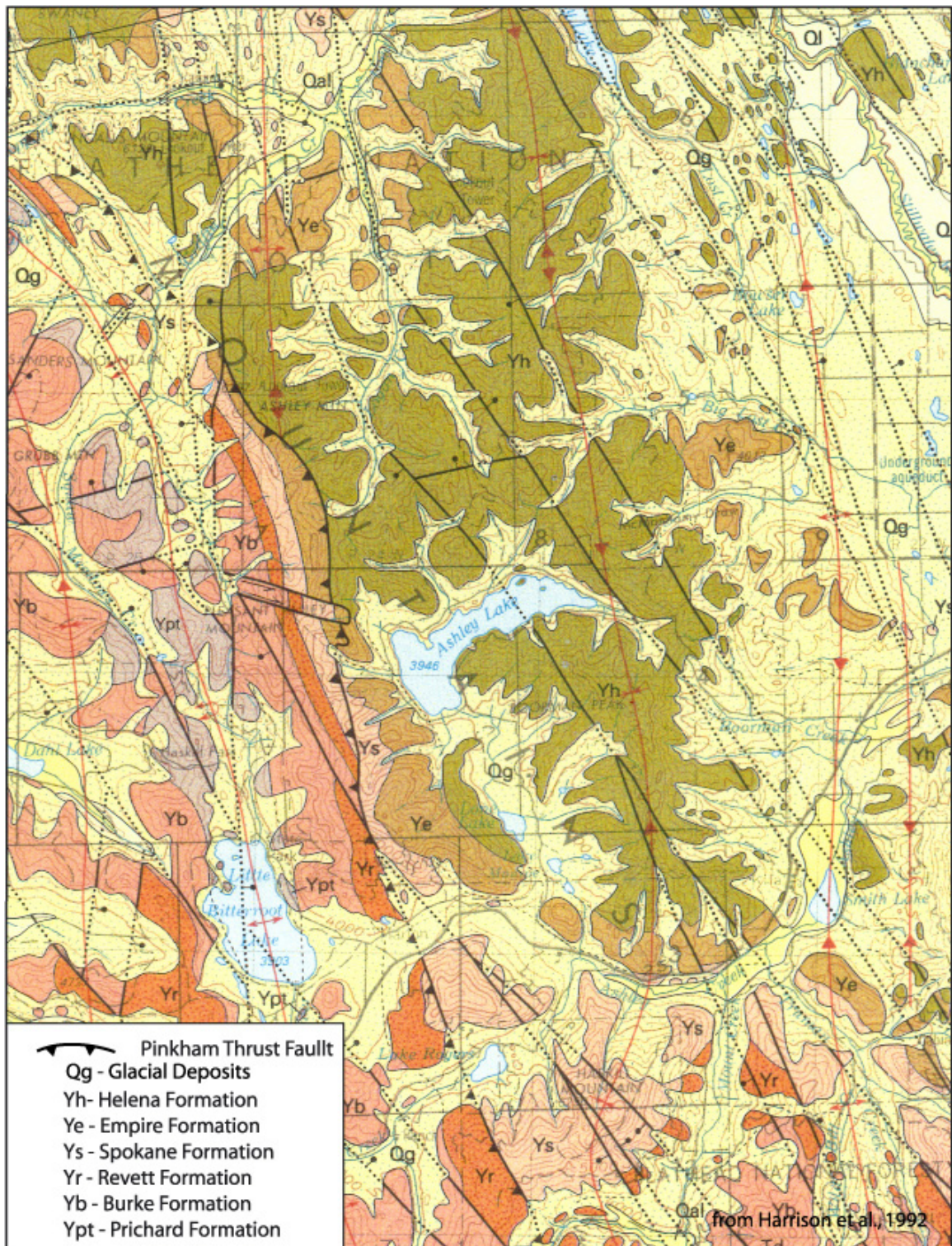


Figure 6 - Lithology and structure map of the Ashley basin

The intermontane valleys in northwestern Montana contain sedimentary strata deposited between the late Eocene and Holocene, but upper Pleistocene deposits related to glaciation and deglaciation dominate the surface exposures and valley fills (Smith et al., 2000). Much of the region has a thin surface mantle of loess that has been influenced by volcanic ash. Most of the ash came from the eruption of Mount Mazama in southwestern Oregon about 6.8 ^{14}C ka (Martinson and Basko, 1998).

During the last ice age, a lobe of the Cordilleran Ice Sheet advanced southward into the Idaho Panhandle, blocking the Clark Fork River and creating Glacial Lake Missoula. As the waters of the Clark Fork River and its tributaries rose behind this 600 m high ice dam, they flooded the valleys of western Montana. At its greatest extent, Glacial Lake Missoula stretched eastward a distance of some 300 km, creating a large lake complex of approximately 2500 km³ (Fig. 7). Periodically, as the rising waters caused failure of the ice dam, the lake emptied catastrophically causing huge floods that swept across Idaho and Washington to create the “Channeled Scablands”. Glacial Lake Missoula filled and emptied numerous times as the Cordilleran Ice Sheet continued flowing south, blocking the Clark Fork River again and again. Over thousands of years, the lake filling, dam failure, and flooding were repeated dozens of times, leaving a lasting mark on the landscape of the Northwest. Many of the distinguishing features of Glacial Lake Missoula and other glacial lakes remain throughout the region today (Ice Age Floods Study Team, 2003).

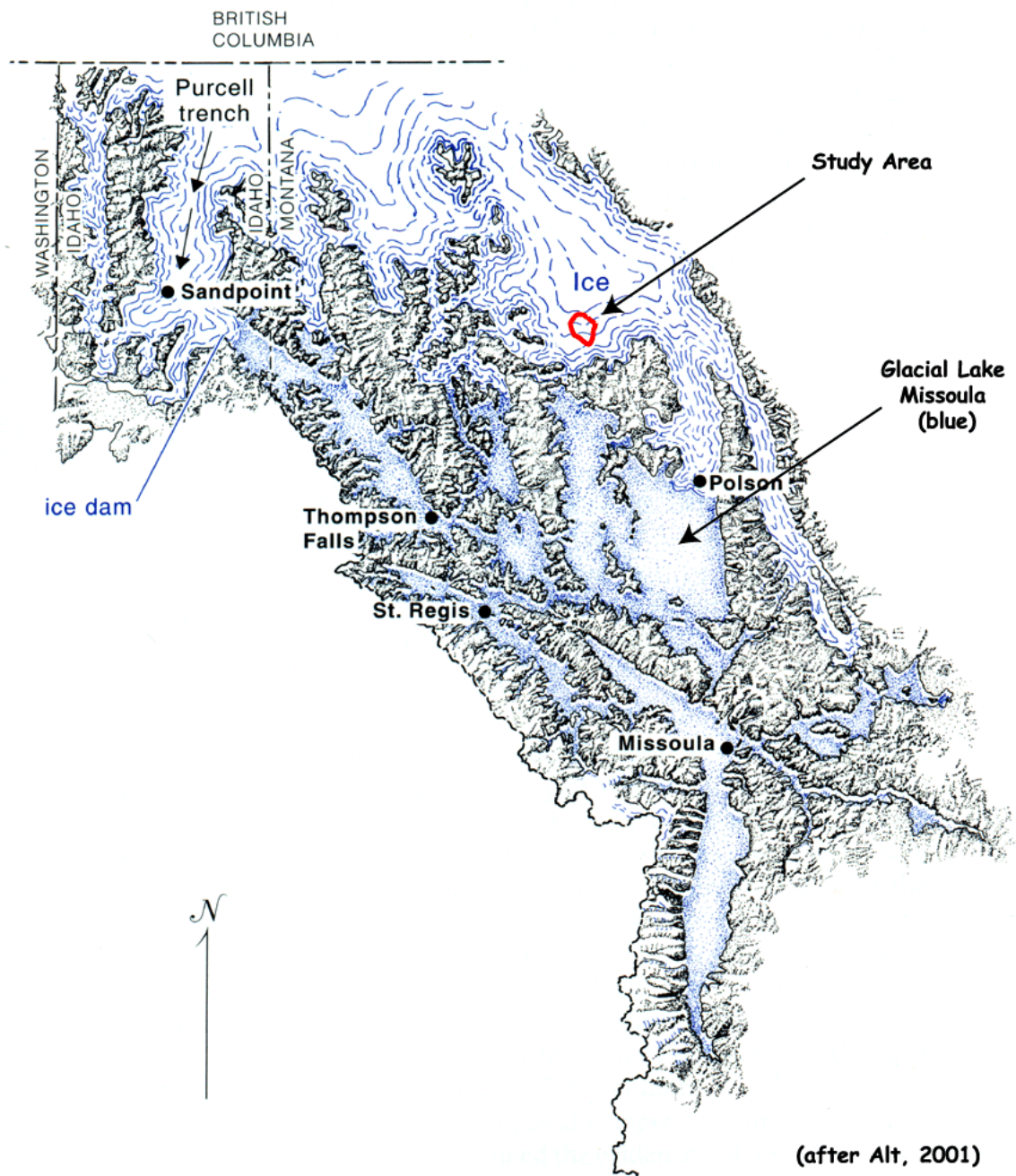


Figure 7 – Glacial Lake Missoula

Although Glacial Lake Missoula is most famous for the extensive, often spectacular erosion that it caused in the Channeled Scablands of Washington, its depositional history may be very important in this region. As the Flathead Lobe of Cordilleran Ice Sheet

receded over time, Glacial Lake Missoula periodically bordered its southern margin. At its maximum stand, the Flathead Lobe of Cordilleran Ice Sheet covered the area of Flathead Lake, the southern margin of the glacier forming the large moraine at Polson. As the glacier margin receded north over the next 2000 years or so, the succession of lakes drowned territory farther and farther north. This margin between ice to the north and Glacial Lake Missoula to the south eventually crept north of Kalispell before the lake itself finally ended (Waitt, 1999). Forty kilometers to the south of Ashley Lake, sediments deposited in Glacial Lake Missoula cover the floor of the Little Bitterroot Valley. They are thickest, between 60 and 90 meters, near Niarada at the north end of the valley (Alt, 2001). An alternative hypothesis on the origin of the lake sediments north of Polson proposes that the lake was actually ancestral Flathead Lake (Johns, 1970). This lake was dammed by the Polson moraine. Lake levels receded due to downcutting through the sediment dams, and ultimately more slowly through the bedrock sill along the Flathead River below the lake (Smith et al., 2000).

Climate. Present climate conditions within the study area are greatly affected by the Rocky Mountains. The mountains act as a barrier to the south moving arctic and central Canadian storms, while extracting moisture from, as well as restricting the eastward passage of moisture bearing prevailing westerly airflow from the Pacific coast (Johns, 1970). Climate conditions in mountainous areas are extremely variable over short distances because of local topographic effects. The Soil Survey of Flathead National Forest Area, Montana states that while Kalispell averages about 42 cm of precipitation per year, the highest mountain ridges average as much as 250 cm per year. The

mountains receive about 80 percent of their precipitation as snow. The greatest amount of rain falls during May and June; July and August are relatively dry. Temperature inversions are common in valleys during winter (Martinson and Basko, 1998).

Locke (1990) determined that the patterns of modern snowpack and paleo-equilibrium line elevations (pELAs) of the late Pleistocene in this region are parallel, with both descending to the north at twice the gradient as to the west. PELAs are elevations on former glaciers where accumulation equaled ablation. This parallelism indicates the Pleistocene pattern of airmass movements apparently was similar to that at present, which is controlled largely by regional airflow bringing cool, moist Pacific air into the northwest corner of Montana.

Not only does paleoglaciation imply a decrease in temperature and/or an increase in precipitation, but also the albedo and even the topography of the region must have differed due to ice and lake-filled valleys (Locke, 1990). There are no paleotemperature data from within the study area; however, published summaries by Peterson et al. (1979), Barry (1983), and Péwé (1983) suggest Pleistocene summer temperatures were at least 10°C lower than present in the Northern Rocky Mountain states. An interpretation of this temperature depression suggests that it has more control over paleoglaciation than changes in precipitation (Locke, 1990).

Ice. The Cordilleran Ice Sheet spread over the northern Rocky Mountains numerous times during the Pleistocene. It advanced southward from source areas in British Columbia, supplemented by local upland ice caps and alpine glaciers. There was significant input from local sources of ice from the mountains of the Swan and upper

Flathead drainages. Multiple lobes, whose flow lines tended to parallel major valleys, characterize the southern limit of the ice sheet (Richmond, 1986).

Regional lobe nomenclature has varied. The terminology used by Richmond (1986) appears to be the most descriptive, so its use has been adopted. The corresponding modern rivers drained meltwater from the Bull River, Thompson River and Flathead Lobes (Fig. 8).

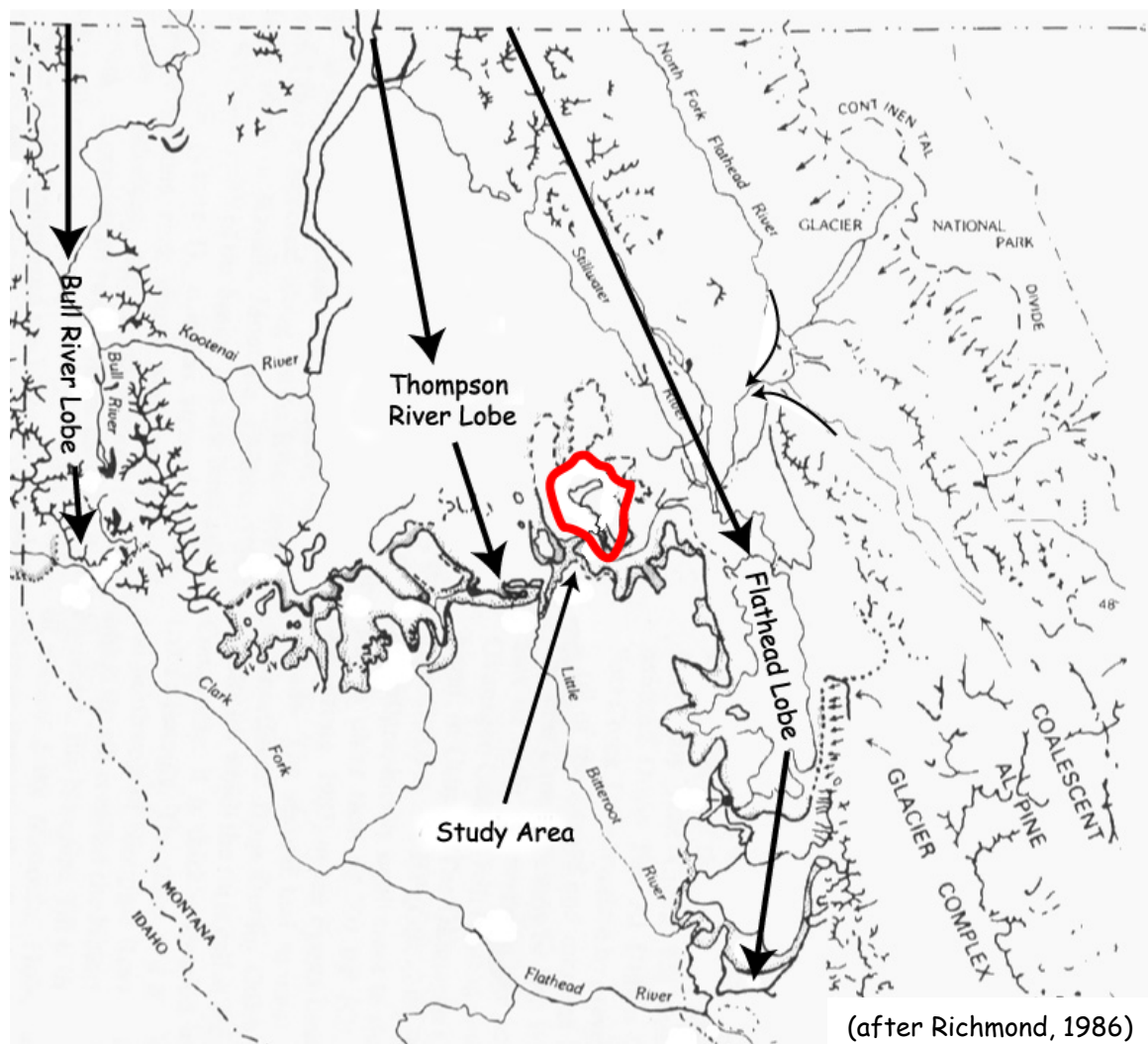


Figure 8 – Lobes of the Cordilleran Ice Sheet in Montana

The subdivision of glaciations into stades and interstades implies that stratigraphic boundaries were caused by climatic fluctuations. However, some stillstands or rapid retreats of the Cordilleran Ice Sheet may have had nonclimatic causes. Topography probably influenced advances and retreats. Richmond (1986) concluded that the lobes of the Cordilleran Ice Sheet did not all fluctuate exactly in phase with each other or with smaller mountain glaciers.

South of the limit of the lastglacial Cordilleran and Laurentide Ice Sheets in Montana, stratigraphic sections and eroded moraines give evidence of early ice sheet and alpine glaciations along the eastern Rocky Mountain front (Karlstrom, 2000). Previous glaciations thus clearly influenced topographic development in the region.

The Flathead Lobe of the Cordilleran Ice Sheet flowed south-southeast down the Rocky Mountain Trench from central British Columbia during the last glaciation. Only a few sites in this region have yielded organic material of late Pleistocene age, minimum dates of deglaciation based on radiocarbon ages are rare. Instead, much of the temporal evidence about the extensive deglaciation in this region is based on the presence of late Pleistocene tephras at various sites north of the glacial maximum. Lobes of the Cordilleran Ice Sheet are known to have receded at least tens of kilometers prior to the deposition of the Glacier Peak G (11.2 ^{14}C ka) and Mount St. Helens set Jy (11.4 ^{14}C ka) tephras. The active margin of the Flathead Lobe was at least 80 km north of its terminal position when the Glacier Peak G ash was deposited (Smith et al., 2000). Flathead Lobe termini for various glaciations are still a major source of debate. Most recently, Levish (1997) interpreted the lastglacial terminus to be at the Polson Moraine. Richmond (1986) details end/terminal deposits of older, more extensive glaciations; however, unpublished

data from subsequent work suggests that at least some of these proposed moraines are of nonglacial and much older origin.

Vegetation. Vegetation is important to this study due to the direct and indirect effects that it has on the morphology of the land and the effects that it may have on the proposed methods of the project. The Soil Survey of Flathead County, Montana summarizes the vegetation of the region. Most of the region is forested or has the potential to be forested. Important tree species are ponderosa pine, Douglas fir, western larch, grand fir, western white pine, lodgepole pine, western redcedar, subalpine fir, and Engelmann spruce. Whitebark pine and alpine larch are important species at the highest elevations. Small sedge meadows are widely distributed in the major valleys throughout the region (Martinson and Basko, 1998).

Natural resource extraction and agriculture has historically been a major driver of the economy in this region. The forest products industry has been active here for over a century. Subsequently, the forest products industry's practices are significantly influencing the current stands of timber. Crops, hay and grazing activities have caused the modification of valley vegetation.

Local setting

Topography. Moderate-relief mountains cover most of the study area. The Ashley basin has one low gradient valley and several tributary valleys (Fig. 3). The mountains are mostly 20-40% slope with some as high as 60%. The highest mountain in the study area, Pleasant Mountain, is located on the divide in the northwest corner of the basin at an

elevation of 1899 m. Ashley Lake itself lies at 1203 m. The lowest elevation is 997 m at the confluence of Ashley Creek with Idaho Creek at the southern terminus of the basin. Total relief in the study area is 902 m. Locally, 25 m vertical cliffs are present due to glaciofluvial erosion. The basin is drained by Ashley Creek, which flows into the Flathead River just north of Flathead Lake.

Geology. Harrison et al. (1992) summarized lithologic and structural information of the study area. The entire bedrock section of the study area is composed of Belt Supergroup rocks, here Middle Proterozoic in age. The Helena Formation floors most of the study area; an exception is the western ridge, which is composed of the Empire, Spokane, Revett and Burke Formations (Fig. 6).

The Helena formation has hundreds of distinct lithologic cycles in each unit. Each complete cycle consists of three parts: 1) an upper bed of dense conchoidal-fracturing dolostone that weathers orange-brown, 2) a middle bed of gray dolomitic siltite that commonly displays horizontal calcite pods or blobs that grade successively upward to vertical pods and blobs and then into molar tooth structures, and 3) a lower clastic bed rarely more than a foot thick of white quartzite or quartzite and black argillite resting on a cut surface. Complete cycles are seen in many exposures, but most commonly the lower bed is missing and less commonly one of the other beds is missing. Pyrite and chalcopyrite are common. Stromatolite beds are present at some localities. The thickness of the formation ranges from zero to about 2750 m.

The Empire formation is a thinly laminated, dark-green and light-green dolomitic argillite and silty argillite or siltite. Fluid-escape structures are characteristic and

horizontal pods of white or pink calcite are particularly abundant in the upper part. Ripple marks, desiccation cracks, and mud chips are common in places. The lower part contains white dolomitic quartzite beds as thick as 60 cm. Pyrite cubes are common in more carbonate-rich strata. A few purple inter-laminated argillite and siltite beds commonly occur near the base, and at places near the middle of the unit. The thickness of the formation ranges from zero to about 600 m.

Most of the Spokane formation has three lithologic members. The upper member has beds of green laminated argillite and siltite that alternate with beds of purple laminated argillite and siltite. Dolomitic cement is common as are ripple marks, mud chips, desiccation cracks, fluid-escape structures, ball-and-pillow structures, and flute casts. The middle member is predominantly pink to purple-gray, very fine-grained feldspathic quartzite or coarse siltite that has planar lamination or long tubular cross lamination. Interbeds of purple argillite are common. The lower member is similar to the upper member but has more purple argillitic beds, is more dolomitic, and has scattered iron-carbonate specks and cement. The thickness of the formation ranges from zero to 1500 m.

The Revett formation is characterized by blocky sets of white, green, or pale purple-gray, crossbedded quartzite beds. In most areas, the Revett can be divided into three informal members: upper and lower quartzite-rich members and a middle member that contains significantly more argillite and siltite. Many beds display climbing ripple marks, ripple cross laminae, load casts, and 0.3 – 1.5 m deep channels. Heavy minerals are common on cross laminae. Argillite beds are purple or green, generally display even- to wavy-parallel laminae, and contain desiccation cracks, mud chips, ripple marks and fluid-

escape structures. Small flakes of secondary biotite and tiny euhedral magnetite crystals are displayed in the formation throughout the study area. The thickness of the formation ranges from about 600 to about 850 m. The Revett interlayers with the underlying Burke formation.

The Burke formation is divisible into three informal members in most areas. The upper member consists predominantly of blocky beds of purple-gray interlaminated argillite and siltite interbedded with greenish-gray interlaminated argillite and siltite. Desiccation cracks and mud chips are common in the upper part of the member. White to purple-gray, parallel-laminated siltite and quartzite beds are interbedded in the upper part of the member and increase in abundance upward. The middle member is predominantly gray to purple-gray, parallel- and thinly-laminated siltite in blocky planar beds that have minor argillite partings or beds. The lower member is predominantly green parallel-laminated, interbedded argillite and siltite; these generally are in beds about 5-50 cm thick and weather to give a blocky-flaggy outcrop. Siltite and some argillite lithologies in all members are commonly speckled by tiny euhedral magnetite crystals and display small flakes of secondary biotite. The thickness of the formation ranges from about 750 to 1100 m.

The major structures of the study area are a thrust fault, two normal faults and an anticline (Fig. 6). The Pinkham thrust fault trends north-northwest across the western ridge of the study area and overrides several older, broad open folds of the Purcell anticlinorium. Other folds have been piggybacked on the thrust plate; the thrust plate has a clockwise rotation and is hinged to the south where the thrust fault ends in gently folded terrane. This flat thrust fault cuts older folds of the Purcell anticlinorium, and at places

cuts down section in the direction of tectonic transport. Two unnamed normal faults, one with an offshoot in the south, parallel each other and trend northwest across the eastern and northern parts of the study area. The axis of the Purcell Anticlinorium, which is to the west of the study area, trends north-northwest (Harrison et al., 1992).

The surficial geology includes moraines, lake sediments and local organic deposits (Martinson and Basko, 1998). These sediments extend up to about 1280 m in elevation and floor the main and connecting valleys (Harrison et al., 1992). They vary in thickness from zero to at least 100 m in the center of the main Ashley Valley as is shown by a water well log (Montana Bureau of Mines and Geology, 2003).

It is unclear if Glacial Lake Missoula ever occupied the Ashley basin. While the maximum extent of the lake at 1280 m above sea level was certainly high enough to flood the basin, ice may have occupied it during this highstand of the lake.

Climate. The mean annual temperature (MAT) at Kalispell, 20 km east and ~300 m lower than Ashley Lake, is 6.0°C (Johns, 1970). Applying an average environmental lapse rate of 5.9°C/km (Locke, 1989), the MAT at Ashley Lake, elevation 1203 m, is 4°C and the very top of the highest point in the study area, Pleasant Mountain, almost reaches the 0°C isotherm.

Because the patterns of modern snowpack and paleo-equilibrium line elevations of the late Pleistocene are parallel in the regional context, it is expected that a similar relationship would exist locally. The presence or absence of the continental ice mass is an important factor on the local late Pleistocene climate. Using the 10°C temperature depression and lapse rates that were applied to the regional climate, the temperature at

Ashley Lake during the late Pleistocene would have been -6°C . Precipitation was probably not significantly different from that of today. Depending on the variable airflow patterns of the region it could have been slightly wetter, or more likely, slightly drier (Locke, 1989).

Ice. The central Salish Mountains do not appear to have supported alpine glaciers. This statement is supported by the lack of cirques and reconstructions of regional pELAs (Locke, 1990), although the highest mountains were certainly close to the glaciation limit. Other slightly higher mountains in the Salish Mountains have obvious cirques. Therefore, ELAs in the study area during the last glacial maximum are interpreted to be approximately 1800-1900 m.

If the area compares with alpine glacial extents in other regions of the U.S., the margins of a previous glaciation will be slightly more extensive than the margins of the last glaciation (Richmond and Murphy, 1965; Colman and Pierce, 1986). However, there are few confirmed deposits of a previous glaciation noted in the area (Alden, 1953; Richmond, 1986).

The local extent of the Cordilleran Ice Sheet in the study area has been a matter of debate among workers. Very limited fieldwork had been done in the study area previously and the work that had been done was of reconnaissance nature.

Vegetation. Locally the vegetation is primarily controlled by topography and secondarily by water availability. Thick stands of second-growth evergreens cover the mountain slopes and ridges, but both evergreens and deciduous trees grow in the valleys (Johns, 1970). Currently, the valley bottoms are hummocky in places but are mostly

meadow that is used in production of hay. Aspen tends to grow where the ground is wet and evergreens grow on the better drained higher ground. Bogs are common and support large colonies of cattails and tall grasses.

Extensive logging has taken place in the study area on land owned by the National Forest Service, Plum Creek Timber Company, Stoltz Timber Company and many private landowners. Logging has both simplified and complicated research. For example, clearcuts allow the morphology of the land to be seen more clearly. Abundant logging roads supply access and road cuts expose underlying materials. On the other hand, logging has disturbed most areas to some degree. As a result, features that appear glacial may in fact be due to past logging practices (roads, ditches, materials no longer *in situ*, *etc*). Also, many, if not all, striations have to be questioned as to their origin because they may have been caused by heavy equipment. Even some of the steepest slopes showed signs of past logging.

METHODS

The Ashley basin allowed the application of several classic techniques of glacial geology to determine the spatial and temporal attributes of the multiple advances of the Cordilleran Ice Sheet. The Quaternary history was determined by: 1) distribution of erratic boulders; 2) location of glacial depositional landforms; 3) location of glacial erosional landforms; 4) relative weathering rind thickness; 5) tephrochronology; and 6) radiocarbon dating.

Erratics

The distribution of erratic boulders was used to determine glacial ice extents. Erratics are rock fragments carried by glacial ice, deposited at some distance from their original outcrop and generally resting on bedrock of different lithology (Bates and Jackson, 1984). Erratics show minimum levels of glaciation; this is a result of their being transported by ice from locations with different bedrock lithologies, and they are rounded by abrasion within the glacier.

Ridges were walked throughout the study area to attempt to find evidence of erratics (Fig. 9). Ridges are chosen because they have an absence of fluvial activity and the only other normal transportation mechanism that exists is gravity, which could only bring clasts down, not up. Erratics are distinguished from country rock by their degree of roundness, faceting and exotic lithology. First, only clasts that are sub-rounded to well rounded were considered. Rounding shows that the clast has undergone abrasion and therefore transportation. Those that fit this criterion were evaluated to determine if they

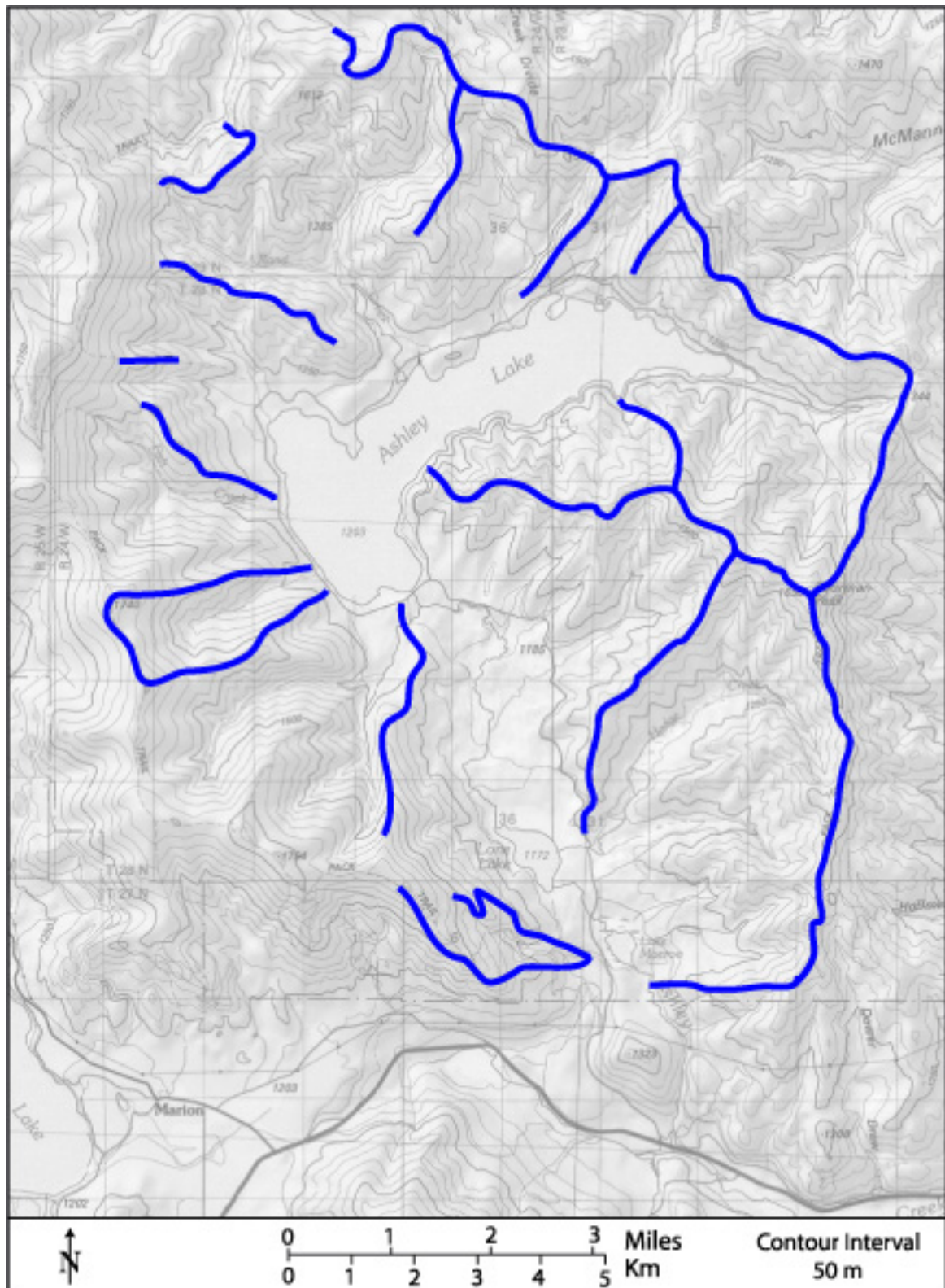


Figure 9 – Ridges walked

were of exotic lithology based on the known geology of the study area and through direct observation of the bedrock. This two-pronged approach was used because classifying a surface rock as an erratic solely based on “exotic” lithology could lead to error due to the extreme heterogeneity of Belt Supergroup rocks. Therefore, the probability of a rounded, exotic clast occurring on a ridge by a process other than glaciation was minimal.

When a suspected erratic was found, its B-axis was measured and roundness, faceting and any surface feature associated with the erratic was noted. The lithology was determined with hand lens and dilute HCl. The position of confirmed erratics was determined by GPS and mapped using digital topographic map software. The highest occurrence of erratics throughout the study area describes an upper limit of past glaciations.

Erratics can also be used as a relative dating criterion. Erratics from different glaciations should show different degrees of weathering. Resistant rocks that are high in silica will tend to show less weathering, while those with high percentages of carbonate will weather relatively quickly.

Depositional landforms

Topographic map and aerial photo interpretation was used to determine probable locations of ice marginal landforms. These landforms were then verified and mapped in the field to determine the extents of advances and major recessional stillstands of the Cordilleran Ice Sheet.

Moraines

Moraines are mounds or ridges of unstratified glacial drift, chiefly till, deposited by direct action of glacial ice (Bates and Jackson, 1984). Moraines are typically recognized by their arcuate shape, hummocky surface and ridge-like profile. When moraines are present in valleys, they often cause closed depressions up-valley and are incised by streams.

Moraine morphology is typically not well developed at continental ice sheet margins and is therefore a poor indicator of age. However, if sub-lobes of ice spilled into the basin over ridges they would effectively be outlet glaciers. This ice may derive some debris from adjacent slopes, making it more likely to display useful morphology. Additionally, identifiable moraine morphology was used to determine the direction ice movement.

Kames

Kames are variably shaped mounds composed chiefly of sand and gravel, formed by supraglacial or ice-contact glaciofluvial deposition (after Benn and Evans, 1998). Because kames are “ice-contact”, they can be used to reconstruct glacial margins.

Glacial depressions

Closed depressions are both obvious and abundant on topographic maps and aerial photos of the study area. Closed depressions have a wide variety of causes. Even within the realm of glacial geology, they can be formed in several ways. They were inspected for general characteristics such as size, shape, nature of vegetation, and water saturation.

In addition, closed depressions that contained bogs were augered in an attempt to recover tephra and organic material for radiocarbon dating.

Erosional Landforms

Erosional landforms are both obvious and abundant on topographic maps and aerial photos of the study area. All workers in the region have noted this. The most abundant erosional landforms in the study area were expected to be glaciofluvial in origin. Other erosional landforms such as U-shaped valleys, trimlines and cirques were investigated as well.

Erosional landforms can also be used to relatively date the processes that formed them. The difference in time between the maxima of the last two glaciations is sufficient to allow significant modification of the features by weathering and differential erosional processes. Erosional landforms from previous glaciations will typically be less distinct than those from the last.

Lateral Meltwater Channels

Lateral meltwater channels, also known as ice marginal channels, are channels cut by meltwater flowing between the ice margin and the valley wall while the valley below was filled with ice (Fig.10). Lateral meltwater channels occur singly or in a series diagonally across the slope of the hillside. These are generally flat floored or v-shaped channels and have a wide range of channel dimensions. They are often floored with slumped till or lag accumulations of boulders and gravel. In this way, the history of the retreat of the ice

margin is indicated by a parallel series of channels that trace the change in the margin of the ice lobe (Fulton, 1967).

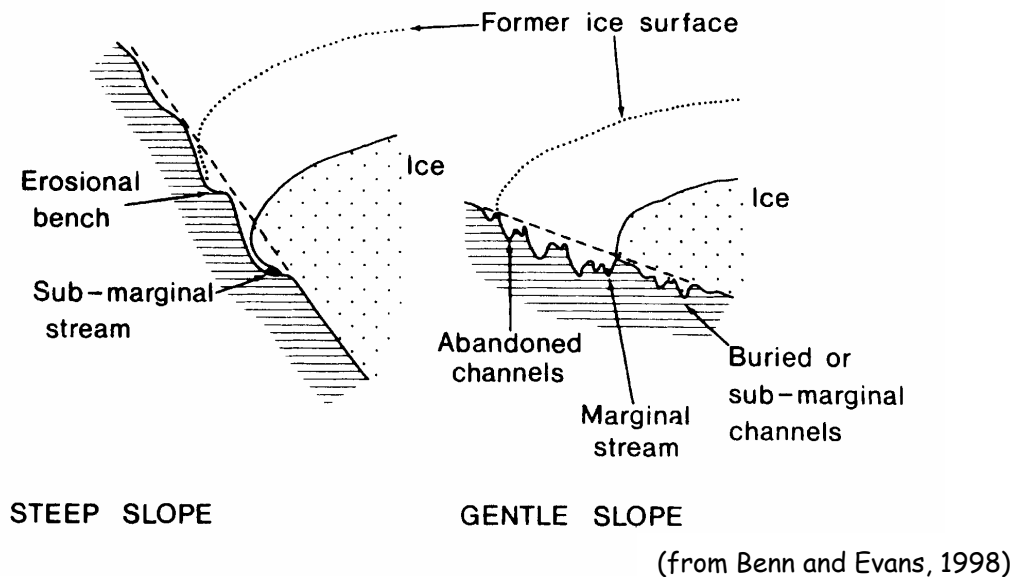


Figure 10 – Lateral meltwater channel formation

The existence of lateral meltwater channels was interpreted from topographic maps and aerial photos and by ground observations. They can be difficult to separate genetically from direct overflow channels that will be discussed next because these forms grade into one another.

Direct Overflow Channels

Direct overflow channels are “wind gaps” of glacial origin. Wind gaps are a notch in the crest or upper part of a mountain ridge or a pass that was formed by a stream, although no stream or a much smaller one flows there now (after Bates and Jackson, 1984). Direct overflow channels are formed where a glacier blocks the normal drainage causing meltwater to pond until it can flow through a low divide away from the ice front

(Fulton, 1967). Because the ice/ground-surface interface must be at least as high as the water, these “spillways” can be used to reconstruct minimal glacial ice elevations.

Ridges in the study area were inspected for direct overflow channels. Topographic maps and aerial photos showed the probable existence of several of these features. It was expected that more would be found during fieldwork; this is primarily due to issues of scale. The study area currently supports large trees that can easily conceal smaller features. Because topographic maps are primarily made from aerial photos, it follows that the maps may not show them either. In some cases, even if there were no vegetative cover, the features would be too small to be mapped at the scale of 1:24,000. Direct overflow channels were located by GPS and the probable direction of water flow determined.

Weathering Rinds

This study of rinds, and those done by many other workers, relies heavily upon detailed work done by Colman and Pierce (1981 and 1986). A weathering rind is a zone of oxidation colors whose inner boundary approximately parallels the outer surface of a clast. Other weathering reactions commonly accompany oxidation, but the rinds are defined primarily by visible discoloration. Weathering rinds begin to develop on clasts in the zone of weathering shortly after clast deposition, so that the duration of weathering represented by the rind is essentially the same as the age of the deposit.

Weathering rinds were employed in an attempt to determine how to subdivide glacial limits. Statistical analysis demonstrates that weathering rinds can be an excellent quantitative indicator of age. Within a sampling area, deposit age is the most important

source of variation in rind thickness, and all differences in mean rind thickness between deposits of different stratigraphic ages are important. Rind thickness data for independently dated deposits from elsewhere demonstrate that the rate of increase in weathering rind thickness decreases with time. Because the rate decreases with time, the ratio of the rind thicknesses of two deposits in the same area provides a minimum estimate of the numerical age ratio of the two deposits. A logarithmic time function was to be calibrated for the sampling area if merited by the data.

A large number of variables or factors potentially affect weathering rind thickness. These factors can be classed into two general groups, here called sampling factors and environmental factors. Elements of the sampling procedure that are capable of introducing variation into the measurement of weathering rinds include: 1) selection of sampling sites; 2) collection of samples; 3) procedures for measuring the selected samples; and 4) the operator who performs the sampling and measuring. Possible variation in rind thickness introduced by these factors was largely eliminated from this study by using a standard set of sampling and measuring procedures. The effect of different operators is negated by personally measuring all rinds. Environmental factors that affect weathering rind thickness, like soil forming factors, are climate, parent material, vegetation, topography, and time. Topography includes the effects of erosion and deposition; climate includes temperature, precipitation, and their seasonal distribution; parent material includes rock type, rock texture, and soil matrix texture. The effects of climate, namely temperature, precipitation and their seasonal distribution, were minimized by sampling over a restricted area. Land surface stability was the primary criterion for selecting sampling sites, erosional or depositional disturbance of the

weathering profile was to be avoided. Sampling sites were located on relatively flat moraine crests or kame surfaces. This type of site precludes burial by colluvium.

The sampling design used in this study was conceived primarily to evaluate the relation between weathering rind thickness and time (age). Sampling methods, measurement procedures, and statistical analyses for weathering rind data in this study are discussed in detail by Colman and Pierce (1981). Several options existed in procedures for selecting the clasts at each sampling site; these options included the lithology of clasts selected and the position in the weathering profile from which the samples were taken (Colman and Pierce, 1981, 1986). Study sites were selected in which there were minimal differences in non-temporal weathering and soil forming factors; this increased the probability that variations in rind data reflect different durations of weathering (Burke and Birkeland, 1979).

All of the deposits sampled were either glacial or glaciofluvial in origin. Data collection concentrated in these areas because they contained well developed sequences of deposits of different ages, because they contained abundant, relatively uniform rock types, and because they had climates conducive to rind formation.

Sites were not chosen by formal random selection procedures from all possible sites of the above description because of the limited number of sites available within the study area (Fig. 11). However, there should be no bias in the selection of study sites because all identified sites were studied and the selected sites are thought to be representative.

Weathering rinds were measured on clasts collected from excavated pits. Clasts were

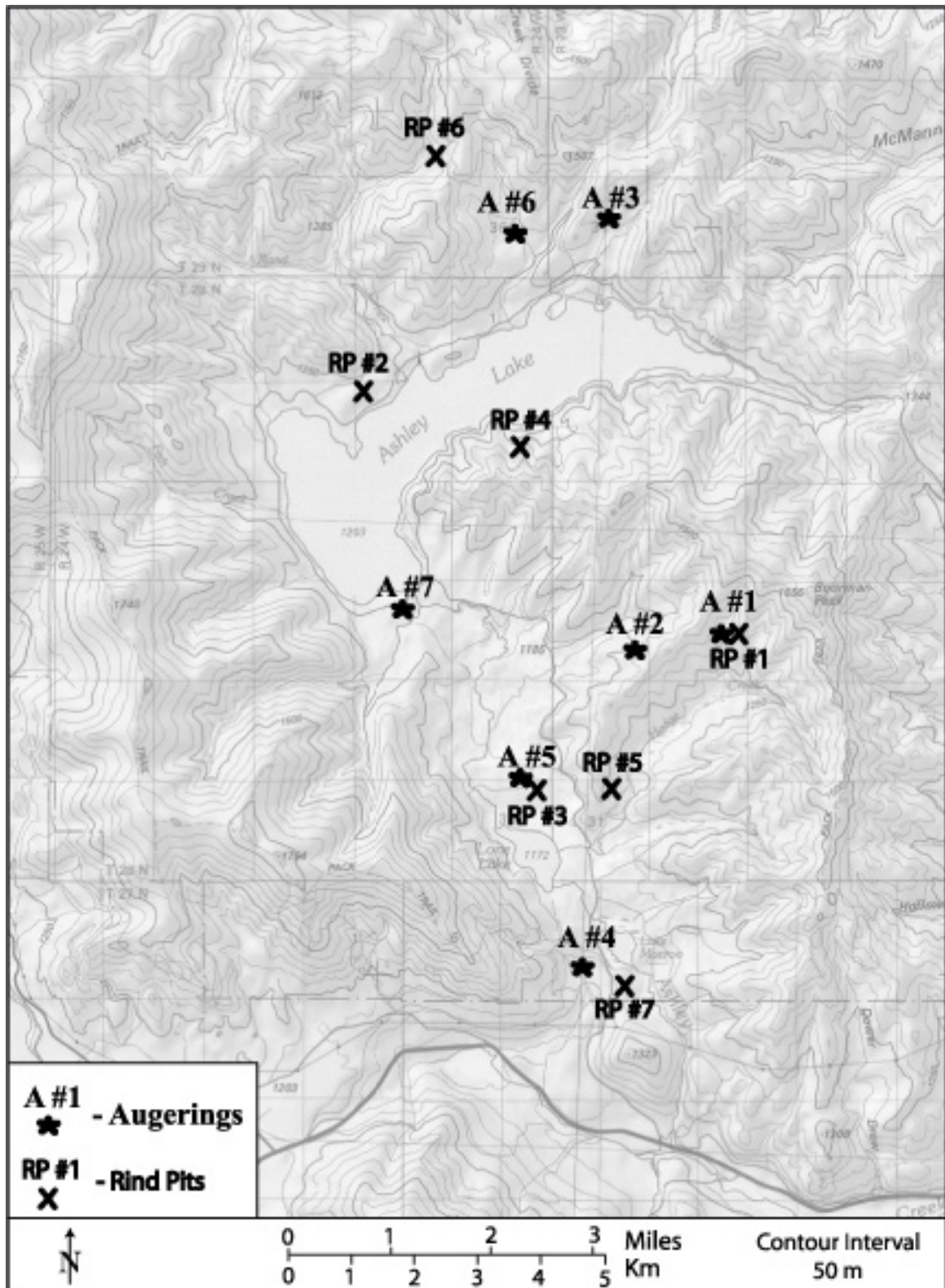


Figure 11 – Weathering rind pits and augering sites

sampled at depths of 20 to 50 cm in the upper parts of the B-horizon of soils developed in the deposits. Weathering rinds are thickest in the upper parts of B-horizons and decrease in thickness with depth (Colman and Pierce, 1981). At least 30 rinds were measured to the nearest 0.1 mm at each sampling site using a seven-power magnifying comparator.

Bog Augering

Bogs were augered in an attempt to recover dateable carbon and tephras to establish a minimum date for deglaciation (Fig. 11). Bogs were mapped, including size, shape, nature of the dam, vegetation, and saturation of sediments. Sampling was done using a 10 cm diameter, double-bitted hand auger. Cut materials are forced into the barrel where they can be recovered. Samples were extruded with a plunger onto a plastic tarp in order of its removal and split lengthwise to allow inspection.

Bogs were selected to maximize the temporal range between each other and to maximize the time that any one bog records in its deposits. By choosing the highest and lowest bogs in the study area and a continuum between them, we hoped to be able to date significant stillstands of the deglaciation of the study area and therefore, the region. To maximize the time that a bog records, they were selected for geometry and significant size. The primary concern when choosing the augering site within any given bog is to get as deep a core as possible. This was accomplished by estimating the deepest point by considering the slopes surrounding the bog. Other factors that were necessary to account for included vegetation, surface water incision and standing water. Depth interval, moist Munsell color, grain size, moisture and bedding characteristics were recorded for each augering. Augerings proceeded until it was no longer possible to continue due the

inability of the auger to penetrate the substrate. Large clasts, deposited by glacial processes, probably inhibited further augering.

Tephra

Tephra studies are based on work summarized by Sarna-Wojcicki (2000). Tephra are volcanic ash deposits that may cover large areas, and consequently are useful for chronostratigraphic correlations. Tephrochronology becomes a particularly powerful chronostratigraphic technique when it is combined with other dating and correlation methods (i.e. isotopic dating techniques). The ability to recognize tephra from specific eruptions in the stratigraphic record enables correlation of stratigraphic sequences over wide areas and in many cases allows assignment of numerical age.

Tephra were sought in the cores of glacial bogs and exposures incised into glacial deposits. Proper identification of a tephra layer began with fieldwork and progressed through various laboratory analyses until the characteristics of the tephra were confidently correlated to those of a known tephra. Rarely, the correlation can be accomplished in the field. Usually some laboratory analyses are needed to identify a tephra layer with some degree of confidence, and detailed mineralogy is needed to identify a tephra positively. Once a potential tephra was located, material was selected that appeared the least contaminated with other sediments. As much material as possible was recovered up to one liter because this is generally an adequate amount for most chemical and petrographic analyses if abundant, fresh glass is present in the layer. In distal airfall tephra layers, the coarsest, purest and densest particles usually will be present near the base of a unit. Properties that were identified in the field include

stratigraphic context, site location, physical continuity, thickness, bedding characteristics, grain size and grading, other aspects of texture, color, grittiness, and reflectivity.

Depending on how diagnostic the field characteristics are of found tephra, several types of laboratory analyses may be performed. The strategy was to begin with the simpler, less expensive procedures and proceed to the more complex, time consuming, or expensive procedures. Petrographic examination was conducted first. The sample was examined using a polarizing petrographic microscope with plane-polarized light and crossed-polarized light capability. The presence and abundance of glass shards was noted and the morphology of the shards described. The degree of hydration of the glass was also noted. In the laboratory, tephra samples were prepped according to methods described in detail by Sarna-Wojcicki (2000). These preparations ease identification. Confidence in correlations is obtained from the convergence of strong similarity of several different sets of physical, chemical, stratigraphic, and chronologic data rather than from any single quantitative or statistical measure (Sarna-Wojcicki, 2000).

Tephra were used to date the stratigraphic sequence by comparing the deposits found in the field with the properties of tephra known to exist regionally (Fig. 12). Tephra layers that were likely to be found at this site included Glacier Peak G (11.2 ^{14}C ka), Mazama (6.8 ^{14}C ka) and Mt. St. Helens D (1980 A.D.), T (0.180 ^{14}C ka) and Jy (11.4 ^{14}C ka) (Carrara, 1989).

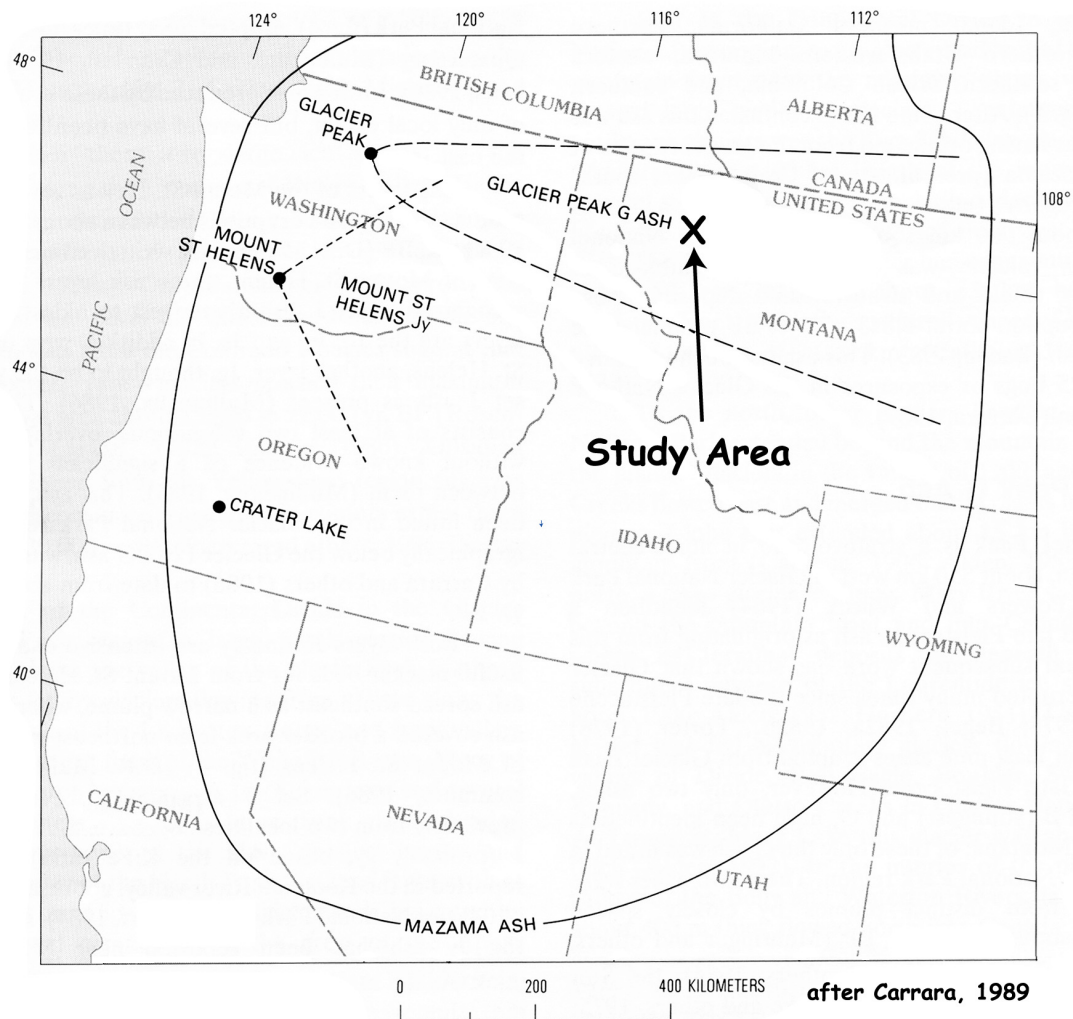


Figure 12 - Distribution of the Mazama, Glacier Peak G, and Mount St. Helens Jy ashes

Radiocarbon

Carefully assessed minimum dates established with radiocarbon can be used to develop an accurate glacial chronostratigraphy. Radiocarbon was sampled primarily from bog bottoms to obtain a minimum date for deglaciation. When possible, age control to calibrate the relative weathering techniques was to be provided by ^{14}C dating of organic matter recovered from bogs adjacent to pits dug for weathering rind analysis.

Techniques explained in detail by Gillespie (1984) were employed. Extreme caution was exercised in sampling in order to avoid contamination. Contamination of the sample can occur at all steps of the process. Material suitable for radiocarbon analysis was sought by splitting the cores augered from bogs lengthwise and doing a visual search. More detailed searches were undertaken in the laboratory by concentration techniques described by Kihl (1975). The organic material dated is likely to be similar in age to the deposit around it, thereby indirectly dating the deposition of the inorganic material.

Ice modeling

An “Excel” based spreadsheet model, GlacPro (Locke, 1995) was used to reconstruct centerline elevations along major flowlines (Fig. 13). Ice terminus positions were defined by major moraine systems identified by the previously proposed methods. Within the mountains the only constraints are the elevations of breached divides and nunataks, both interpreted from topographic maps based on presence or lack of obvious glacial scouring and shaping, and ground observations such as presence of erratics and striations (Locke, 1999).

Because glacier ice is a relatively well characterized material, its behavior is generally predictable. As recognized by Nye (1952 a and b) and applied by others in the intervening years, this predictable rheology can be used to reconstruct former glaciers. Only minimal field data are required, specifically, a defined starting point and valley geometry. Locke (1999) has summarized the use of spreadsheets in glacial modeling. Alternatively, if the vertical extents of former glaciers can be reconstructed on

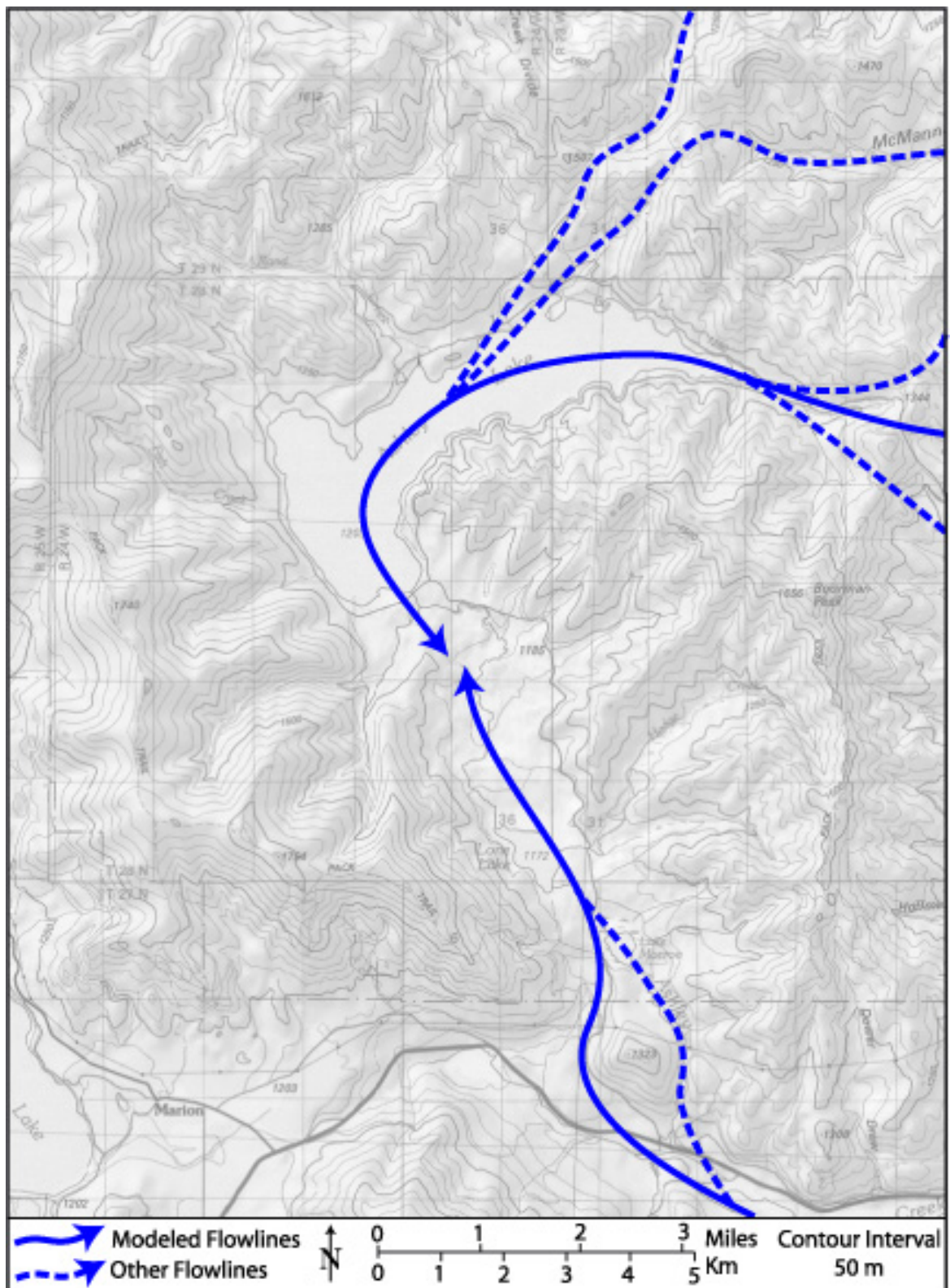


Figure 13 – Major ice flowlines

the basis of field evidence (moraine crests, trimlines, nunataks, *etc.*), flow models can be used to interpret variability in the factors which control ice flow, especially effective basal shear stress. These methods involve the application of iterative models to former ice cover of the study area (Locke, 1999).

RESULTS

Erratics

Erratics were widespread throughout most of the study area (Fig. 14). Many more observations were made than are shown on the map due to issues of scale. Erratics were easy to identify. The bedrock of the area tends to weather into blocky clasts because it has well developed bedding planes and cleavage. As a result, clasts that are sub-rounded to well rounded stand out in stark contrast to the country rock. In most cases, the sub-rounded to rounded clasts are obviously exotic in lithology. Sub-rounded to rounded, apple green argillites are a quick, easy, and common erratic because this rock was not observed as bedrock anywhere in the study area. Very well rounded, extremely hard, diabase exotics, while somewhat rare, were immediately recognizable as erratics because their lithology contrasted so much with Belt rocks.

At lower elevations, erratics are common and obvious. As elevation is gained along the ridges, the erratics abruptly become scattered and less obvious (less rounded, more weathered) until, on some of the highest ridges none were found (Fig. 14).

Erratics degree of weathering, combined with their lithology, suggest their relative ages. The lower, common erratics range widely in their size and lithology. Their surfaces are relatively free of pitting and other obvious weathering. Scattered erratics (yellow “X”s, Fig. 14) are more weathered and of an overall limited number of lithologies compared with those lower in elevation.

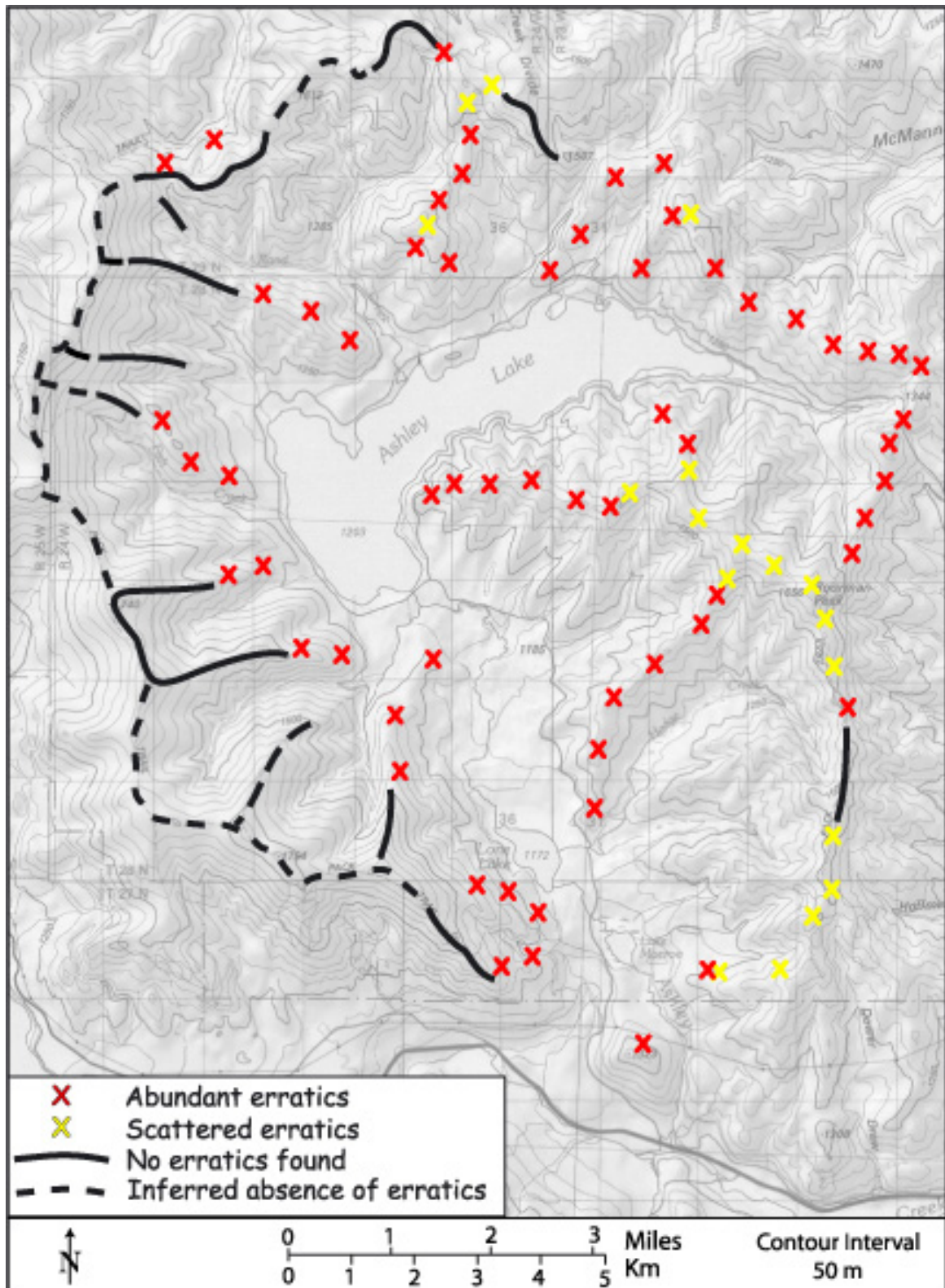


Figure 14 – Distribution of erratics

Depositional landforms

Depositional and erosional landforms were located and mapped. Landforms that were large enough to show morphology at the scale of figures in this thesis are mapped according to color (Fig. 15). Those that are too small are coded by a representative letter. These included those previously indicated by topographic maps and aerial photo interpretation, and those that were discovered while completing fieldwork.

Moraines

Moraines in the study area are diverse in their topographic setting and morphology. Moraines vary in length from 200 - 1700 m and in height from a meter or less to approximately 15 meters (Fig. 15, orange areas and “M”s). The largest are located in the Ashley valley bottom. In most cases, their boundaries are poorly defined; they are hummocky with abundant bogs and are incised by local streams. The smallest and most well defined moraine is located on the major southwest-northeast trending ridge east-southeast of the Ashley Lake outlet. It is linear, with steep proximal and distal sides that cause a closed depression that has escaped fluvial incision. There is a continuum of moraines between the largest and the smallest.

Moraine morphology was not considered as a measure of age. Most occur in widely varying settings (topography, bedrock vs. sediment floored, *etc*) and do not allow meaningful numerical comparison.

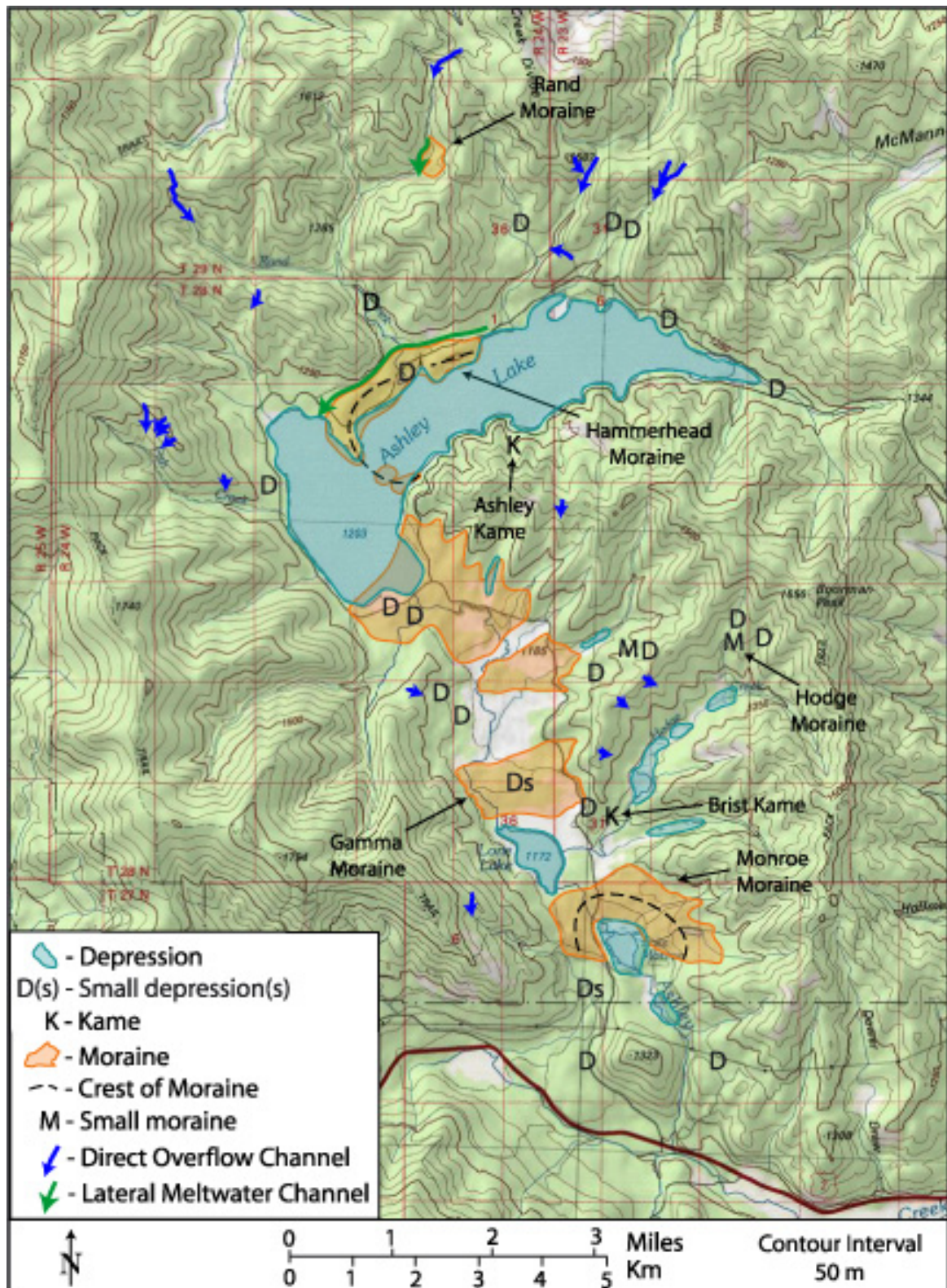


Figure 15 – Depositional and erosional landforms

Kames

Two kames were located and mapped (Fig. 15, “K”s). Kame deposits are sorted and contain clasts that are predominantly rounded, heterolithic and deposited on inclined bedding planes, assuming that the deposit has not been tilted since deposition (Fig. 16). Both mapped kames are presently being mined for sand and gravel, one commercially, the other privately. Exposure by mining greatly simplified characterization of these two kames.



Figure 16 – Brist Kame cross-sectional view (rock hammer for scale)

Glacial depressions

Glacial depressions are divided into two categories based on the presence or absence of an obvious overland drainage channel. Closed depressions have no apparent overland drainage whereas bogs and lakes have at least a partially developed drainage. These

depressions were visually inspected for their general attributes and mapped (Fig. 15, blue, and “D”s).

Depressions vary widely in their size, shape, nature of their containment, vegetation and indicators of moistness. Sizes range from a few square meters on morainal hummocks to approximately 10 km² in the case of Ashley Lake itself. Morainal settings vary from a well defined landform that dams a bedrock floored drainage, thus creating a depression, to a simple low depression in an otherwise relatively level area on unconsolidated materials. Over two seasons of fieldwork, some depressions were inundated with water, while others were internally well drained. Moist depressions generally contain a variety of aquatic plants. Drier depressions contain abundant low grasses and sedges, with some also having a variety of large trees.

The deposits contained in depressions can be valuable for relative and absolute dating. Stratigraphic markers within deposits can aid in relative dating and organic materials can be dated numerically by radiocarbon techniques. If a depression had characteristics that showed promise of containing a relatively long sedimentary record, a more detailed study was conducted and is presented later in this chapter. In addition, their patterns and positions may help reconstruct glacial margins.

Erosional Landforms

Lateral Meltwater Channels

Two lateral meltwater channels were mapped adjacent to moraines (Fig 15, green arrows indicating flow direction). Channel cross-sections were v-shaped and/or flat-bottomed. The largest lateral meltwater channel, which was adjacent to the moraine across the “neck” of Ashley Lake, had an outwash plain or delta immediately beyond the toe of the moraine. The other lateral meltwater channel had no obvious outwash plain or delta. The two lateral meltwater channels had some similar attributes: 1) their gradients decreased to the west-southwest; 2) slopes on both sides of the channels were very steep sided; and 3) both occurred well below maximum ice elevations. Lateral meltwater channels, while common in the region, were limited within the study area.

Direct Overflow Channels

Direct overflow channels were found throughout the study area (Fig. 15, blue arrows indicating flow direction). Channel dimensions varied considerably. Some identified were only 10 meters long, while others extended for at least 400 m. Some cut deep gouges into the bedrock while others scarcely eroded it at all. Slopes on channels walls were often steep sided and, in some cases, vertical. Because many direct overflow channels cut across moderately steep ridges, there was not an obvious gradient to show which way the water was flowing. However, one side of the ridge was typically eroded much further down than the other (Fig. 17).

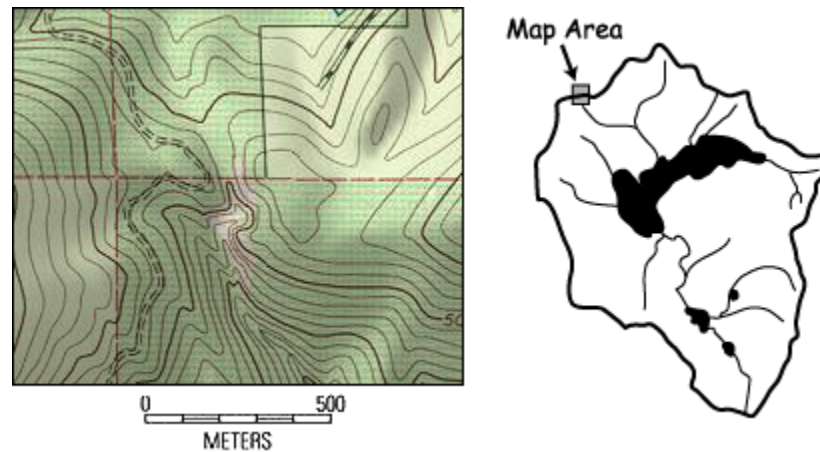


Figure 17 – Direct overflow channel erosion

Weathering Rinds

Weathering rinds were measured at seven sites. As many as seven different lithologies were identified in any one excavation (rind pit). Lithologies, listed approximately by the number of occurrences, included gray limestone, gray dolostone, gray argillite, green argillite, gray quartzite, purple quartzite, brown siltstone, brown quartzite and diabase (see tables in Appendix A for complete listings of data). Experience showed that blocky clasts did not have consistent rinds. Due to these inconsistencies, and the fact that most clasts were blocky, most were not measured. Therefore, to increase efficiency, only sub-rounded to well rounded clasts were measured. Few clasts showed evidence of having been deposited with preexisting rinds and clasts that showed such evidence (asymmetric rinds, for example) were not measured.

Soils contained in the sampling horizon of rind pits showed widely variable carbonate content when tested with dilute HCl. Results of these tests are summarized below. The higher sites tended to contain less carbonate.

<u>Rind Pit</u>	<u>Reaction of soil to dilute HCl</u>	<u>Elevation (m)</u>
Monroe Moraine	No observation	1165
Gamma Moraine	No observation	1180
Brist Kame	Vigorous effervescence	1200
Ashley Kame	Vigorous effervescence	1215
Hammerhead Moraine	Vigorous effervescence	1250
Hodge Moraine	None	1325
Rand Moraine	Very little effervescence	1400

Table 2 – Soil Carbonate in Rind Pits

Limestone was the most common clast measured. As a result, these clasts were used for statistical analyses. Complete data are in Appendix A. Rinds as thick as 14.4 mm were measured. However, the maximum frequency of any thickness other than 0.1 mm, which was the most common thickness measured, was two rinds at 7.6 mm on the Rand Moraine. Rinds typically consisted of a fine-grained, chalky, light gray material that was less dense than the unaltered rock. Rinds showed no reaction to a dilute HCl solution, while the inner, unaltered rock effervesced. It is likely that the rinds formed through dissolution of the carbonate. The remaining material was unidentified. Clasts with very thick rinds were noticeably less dense than those with thinner rinds.

Bog Augering

Seven selected bogs were sampled by hand auger to recover tephra and organic material. Augerings A1 - A6 proceeded until being stopped by large clasts, which were probably in till (see example of log in Fig. 18, all logs can be found in Appendix B). Clasts precluded further augering due to the physical inability of the teeth to cut through them. The teeth were bent several times when large clasts were suddenly encountered. A7 had to be discontinued due to a collapsing borehole. The borehole was largely water saturated sand that collapsed multiple times when the auger was removed. As the borehole proceeded, a point was reached in which sand collapsed into the hole as fast as it could be removed and augering was terminated.

Tephra

Several intervals that contained potential tephra were recovered by augering. In the field, they were differentiated from other deposits by their light color, extreme grittiness, relative stratigraphic location, bedding characteristics such as a relatively sharp contact with other sediments, low competence of a layer versus high competence of bounding layers, and abundance of mafic grains. Deposits identified as potential tephra in the field were then further studied in the laboratory, as field characterization was deemed insufficient.

Augering: 4

Date: 07/17/02

Location: Drainage Exit Bog

Worker(s): Denny Capps

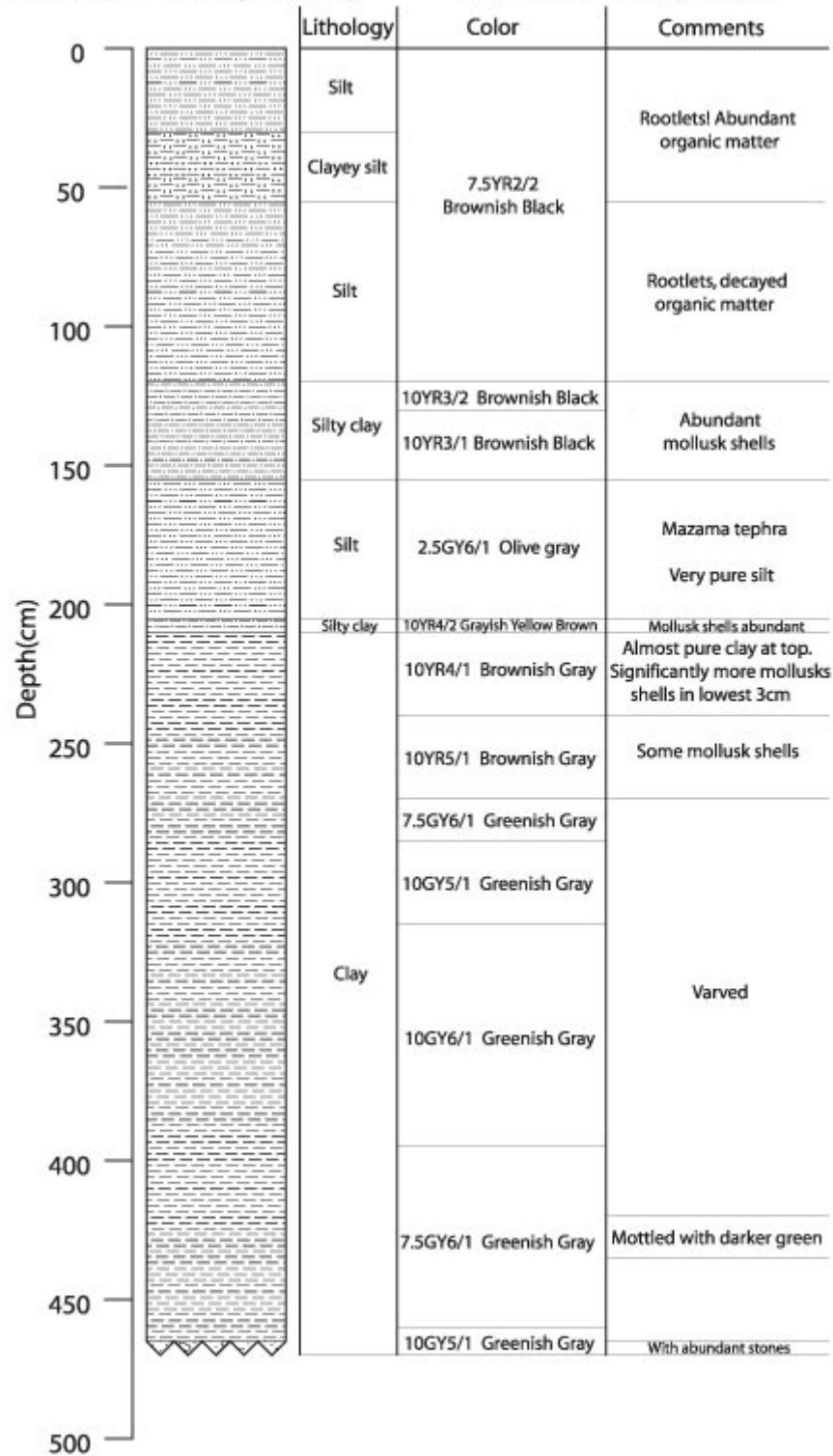


Figure 18 – Sample augering log

Carrara's (1989) detailed tephra analyses in bogs approximately 65 km to the northeast in Glacier National Park were used to characterize those found in the study area. Because it is a relatively short distance from this study area to Carrara's compared with the total distance traveled by the tephras, they are expected to be very similar. Prepared materials were characterized primarily by glass shard morphology and relative abundance of mafic crystals, which Carrara found to be good criteria to differentiate the tephras (Table 3). Three occurrences of the Mazama tephra and three occurrences of the Glacier Peak G tephra were provisionally identified in the laboratory (see Table 4 for an example of a worksheet, all worksheets can be found in Appendix C). Augering #2 was the only bog in which both tephras were identified. Further testing was done on a tephra identified as Mazama to offer more assurance of its identity. The Mazama, Glacier Peak G, and Mount St. Helens Jy tephras all contain crystals of plagioclase, hornblende, hypersthene and opaque oxides; the Mazama ash also contains crystals of augite and apatite (Lemke et al., 1975 and Sarna-Wojcicki et al., 1983). Crystals of augite and apatite were found in the sample identified as Mazama by glass shard morphology and relative abundance of mafic crystals. In addition, the refractive index of the Mazama glass is > 1.502 , higher than any other Holocene or late Pleistocene ash reported in Montana (Wilcox, 1965). The tephra identified as Mazama from the above techniques had a refractive index > 1.502 . The refractive index of a suspected Glacier Peak G ash was found to be < 1.502 .

<u>Tephra</u>	<u>Glass shard morphology</u>	<u>Abundance of mafic minerals</u>	<u>Refractive index</u>	<u>Thickness (cm)</u>	<u>Date</u>
Mazama	Commonly elongate	~5%	>1.502	3 - 100	6.8 ¹⁴ C ka
Glacier Peak G	Relatively equidimensional	15-20%	~1.498	0.3 - 3	11.2 ¹⁴ C ka
Mt. St. Helens Jy	Relatively equidimensional	~5%	~1.498	0.1 - 1	11.4 ¹⁴ C ka

Table 3 - Criteria for Differentiation of the Major Tephra

Site Location: Gamma Bog

Date: 07/12/02

Category	Property	Observation
Field Characteristics	Thickness	1 cm
	Bedding characteristics	Two lenses 5 cm apart, sampled lowest
	Grain size	Silt
	Color	Light Gray
	Stratigraphic context	73 and 78 cm, 229 cm total depth
	Grittiness	Glassy
	Reflectivity	None noted
Lab Characteristics	Percentage mafic crystals	~2-3%
	Degree of hydration	Low
	Morphology of glass shards	Very acicular with abundant bubble-wall and bubble-wall junctions

Probable tephra - Mazama

Table 4 – Sample Tephra Characterization Worksheet

Radiocarbon

Carbon was recovered from augerings by techniques detailed in Kihl (1975).

Macrofossil mollusk shells and microfossil plants were recovered from basal cores.

Augerings A1 and A4 contained macrofossil mollusk shells of snails and bivalves that were possibly the same two species (*Gyraulus parvus* and *Pisidium variable*) that Carrara (1989) identified in bogs approximately 65 km to the northeast in similar environments.

These materials were *not* dated for two main reasons. 1) These were aquatic organisms and the chemistry of their shells could therefore be influenced by the hard-water effect. This problem arises when aquatic organisms take up carbon from water containing bicarbonate derived from old, inert sources. This is a particularly difficult problem in areas where limestone and other calcareous rocks occur (Bradley, 1999). Because the country rock here is high in carbonate, very old (~1 Ba) and therefore completely dead in relation to radiocarbon, dates significantly older than their true age may result. It was therefore prudent not to use this carbon. 2) The shells are particularly susceptible to contamination by modern carbon because they readily participate in chemical reactions with rainwater and/or groundwater. Most mollusks are primarily composed of calcium carbonate in the metastable form, aragonite. This aragonite may dissolve and be redeposited in the stable crystal form of calcite. During the process of solution and recrystallization, exchange of modern carbon takes place and the sample is thereby contaminated (Bradley, 1999).

Augerings (A) 1, 2, 3, 4, 5 and 6 yielded small fragments of plant-like organic material. Dr. Sharon Eversman of the Montana State University Ecology Department analyzed these fragments under a reflective microscope for identification. A2 contained obvious surface contamination (green grass) and no identifiable subaerial plant material. Subaerial plant material is desirable to avoid problems associated with the hard-water effect. Parts of plants that are receiving their carbon from the atmosphere will be in relative equilibrium with atmospheric carbon and will therefore give realistic ages. Plants, like the previously mentioned mollusks, that grow wholly underwater will be influenced by the hard-water effect (Pierce, 2003, verbal communication). A3 was

contaminated by modern rootlets. A1, A4, and A6 had no *identifiable* subaerial plant material. A5 had one identifiable subaerial plant fragment. This sample was not dated for two reasons. 1) The fragment appeared to be a fresh, small wood chip that was probably contamination. 2) It would be the only date in the study area and therefore without the support of other data. A basal core of A7 could not be obtained due to the before mentioned problems with a collapsing borehole. As a result, no deep organic material was recovered from this augering.

Even if radiocarbon dating was to be done, the dates from these bog bottoms would not reflect the age of deglaciation, but instead represent the inception of organic rich sediment deposition that, in many cases, began well after bog formation. Estimates of the lag time between deglaciation and organic rich sediment deposition, represented by either no deposition or the deposition of inorganic fine-grained clastics, vary widely. The assumption that plant growth occurs soon after deglaciation is, in part, the basis for the use of bog bottom dates to determine the age of deglaciation.

Varves

Varves were encountered in the core of A4. Varves were not measured in detail because of the nature of the specific auger used. The auger disturbed the deposit and did not collect them all in one continuous sample. Couplet thickness varied, but averaged 5 mm with at least 30 couplets represented.

Ice modeling

Identified end moraines and valley geometries measured from topographic maps were used to reconstruct glacial lobes using GlacPro. The lobes were constrained by erratics, and depositional and erosional landforms. These constraints allowed estimation of the effective basal shear stress. Two major flowlines were modeled for glacial profiles at different times (Fig. 13). The Ashley Piedmont Lobe, effectively an outlet glacier, entered from the north and east where it spilled over the drainage divide into the Ashley basin. The Ashley Creek Lobe entered the valley from the south and faced no major obstacles other than an uphill gradient. Evidence of this upvalley advance was first documented by Alden (1953) in his observations of moraines, outwash plains, rounded glaciated ridges and striations.

The Ashley Piedmont Lobe was modeled for three different ice margins: 1) the lastglacial maximum “LGM”; 2) lastglacial recession 1 - “LGR 1”; and lastglacial recession 2 - “LGR 2”. The model was run along the center bifurcation of the flowlines shown in Figure 13. The “LGR 2” margin was modeled based on the end/lateral moraine system that forms the “neck” of Ashley Lake. This site was modeled first because it is the only location where an end/lateral moraine system was found. The end/lateral moraine combination provides more constraints to the model. The end moraine provides a starting point and the lateral moraine helps to constrain ice thickness that allows calibration of the main variable of the model, effective basal shear stress. An effective basal shear stress of 0.3 bars causes the ice elevation to closely match the elevation of the lateral moraine (“LGR 2”, Fig. 19). This calculated effective basal shear stress contrasts with a typical

number of 1 bar in other similar mountain settings (Pierce, 1979). The second margin, “LGR 1”, was modeled based on an end moraine approximately one kilometer southeast of the Ashley Lake outlet. Direct approximation of the effective basal shear stress was not possible on this or subsequent ice margins because they do not have lateral moraines or other attributes that allow such. An effective basal shear stress of 0.3 bars was used because the controls on this margin are not expected to be largely different from those of LGR 2 (“LGR 1”, Fig. 19). There were scattered deposition and erosional landforms that matched up with modeled margins. However, they did not match up consistently. As a result, they were not plotted on the graph to avoid clutter. The third and highest margin, the lastglacial maximum, “LGM” was modeled based on limited information provided by highest common erratics. To match the modeled ice margin to highest common erratics necessitated a very low effective basal shear stress of 0.1 bars (“LGM”, Fig. 19).

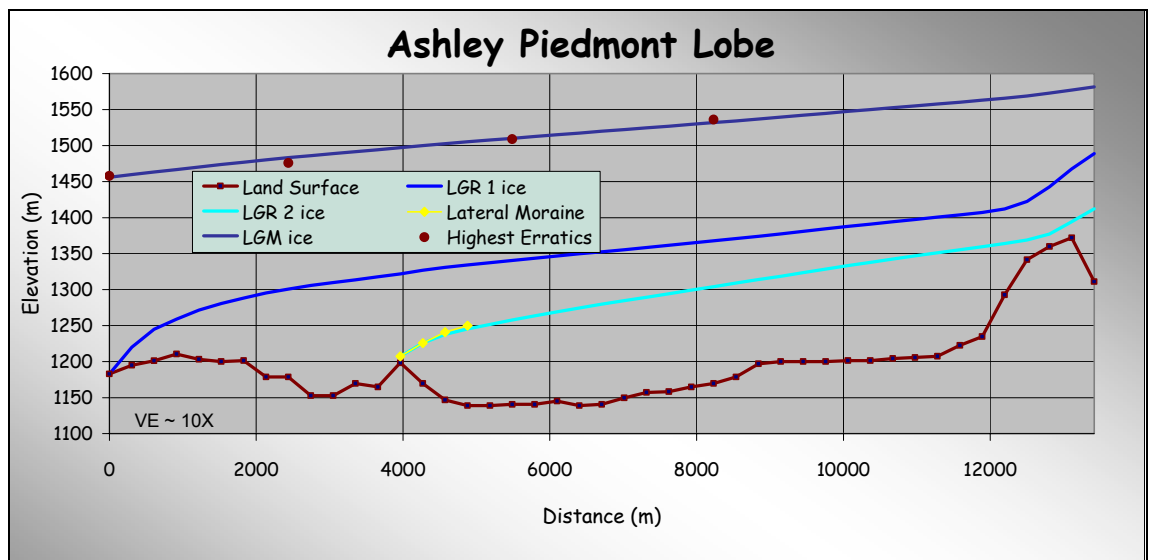


Figure 19 – Graph of Ashley Piedmont Lobe ice margins

The Ashley Creek Lobe intruded from the south and was modeled along the western bifurcation of the flowline in Figure 13. The Ashley Creek Lobe was modeled for two of the three same ice margins as above (Fig. 19). The first margin, “LGR 2”, was modeled starting at an end moraine just upstream Ashley Creek from its confluence with Idaho Creek (Fig. 20). Unlike “LGR 2” of the Ashley Piedmont Lobe, there were no lateral moraines to constrain the effective basal shear stress. An effective basal shear stress of 0.3 bars was assumed because it is expected that this lobe had similar conditions at its terminus to that of the terminus of the Ashley Piedmont Lobe. “LGR 1” was modeled based on an end moraine just south of the end moraine modeled in “LGR 1” of the Ashley Piedmont Lobe. As was the case with the margin of the Ashley Piedmont Lobe, there were scattered depositional and erosional landforms that matched up with modeled margins. However, they did not match up consistently. As a result, they were not plotted on the graph to avoid clutter. Again, an effective basal shear stress of 0.3 bars was used for reasons already discussed. The lastglacial maximum was not modeled on this profile because during this time, ice was being supplied to this area on all sides except the west, rendering a simple model as used here inadequate. The modeled elevation should not match the highest erratics of this area because at this higher stage, the two ice lobes had coalesced and a new, single gradient to the south or west was achieved.

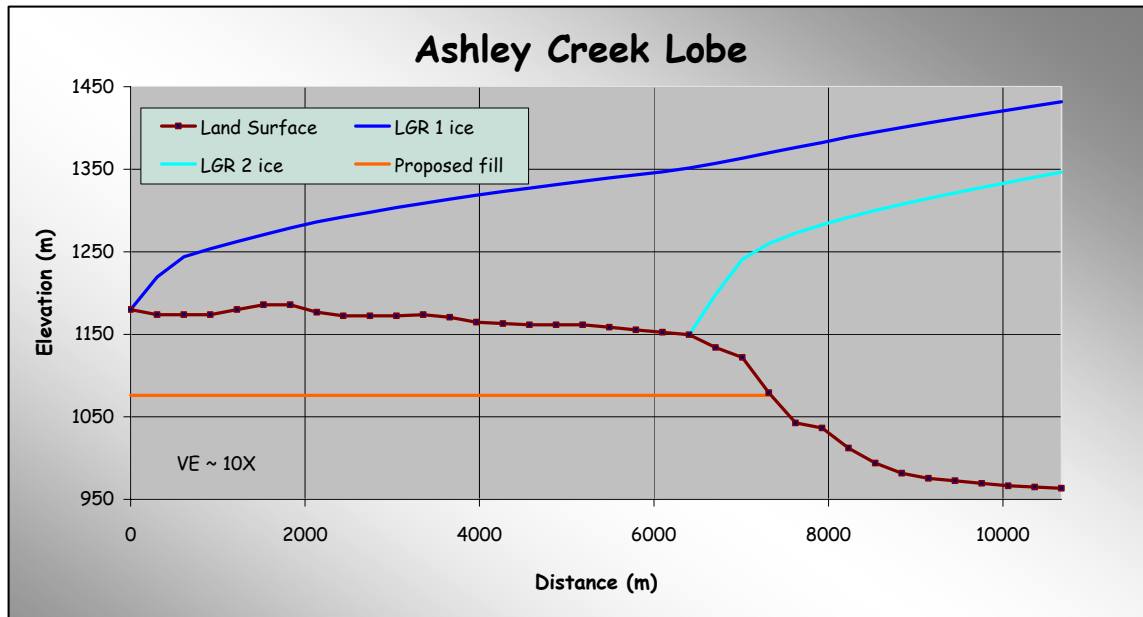


Figure 20 - Graph of Ashley Creek Lobe ice margins

DISCUSSION

The discussion will be presented in the following format: 1) discussion of features; 2) defining significant ice margins that created these features; and 3) defining a glacial history.

Discussion of features

Erratics

Erratics are widespread and relatively easy to interpret. Throughout most of the low and mid-elevations of the study area, erratics are common. Simply walking any direction for 100 m will normally yield at least one erratic. From any one vantage point, it was commonly possible to see several erratics. As elevation is gained along ridges, the boundary between common and scattered erratics is determined by a very rapid decline in abundance. Initially, it was thought that this was an upper limit for erratics. Further careful observation showed that more erratics could be found above this level, but in much less abundance (scattered). Above the boundary, several hundred meters could be walked without identification of any erratics. Not only were the erratics scattered, but they were harder to recognize. Typically, these scattered erratics were more weathered and of a limited number of lithologies compared with those lower in elevation. Locally derived erratics tend to weather much quicker than erratics of other common lithologies. For example, dolostone erratics derived from local bedrock are easily weathered by dissolution due to their high carbonate content. Local bedrock also tended to have very well defined bedding planes that encouraged splitting, especially by frost action; splitting

of the scattered erratics made them harder to recognize due to their more angular appearance. Lithologies of erratics derived from outside the study area included diabase and argillite. These lithologies often showed no signs of weathering, as if they were only recently deposited. For example, the diabase showed no bedding planes because it is a massive, intrusive igneous rock. Argillite erratics, especially the apple-green variety, commonly showed no well developed bedding planes. The lack of well defined bedding planes in these types of erratics helped preserve their glacially sculpted shapes in addition to being very resistant to rind formation.

Depositional Landforms

Several of the moraines are located at relatively narrow spots in the main valley (Fig. 15). This suggests that the glacier initially may have been floating in a proglacial lake and its terminus was stabilized by the constriction. Glaciers terminating in deep water are potentially unstable and vulnerable to catastrophic retreat by rapid calving. Ice margin stability is therefore favored by the presence of pinning points or constrictions in the enclosing channel. Such constrictions in the channel occupied by a glacier reduce losses by calving and if the calving rate is less than the ice flux, the glacier will thicken and stabilize. However, if the glacier pulls back from a pinning point into deeper water, the calving rate may exceed the ice flux and rapid retreat will result (Benn and Evans, 1998).

The two mapped kames had not been identified prior to fieldwork. One was noted as a possible moraine because it was shown on the map as a gravel pit. The other was found because it had been incised by a county road. It is therefore likely that there are many more kames present in the study area that have not been identified.

Depressions are common throughout the study area, as is the case in many areas that were glaciated in the last 20 ka. Many processes can form closed depressions. Three different mechanisms are interpreted to be at work within the study area. Two were a result of the direct action of ice and one was the result of mass wasting processes. One glacial mechanism is the formation of kettles. Kettles are depressions in glacial drift, often containing a lake or swamp, formed by the melting of a detached block of stagnant ice that was buried in the drift (Bates and Jackson, 1984). Ashley Lake, east of the Hammerhead Moraine, probably owes much of its existence to the process of kettle formation. Another, much smaller closed depression, located at the eastern tip of Ashley Lake (Fig. 15), is also interpreted as a kettle. Like many kettles, it is surrounded by a slightly raised, hummocky rim and has a poorly developed outflow channel. This rim is vegetated by large coniferous trees and the depression by water tolerant plants. During the summer of 2001, it was largely dry, but in the summer of 2002, it was occupied by shallow water. The second glacial mechanism is the formation of ice marginal depressions. Ice marginal depressions are created by glacial ice directly depositing till that blocks drainages. These depressions are the most common type of closed depression in the study area. Some of these ice marginal depressions were small (a few meters across), dry and contained no outlet channels. Others were large (up to a square kilometer), contained water and had well developed channels entering and exiting them (Lone Lake). There is a continuum between kettles and ice marginal depressions, so a division into two neat categories is somewhat arbitrary. There is an apparent mass movement induced closed depression on the edge of the northeast shore of Ashley Lake (Fig. 15). An elongate topographic low with relatively sharp edges trending down a slope

with a lobate mass on the valley bottom suggests mass movement. The lobate form on the valley floor contains a closed depression with standing water.

In some locations, there is at least 100 m of sediment fill in the main valley as indicated by topography, well logs and Alden's (1953) conclusion that ice crowded into the basin and "there deposited the great fill of drift that holds Ashley Lake"(p. 118). In the upper basin, Ashley Lake is 68 m deep. Where the Ashley Creek valley meets the larger valley just before the confluence with Idaho Creek the elevation drops precipitously by at least 75 m. Both sites point to the existence of an extensive fill.

Erosional Landforms

Lateral meltwater channels and direct overflow channels are genetically related and form a continuum. However, they have been differentiated based on their dominant characteristics. The two identified lateral meltwater channels are parallel to moraines (Fig. 15). If the depositional landform is correctly identified as a lateral moraine and the channel-like erosional landform next to it is a channel, then a channel flowing parallel to a moraine is a lateral meltwater channel by definition. Numerous direct overflow channels were found while walking ridges. Most had not been previously interpreted from topographic maps and aerial photos due to their relatively small sizes and forest cover.

Weathering Rinds

Weathering rinds were only measured on glacial deposits. To measure rinds by the proposed methodology required a deposit of at least 0.2 m in depth, but not deeper than

0.5 m. Surface erratics could not be used because they did not satisfy this requirement. The only locations where glacial deposits of the appropriate depth were found were at mid-to-lower elevations; therefore, rind measurement was limited to these areas. The abundance of gray limestone and dolostone clasts in deposits is explained by the fact that these two rocks make up much of the bedrock of the study area. Other lithologies transported in from outside the study area were fewer in number because their relative abundance was diluted by the local bedrock.

Bog Augering

Augering, like weathering rinds, took place at relatively low elevations due to availability of depressions. As a result, tephras, organic material and varves were recovered from low elevations and limited extents. Immediately after deglaciation, ice marginal depressions will typically be floored by till or other coarse glacial deposits and initial nonglacial deposition will be rapid. Rapid sedimentation is a result of the high availability of sediment and water at the margins of warm based glaciers and a probable lack of vegetative cover. Vegetation inhibits sediment mobility and runoff. Varves, in addition to those found while augering, were encountered in a test soil pit excavated in the Gamma moraine. Here, the total varve package measured 50 cm in thickness, with the lowest of at least 15 couplets being 5 cm in thickness. The varves decreased in thickness up-section until they were no longer recognizable from the overlaying strata.

Defining significant ice margins that created features

Erratics

The presence of erratics provides direct evidence for defining former ice elevations. Any extent of a ridge that is mantled with erratics was once covered with ice. Erratics can only define the most extensive ice margin within any one glaciation. Anything below the most extensive ice margin will likely be mantled by erratics.

A new finding of this study, primarily based on common erratics, is that ice elevations were high enough to advance over divides into the Ashley basin. This provides new information on the surface elevation and minimum thickness of the Flathead Lobe of the Cordilleran Ice Sheet. At the latitude of Kalispell, the ice was at least 1500 m in elevation (Fig. 21). These findings are a significant departure from the work of Richmond (1986) as well as topographic map and aerial photo interpretation of Locke and Capps prior to fieldwork. Of significant note is that there is very little ice surface gradient in any direction. The low to nonexistent north-south gradient in the Ashley Lake area appears consistent with the observation that drumlins in the Kalispell area indicate that the lastglacial flow was east to west. In addition, striations in the lower Ashley Creek valley show east to west flow (Alden, 1953).

Due to lack of evidence, erratics do not provide an identifiable maximum for the pre-lastglacial, only a minimum (Fig. 22). There were areas within the study area in which erratics were not found (Fig. 14). There are three possible explanations. 1) These areas were never glaciated. This is the most probable explanation for the highest areas such as the top of the western ridge. For example, the top of Pleasant Mountain is over 360 m

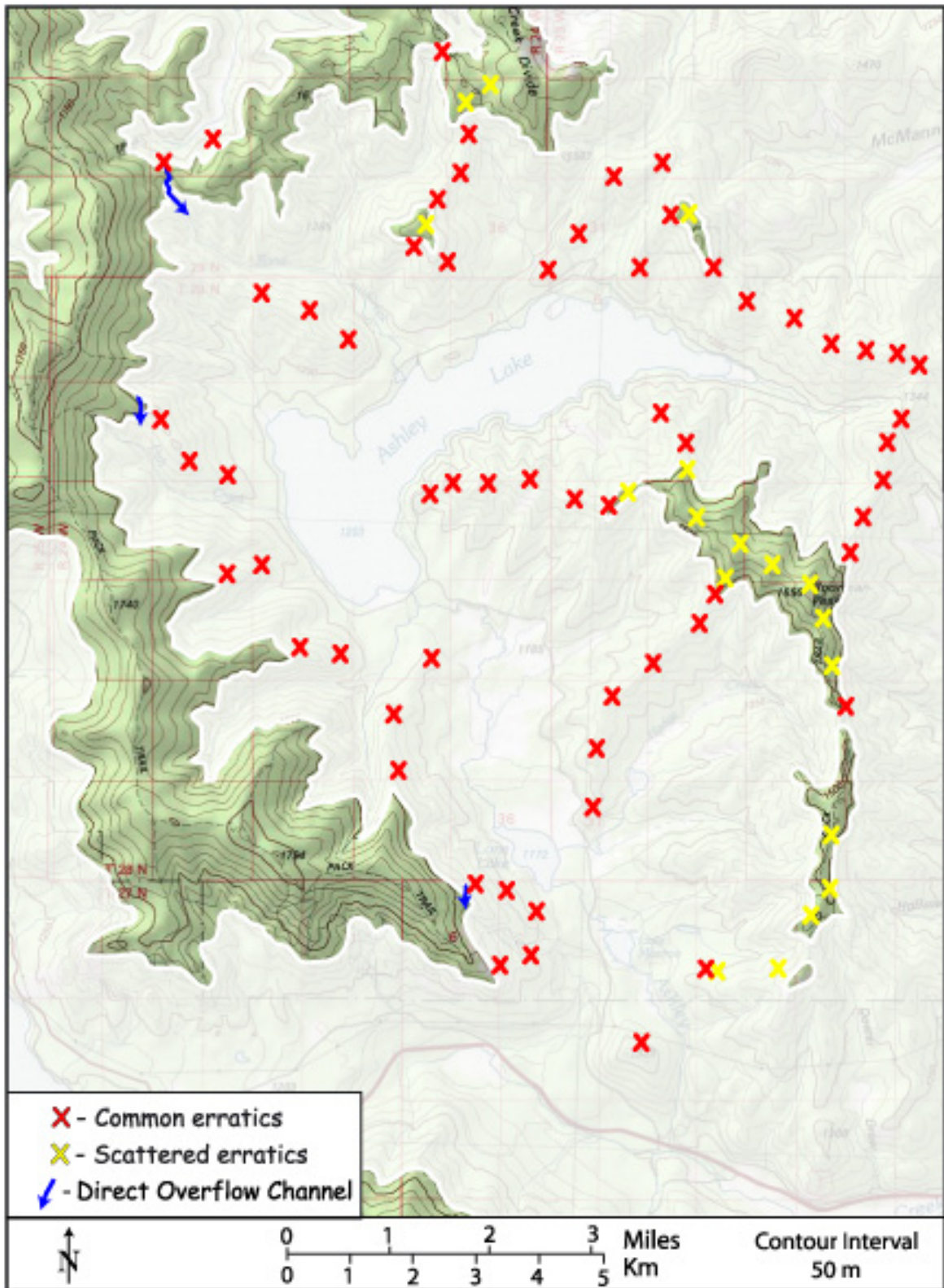


Figure 21 – Interpreted lastglacial maximum

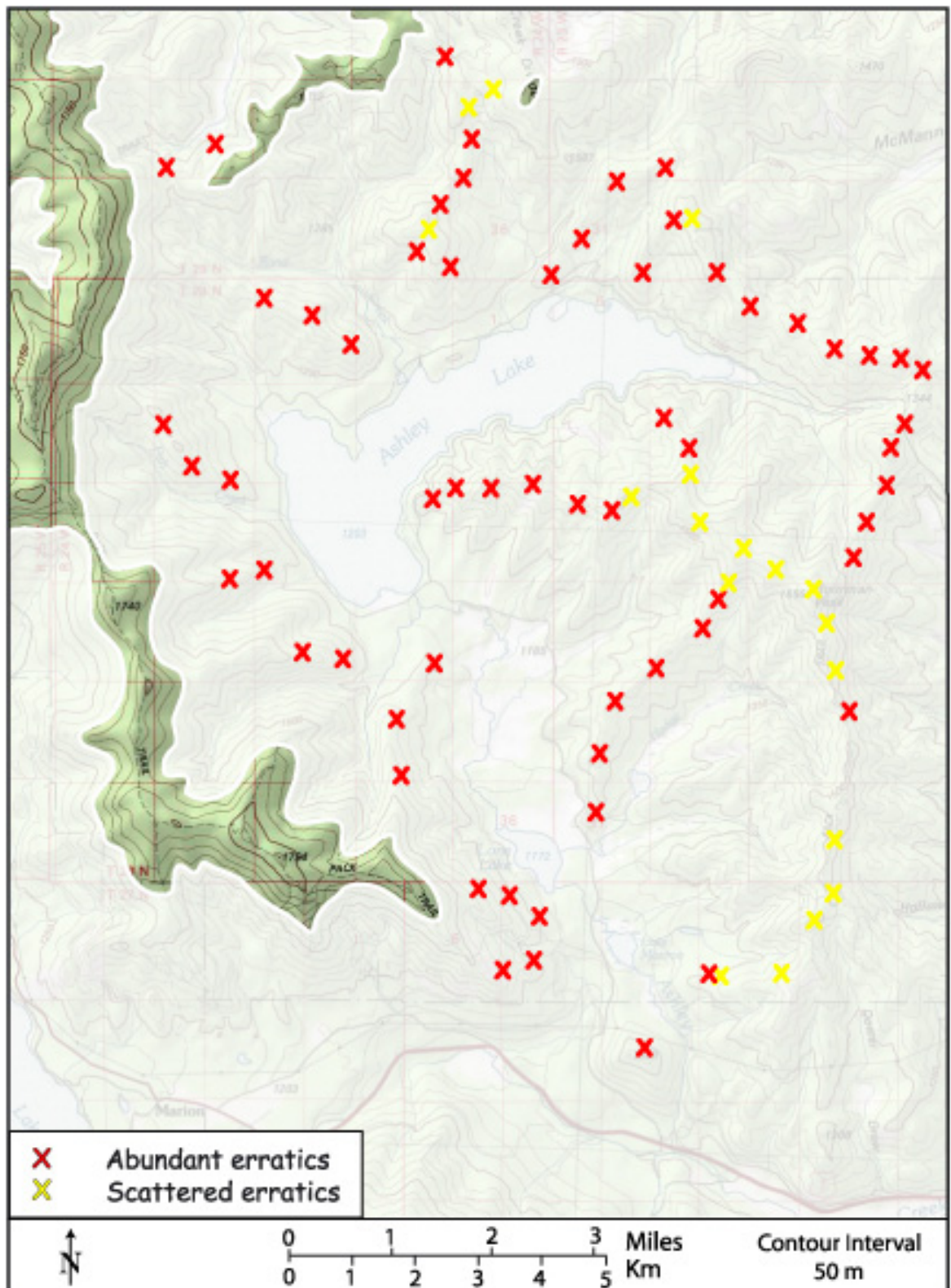


Figure 22 – Interpreted pre-lastglacial minimum

above the highest “common” erratics. 2) These areas were glaciated and erratics were left behind but were not found during fieldwork. Some ridges, especially those in the north-northwest corner, are covered mature forests on low gradient slopes. These areas produce significant organic matter that may conceal erratics. 3) These areas were glaciated, but due to subsequent erosional processes, erratics were moved down slope or destroyed by natural processes such as fire spalling. Fires, both natural and human caused, are very significant drivers of geomorphic change in this region. Multiple examples of fire spalling were observed. Recently burned logging debris piles caused clasts as large as 35 cm to break into much smaller fragments. Nunataks and other areas above past glacial margins were subjected to intense frost shattering processes. This could result in the colluviation of erratics from previous glaciations to lower elevations or cause them to be broken, rendering them unrecognizable as an erratic. Normal slope processes likely transported erratics to lower levels through time as well.

Depositional Landforms

The record of glaciation as defined by moraines is a selective one in the study area and elsewhere. Glacial advances generally obliterate moraines deposited by less extensive, previous glacial advances. Obliterative overlap is the process of obliteration of a moraine when overlapped by a later, more extensive glacial advance (Gibbons et al., 1984). However, moraines were very useful in interpretation of the last deglaciation. In some cases, their location combined with arcuate morphology, which showed the direction of retreat, allowed reconstruction of the last glacial recession (Fig. 23). The arcuate form can be seen in at least three locations: 1) the Monroe Moraine, located near

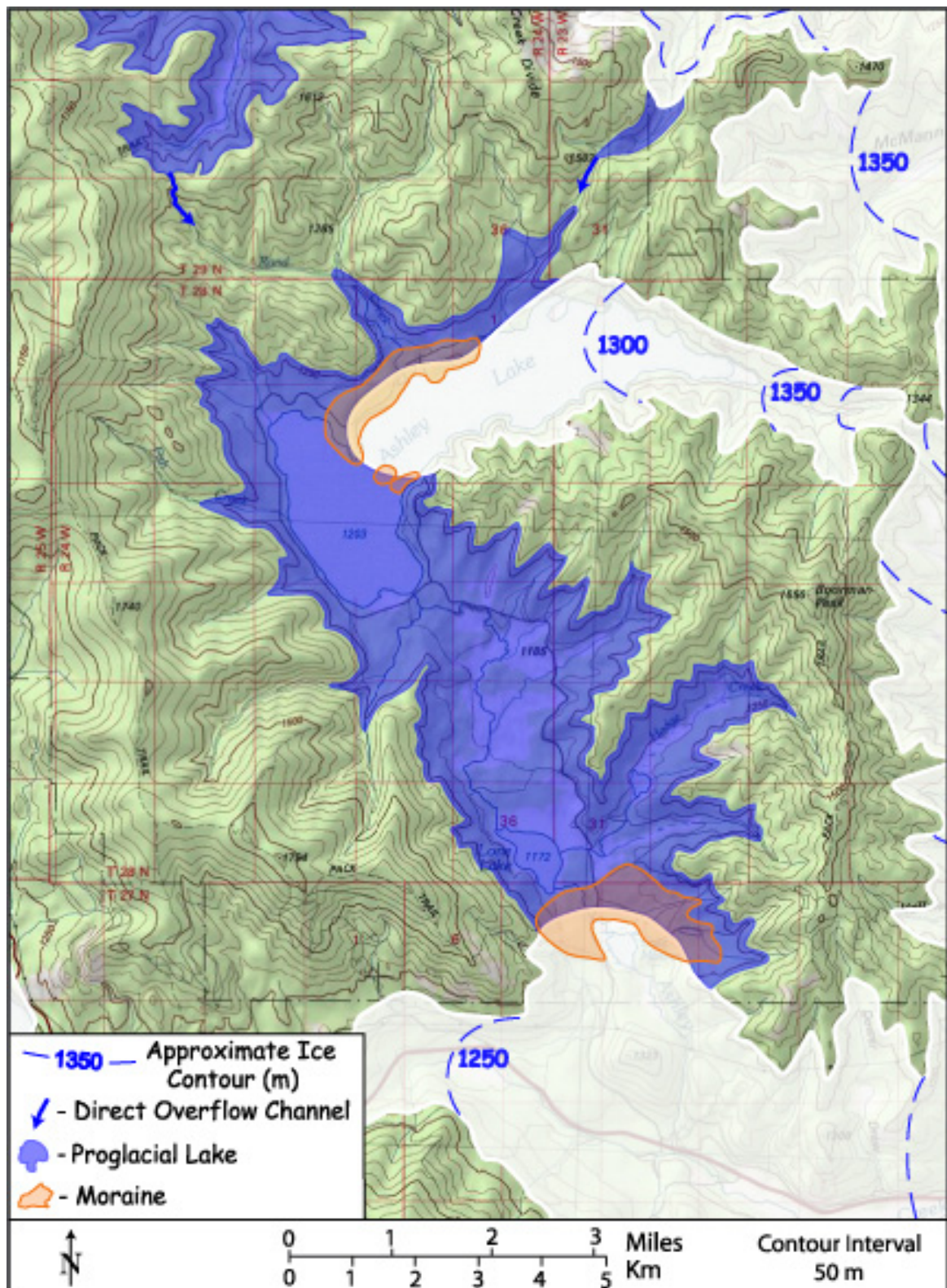


Figure 23 – Interpreted lastglacial recession

the southern end of the drainage; 2) the Hammerhead Moraine stretches across the neck of Ashley Lake; and 3) the Rand Moraine, located within the northeast fork of Rand Creek. Other, more subtle, lobate morainal landforms could arguably be considered as well, but the shape of associated closed depressions that will be discussed later better characterizes these deposits.

The presence of kames indicates that glacial and fluvial processes were taking place at these locations (Fig. 15). These processes were below the ELA where abundant meltwater was available. Both kames were at low elevations and therefore define later stages of deglaciation.

Closed depressions were also located at low elevations and only represent deposition from later stages of deglaciation. Because kettles are indicative of stagnated ice, they can provide valuable information on the condition of the ice at the time of their deposition. The ice of the Ashley Piedmont Lobe became stagnant once it was pinched off from its source by down wasting. The drainage divide to the east and northeast cut it off from the main lobe of the ice sheet to the east. The mass movement induced closed depression is interpreted as having been triggered by oversteepening of till during recession of the ice. Mass movements are common in recently deglaciated areas. Since the head of the scarp is located at approximately 1430 m and the toe of the transported material is at approximately 1200 m, the basin would have had to have been almost completely deglaciated by the time the mass movement took place.

Most of Ashley Lake was formed because of a large mass of stagnant ice. This stagnant ice was present while the remaining sediment supplying glaciofluvial processes were still taking place in the valley. As the active ice margin receded away from the

basin, sources of glacial and glaciofluvial sediment were greatly reduced, leaving the ice to melt out and leave behind a closed depression that forms much of what is now Ashley Lake.

Assuming that there is a “great fill” of sediment in the Ashley valley, an explanation is needed as to how it was deposited. This material was likely deposited upvalley from the Ashley Creek Lobe and downvalley from the Ashley Piedmont Lobe as they advanced and retreated from the basin. This setting would provide an ideal sediment trap with abundant sediment supplied by meltwater and ablation from both lobes. It is also possible that a similar ice lobe geometry existed in previous glaciations that could have deposited some of the material in previous episodes. It might be expected that since there was a “great fill” deposited here that it may have regional significance. However, another explanation is more likely. First, no consistent, similar features were found throughout the region at similar ice margins. Second, the “great fill” is likely due to the sediment trap described above which was due to local topography and not a climatically induced stillstand of the ice.

Erosional Landforms

Fluvial-like channels with little or no modern discharge are described as lateral meltwater channels because they flow parallel to probable ice margins and they do not cut across obvious, large-scale bedrock ridges. Lateral meltwater channels, although limited in number in this study area, can be used to reconstruct regional patterns of ice recession. In subpolar regions like the study area, lateral meltwater channels are explained as the product of: 1) high subglacial water pressures deflecting some surface

water along the glacier margins; 2) spring snow melt on surrounding slopes and meltwater drainage along the glacier margin while the bed was still frozen after the penetration of the winter cold wave; or 3) cold-based conditions in the outer marginal zone of otherwise temperate glaciers (Fulton, 1967). The channel morphology of the two examples in the study area closely paralleled adjacent moraines that have previously been discussed. The moraines and lateral meltwater channels describe late stages of deglaciation of the lastglacial.

Because many direct overflow channels are eroded only on the very tops and one side of the ridge, the direction of water flow could be determined. Water can be ponded by ice and bedrock dams until it overtops a low divide. When this happens erosion cuts headward causing more erosion on the downstream side. Most direct overflow channels occurred at or near maximal ice elevations. The most significant of these is the “Logan” direct overflow channel (Fig. 17), so named because it connects the Ashley basin with the Logan Creek basin to the north. It is the largest and most well defined of all direct overflow channels in the study area. It is significant in that it shows that water was entering in to the Ashley basin from the Logan Creek basin and carving a very ragged channel into bedrock to lower elevations. The most plausible explanation is that a lobe of ice moved up Logan Creek from the north damming it and forming a reservoir that eventually spilled into the Ashley Lake basin. It is possible that Ashley Lake itself was higher in surface elevation as evidenced by possible depositional features (perched deltas) that were fed by the overflow from the Logan basin. Common erratics ended just below this divide, indicating that the lastglacial ice did not overtop it, yet water appears to have flowed through it and cascaded to much lower levels in the Ashley basin. This

suggests that water probably overflowed this divide at widely varying ice elevations through time. This theory is further supported by the fact that it is the most well developed direct overflow channel found in the study area. Water overflowing a drainage divide at an elevation of approximately 1530 m means that the ice dam was at least this high to the north. The large number of direct overflow channels in the study area shows that there was an abundance of available water, a condition that would be expected for a warm-based ice sheet below the ELA.

Bog Augering

The presence of varves in the core of augering A4 and the test soil pit is evidence of a possible glacial lake. The varves in A4 overlie gravel, suggesting that immediately after deglaciation a proglacial lake occupied the area of the bog. Three likely hypotheses exist. 1) These deposits could be the result of Glacial Lake Missoula. Glacial Lake Missoula reached a maximum elevation of 1280 m, so its water could have reached the location of these deposits. If the 1280 m level of Glacial Lake Missoula occurred during the maximum extent of the Pend Oreille Lobe, the lake may have existed at the Flathead Lobe ice maxima. This would not have allowed the deposition of Glacial Lake Missoula sediments because all areas below 1280 m were covered locally by the Flathead lobe. However, it is quite possible that the maxima of the Pend Oreille and Flathead Lobes were not synchronous. In this case, the highstand of Glacial Lake Missoula would have corresponded with times of deglaciation in the study area, possibly allowing varve deposition in closed depressions. 2) These deposits could be the result of Ancestral Flathead Lake. Recent research has shown that the maximum level of Ancestral Flathead

Lake was no higher than 958 m above sea level, far too low in elevation to have deposited the varves at approximately 1115 m above sea level (Smith, 2003, verbal communication). 3) Lastly, these deposits could be the result of a localized proglacial lake. Water would almost certainly have been impounded by ice receding out of the Ashley Creek valley toward Kalispell. It is therefore more likely that these varves are a result of locally impounded glacial lakes. No evidence was found to show how the water from this lake eventually drained. The water may have drained subglacially, with subsequent reworking of the depositional/erosional features by the glacier. It is also probable that other, smaller proglacial lakes existed during deglaciation in the tributary valleys such as Rand Creek. A glacier front that is becoming more distant, thereby introducing the sediment at an increasingly greater distance, may explain the decreasing upsection thickness of the varves. Other working hypotheses include a reduction in water/sediment input, vegetative stabilization of recently deglaciated slopes or climate change.

Ice Modeling

Any theoretical glaciation can be modeled. However, more evidence to tune the model means that more confidence can be placed in the outcome. All major groups of erratics, and depositional and erosional landforms were modeled. The data are permissive of a model but do not constrain all margins of the model. Of all stages of the Cordilleran Ice Sheet considered in this study, the lastglacial recession was most realistically modeled because it had the most control. In one case, an end/lateral moraine was measured to delineate the ice margin.

According to evidence of moraine morphology, the ice separated into two lobes as deglaciation progressed. Because of the widely varying topography affecting the two lobe's entrance to the basin, they had two very different modes of deglaciation. As the Ashley Piedmont Lobe retreated, it was punctuated by pinning points that stabilized it in deep water. Ultimately, major thresholds, the north and eastern divides to the drainage basin, controlled deglaciation of the Ashley Piedmont Lobe. As the Ashley Creek Lobe retreated, it was punctuated by pinning points and the relatively steep entrance to the Ashley basin. However, there were no major topographic ridges to pinch off the ice from its source and cause major stagnations. Modeling was deemed worthwhile to facilitate the understanding of how topography affects deglaciation and to obtain estimates of the effective basal shear stress.

It is probable that both of the termini of these lobes were floating in proglacial lakes for some time during deglaciation. The Ashley Piedmont Lobe was almost certainly floating at some point during its retreat and stagnation. Limited end and lateral moraine evidence (Fig. 19) gives direct support in the model for lower effective basal shear stresses than would be normally expected. This deep water would explain the calculated effective basal shear stress of 0.3 bar, which is much lower than an expected number of 1 bar that would be typical of ice in this setting (Pierce, 1979). In most cases, effective basal shear stress could not be determined due to lack of constraining features. Because it appears that the first part of the study area to be deglaciated was near the center, proglacial lakes were likely impounded at varying levels. Lake levels would have a wide variety of controls. The Ashley Creek Lobe was responsible for damming this lake. Water contribution to the lake included precipitation, meltwater from ice to the south and

northeast, and for some interval, direct overflow channels that entered the basin from the north.

The lastglacial maximum, as modeled here, showed a very low effective basal shear stress. Because a lake level of at least 1400 m in elevation, for which there is no evidence, would have been required to lower the effective basal shear stress to the modeled number, another explanation is necessary. One possible explanation is that there was very little ice discharge through the system. In addition, since the ice entered the basin from at least two different directions, there would be no simple, unidirectional ice gradient from the north. As the ice became deeper, topography controlled it less and less. In fact, it is likely that the ice surface sloped down to the southwest or perhaps even to the west due to major inputs of ice from the east such as those from the area of current day Glacier National Park and the Swan Valley.

Defining a glacial history

Erratics

The presence, abundance and characteristics of erratics can be used to infer a glacial history of the study area. Because the best examples of erratics in the area are from the lastglacial, they will be introduced first, then older features compared to them.

Erratics are not helpful in differentiating lastglacial recessional deposits. Unless significant time passes between maximal ice positions and complete deglaciation, erratics at all levels appear similar. Erratics were the best indicators of the lastglacial maxima. The change of erratics from “common” to “scattered” is interpreted as the boundary

between lastglacial deposits (common) and a previous stade or glaciation (scattered). This boundary is well illustrated by erratic evidence on the multiple ridges of Boorman Peak in the east-central part of the study area (Fig. 21). For this reason, the glaciation that includes “common” erratics will be called the “Boorman Peak Glaciation”. Boorman Peak Glaciation erratics at all elevations are similar in their individual attributes and abundance. Because these erratics occur in areas where there are numerous closed depressions and other relatively fresh glacial landforms, the Boorman Peak Glaciation is interpreted as being approximately equivalent to Wisconsin deposits of the Laurentide Ice Sheet and to Pinedale deposits of Rocky Mountain alpine glaciers.

The highest proposed glaciation, where “scattered” erratics were found, extends to elevations above the top of Boorman Peak. As a result, Pleasant Mountain, on the higher ridge to the west, is used to define this glacial episode. The stratigraphic unit that includes “scattered” erratics will be called the “Pleasant Mountain Glaciation”. The characteristics of these erratics can be used to infer which glaciation deposited them. Based on the differences in weathering, identified lithologies and abundance of erratics, it is inferred that a long period of time elapsed between the Pleasant Mountain Glaciation and the Boorman Peak Glaciation. With the additional time between the two stades, Pleasant Mountain Glaciation erratics could weather significantly and become harder to recognize. The higher erratics tended to be lithologies that contained relatively high percentages of silica, which is more resistant to weathering than those that are predominantly calcareous. Examples included apple-green argillite and quartzite. They occur in areas where there are no identified glacial landforms. Consequently, the Pleasant Mountain Glaciation is interpreted as being approximately equivalent to Illinoian or even

older deposits of the Laurentide Ice Sheet and to Bull Lake or older deposits of Rocky Mountain alpine glaciers. Furthermore, Illinoian and Bull Lake deposits are typically more extensive than Pinedale or Wisconsin glaciations. Older, more extensive glaciations than the Illinoian or Bull Lake are not commonly reported in the Rocky Mountains so an Illinoian or Bull Lake age is favored for this area.

Depositional Landforms

Moraines from Pleasant Mountain Glaciation and the Boorman Peak maxima were not recognized. Erratics show that the Boorman Peak Glaciation reached higher elevations than any identified moraines. There are several ways to explain the absence of moraines. 1) The ice occupied its maximal position for a relatively short time; because of this, there was not adequate time to deposit recognizable moraines. 2) Recognizable moraines may have been deposited, but due to the moderate to steep slopes, the deposits were subject to mass wasting and severe erosion, thereby erasing evidence of their existence. 3) The deposits may have been too small to recognize. 4) The higher elevations may have been close to the ELA, causing less melting of ice and therefore less deposition. In all likelihood, a combination of these factors was at play.

Moraines of previous glaciations would be above the elevation of highest lastglacial erratics, but no such moraines were found there. There are at least three reasons why these Pleasant Mountain Glaciation moraines are not to be expected: 1) all reasons enumerated in the previous paragraph; 2) only moraines that were more extensive than subsequent glaciations stand a chance of preservation because older deposits are reworked (Gibbons et al., 1984); and 3) pre-lastglacial moraines, if they did exist in the

study area, would have significant degradation of their morphology due to the large amount of time that has passed since that glaciation, making them difficult to identify.

All moraines identified in the study area are interpreted as belonging to the Boorman Peak Glaciation recession. Two of the most significant moraines may correlate and illustrate how the ice split into two sub-lobes (Fig. 23). Between these lobes, it appears that there was likely a higher stand than the current level of Ashley Lake. This is supported by extensive areas of subdued topography caused by a drape of lacustrine deposits and postulated perched deltas up to an elevation of about 1280 m. These landforms may be worthy of further study in the future to determine their genesis. It is curious to note that 1280 m is also the elevation that Waitt (1985) and Alt (2001) have given for that maximum elevation reached by Glacial Lake Missoula. However, it is most likely that this was a higher stand of Ashley Lake dammed in by the ice lobe receding to the south. This lake probably drained through lateral meltwater channels and/or subglacially. It may have drained catastrophically to the Little Bitterroot drainage to the west although this possibility was not investigated in depth.

Erosional Landforms

All lateral meltwater channels and direct overflow channels identified were found at or below the Boorman Peak Glaciation maxima as defined by erratics and are therefore not useful in identification of multiple glaciations. Lateral meltwater and direct overflow channels, like depositional landforms, degrade over time and therefore can be difficult to detect. In addition, these landforms can be exploited by subsequent glaciations making it difficult, if not impossible, to date their inception. Only degraded pre-lastglacial features

that lie above subsequent glacial maxima can be interpreted from the record. No such features were identified.

Weathering Rinds

Relative dating by the use of weathering rinds failed to provide a detailed chronology. There are many reasons why weathering rind development may have failed to establish a relative dating criterion for the study area. Although the lithologies of clasts found in the study area have many favorable characteristics for rind formation such as being fine-grained, mechanically resistant, iron rich and have provided a criterion for establishment of rinds in other regions, several local variables may have confounded this effort.

Some clasts contained abundant cubes of pyrite. Pyrite is a very active oxidizer that produces sulfuric acid. As a result, pyrite cubes formed halos of oxidation around them. This oxidation often exploited bedding planes within the rock. These bedding planes then tended to fail preferentially, giving uneven weathering rinds (Pierce, 2002, verbal communication).

The relative alkalinity of soil and unconsolidated debris in the study area may have hindered rind development. Here, the alkalinity is due to the bedrock lithology, which is predominantly calcareous. Oxidation takes place more readily in acidic environments. Rind growth may have been inhibited for this reason. Alkalinity of soils varied widely in the study area. Soils below 1300 m effervesced wildly with the addition of dilute HCl. Soils above 1300 m effervesced mildly and not at all at higher elevations. Although it might be expected that higher areas may be more intensely leached of their carbonate due

an increase in precipitation with elevation (Richmond, 1962), there is probably not enough variation in elevation or aspect between these sites to explain the variation.

Although successful studies have been done on mafic and felsic rocks in other areas, the largely calcareous/felsic nature of the rocks in deposits within the study area probably hindered rind development. Silica rich rocks, such as many of those found in the study area, are relatively resistant to weathering compared with those that have more mafic minerals.

Lithologic heterogeneity between and within clasts can be large. As many as seven different lithologies were identified in any one rind pit. To establish a meaningful rating curve, clasts of similar composition must be used. Although limestone was the most numerous clast, it made up less than half the number of all those sampled. Rind pits in moraines were matrix supported, so a large amount of material had to be excavated to acquire enough clasts to allow appropriate statistical analyses. This made the process very inefficient. Intraclast variability is often gradational making determination of lithology difficult. Calcite blebs are abundant in some rocks. These blebs weather more readily than the surrounding rock. This variability causes uneven rind development. There was an apparent continuum between limestone and dolostone. Differentiation of lithology into two different groups is somewhat subjective in the field because it was based on the degree of effervescence with dilute HCl.

Although weathering rinds failed to establish a detailed chronology, their thickness suggests that they are from the Boorman Peak Glaciation. The modal measurement was 0.1 mm, indicating a relatively short period of time for rind formation. In addition, soils

in rind pits were weakly developed, providing further evidence for a relatively short period of time since they began to form.

Bog Augering

All augering took place in Boorman Peak Glaciation deposits. There were no ice marginal depressions at or above interpreted highest lastglacial ice positions as supported by studies of erratics. Ideally, Pleasant Mountain Glaciation and pre-Pleasant Mountain Glaciation closed depressions would have been augered. Closed depressions are generally absent on glacial deposits of this age. No features of this age were recognized, so no such work was done.

Tephra provided a minimum age for deglaciation for the study area. In augerings A1, A2 and A6 (Fig. 11), the Glacier Peak G ash was found showing that the ice had retreated from these localities by at least 11.2 ^{14}C ka. This is consistent with the tephra based results of other workers in the area (Carrara, 1989, Smith et al., 2000).

Using tephra's stratigraphic context within deposits, it is possible to extrapolate an age for the base of the sediment in the depression. Because tephra age is known and their depth was measured, it is possible to calculate an average sedimentation rate since deposition of the tephra. The average sedimentation rate can then be extended by extrapolation through to the base of the core, assuming a constant rate of sedimentation. While this assumption is probably not valid, the results may be informative. This procedure is a very rough approximation and has several confounding variables, one of which is that basal sediment in ice marginal depressions probably accumulates more rapidly than later sedimentation. Rapid basal sedimentation is due to at least two causes:

1) the close proximity of source sediment and water to transport it from the retreating glacier front; and 2) the relative lack of vegetated terrain at the time of deglaciation (Pierce, 2003, verbal communication). The results are summarized in Figure 24.

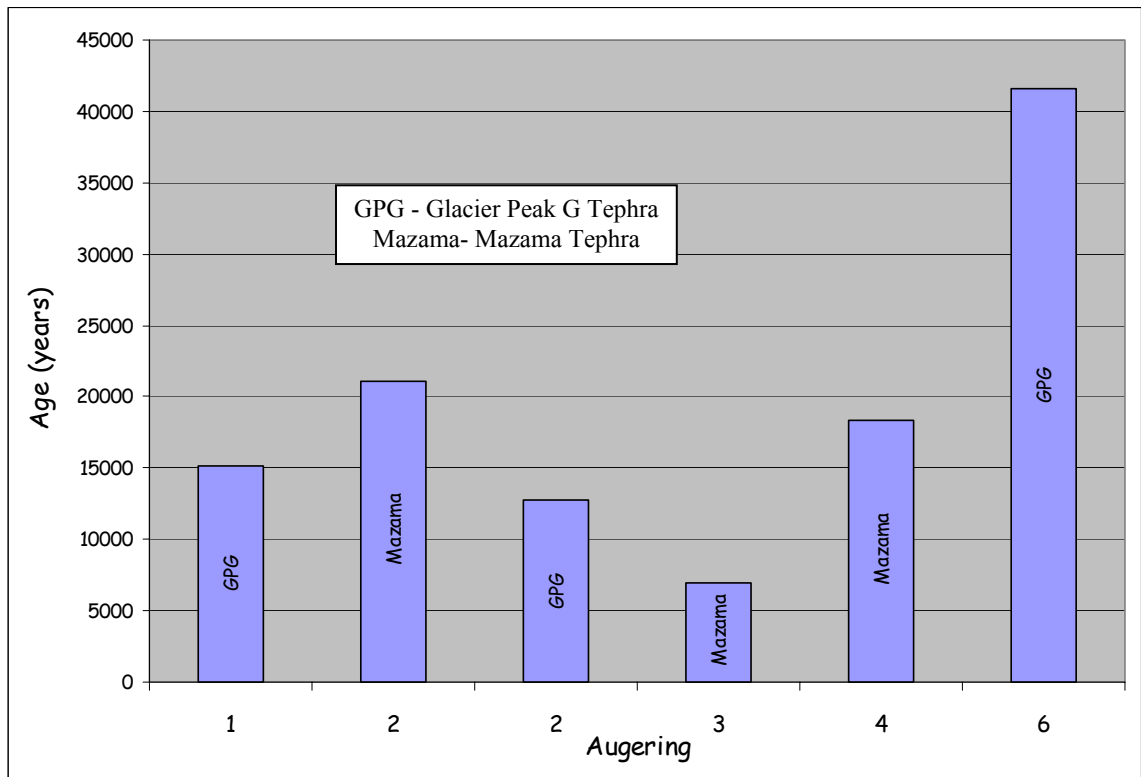


Figure 24 – Graph of inferred ages of bog bottoms based on sedimentation rates

Augerings 1 and 2 gave very reasonable minimum dates of deglaciation that roughly agree with other regional correlations. These estimates are based on uncorrected ^{14}C ages for the two tephras, and thus have an additional error when used for extrapolation. In augering 2, the Glacier Peak G tephra date matches other chronologies better than the date yielded by the Mazama tephra, which was found in the same augering. This is to be expected since the Glacier Peak G tephra was emplaced much closer to the time of deglaciation and would therefore logically yield a better date. Augerings 3, 4 and 6

yielded much less reasonable dates. The drainage above augering 3 may have only been subjected to more (intense) forest fires or other disturbances to the system after the emplacement of the tephra that skewed the sedimentation rate. Once again, it should be noted that the variable sedimentation immediately after deglaciation can have large effects on this type of extrapolation. Augering 4 may have continued to experience a large amount of deposition from a proglacial lake after the glacier retreated from the immediate area as evidenced by the thick varve package near the bottom. Augering 6 may have experienced significant infilling by sediment immediately after deglaciation due to the relatively large drainage area above it. While little confidence can be placed on these estimates for dating the precise retreat of the ice, the results do show that these bogs date from the last glaciation.

Since it is known that the lastglacial Flathead Lobe had retreated to at least 80 km north of its terminal position at Polson by the time the Glacier Peak G ash was deposited (11.2 ^{14}C ka), active ice would have already retreated from the Ashley basin before this time (Smith et al., 2000). Since there are no major moraines between the Ashley basin and the point mentioned above where Glacier Peak G ash is first found, it is hypothesized that the ice retreated relatively quickly and uniformly. This points to a time of deglaciation in the Ashley basin of not long before 12 ^{14}C ka. Augering 2 (Glacier Peak G) sedimentation rate dating suggests an age of deglaciation of approximately 13 ^{14}C ka. This date appears to be very reasonable and is likely a good approximation.

Regional Correlation

No correlation of the Pleasant Mountain Glaciation with other areas is attempted. Interpretations of many deposits in the region, such as the “Mission Moraine” of Richmond (1986), have numerous complications. The Mission Moraine has been reinterpreted by Levish (1997) as a lacustrine deposit rather than a glacial moraine and therefore no correlation to glacial features within the Ashley basin is needed. Other deposits in the region such as the “old moraine remnants” described by Richmond (1986) in the Thompson River Lobe have no numerical or relative ages assigned to them other than “pre-Wisconsin”. This leaves open the possibility that these deposits may be correlative with Illinoian deposits of the Laurentide Ice Sheet or deposits of an even older glaciation.

The Boorman Peak Glaciation maximum defined in the study area by erratics is correlated with the Polson Moraine (Richmond, 1986) 60 km to the southeast in the Flathead valley. While there is no moraine in the Ashley basin comparable with the size of the Polson Moraine, none is to be expected. At lastglacial maximum conditions in the Ashley basin, there were only isolated, medium-relief nunataks that are not very conducive for moraine deposition or preservation. Additionally, it was not the location of terminal glacier fronts. Therefore, it was not a major depositional center. The Polson moraine, on the other hand, was deposited in a wide, low relief valley. Some workers name the Polson moraine as the lastglacial terminal moraine and therefore a major depositional center; however, there are no exiting sets of reliable radiocarbon dates to support this conclusion.

The Boorman Peak Glaciation recession in the study area has no clear correlation with other areas. Although unpublished data by a team of University of Montana scientists indicates the presence of a recessional moraine in the bottom of the Big Arm of Flathead Lake (Hendrix, 2003, verbal communication), no correlation is attempted due to the present lack of a full disclosure of information. The next known recessional moraine of note as the Flathead Lobe retreated is the Kalispell Moraine (Richmond, 1986). It cannot be correlated with any features within the study area because it was deposited well after ice had retreated from the basin. This is shown by the valley geometry, which necessitate that the ice terminus would be near Kalispell only after it had left the Ashley basin. Therefore, major depositional and erosional features contained within the study area have no known correlative features in the main lobe of the Flathead to the southeast. If there is no correlation, why are there significant sets of depositional and erosional landforms in the Ashley basin? The significant stillstands of the advance and retreat of ice in the basin was primarily controlled by topography! As a result, the unique topography of the basin creates a unique pattern of ice advance and retreat.

CONCLUSION

The objective of this study was to determine the spatial and temporal attributes of the Flathead Lobe of the Cordilleran Ice Sheet by classic techniques of glacial geology. More specifically, these attributes were sought to define multiple advances of an ice sheet that have not been well defined. This was accomplished by first identifying features, then defining significant ice margins that created these features and finally, defining a glacial history.

It had been hypothesized that the Ashley basin was largely a nunatak during the lastglacial maximum and that if ice had intruded into the basin, its extent was probably minimal. After extensive fieldwork, this hypothesis is rejected. Numerous lines of evidence suggest that the basin was largely covered by ice during the last glaciation. Other evidence suggests that during an older glaciation the ice was even more extensive and probably covered the entire basin except the highest part of the western ridge, however only a minimum ice extent could be determined, not a maximum.

Interpreting lastglacial ice margins from depositional landforms alone in this study area was found to be inadequate. Numerous erratics and direct overflow channels show that the ice was at least 120 m above any recognized depositional landforms. Additionally, these erratics and direct overflow channels were approximately 240 m above depositional landforms that other workers had used to estimate glacial margins. This underestimation was probably the result of the reconnaissance nature of their studies. It is therefore probable that terminal margins of the last glaciation in this region have been under interpreted. Detailed fieldwork must be carried out to determine ice

extents. This fieldwork should include extensive work on erratics and small scale erosional landforms such as direct overflow channels, as indicated by this study. Topographic map and aerial photo analysis, modeling, and laboratory work, while informative, cannot accurately determine glacial extents in areas such as the Ashley basin on their own.

A local stratigraphy was developed and correlated to regional stratigraphies. The deposits relegated to each time division were separated by a major unconformity, primarily demonstrated by erratics, and depositional and erosional landforms. The well defined Boorman Peak Glaciation is interpreted as being approximately equivalent to late Wisconsin deposits of the Laurentide ice sheet and to Pinedale deposits of Rocky Mountain alpine glaciers. The next older glaciation is named the Pleasant Mountain Glaciation and is exclusively defined by the presence of scattered erratics. Based on the differences of identified lithologies and abundance of erratics, it is inferred that a long period of time elapsed between emplacement of the Boorman Peak and the Pleasant Mountain deposits. Pleasant Mountain Glaciation deposits, because they occur significantly above Boorman Peak Glaciation features, are interpreted as being correlative with the Illinoian deposits of the Laurentide ice sheet and to Bull Lake deposits of Rocky Mountain alpine glaciers. Due to a lack of numerical age control, the possibility that the Pleasant Mountain Glaciation is correlative with an even older glaciation must be presently considered.

Topography is the key to the style of deglaciation in this study area. The moderate-relief topography and shallow ice gradient combine to exert the main controls on the ice here. Despite the generally accepted idea that the Flathead Lobe advanced from the north,

the center of the study area was the first to be deglaciated due to local effects of topography. The two main sub-lobes of ice that existed in the Ashley Lake valley had different controls on their retreat. The Ashley Piedmont Lobe, effectively an outlet glacier, entered from the north and east where it spilled over the high drainage divide into the Ashley basin. The Ashley Creek Lobe entered the valley from the south and faced no major obstacles other than an uphill gradient.

The information gained from the Ashley basin in this study can be used to further understand the surrounding glacial history (Fig. 25). The extent of the last glaciation of the Cordilleran Ice Sheet in this area has now been determined by detailed fieldwork and found to be several hundred meters higher than previously mapped by Richmond (1986). The precise timing has still not been derived due to the lack of materials suitable for radiocarbon dating and the failure of relative weathering techniques to differentiate deposits. Through detailed fieldwork *and* laboratory analysis of the Ashley Lake study area, past glaciations of the Cordilleran Ice Sheet have been more accurately defined.

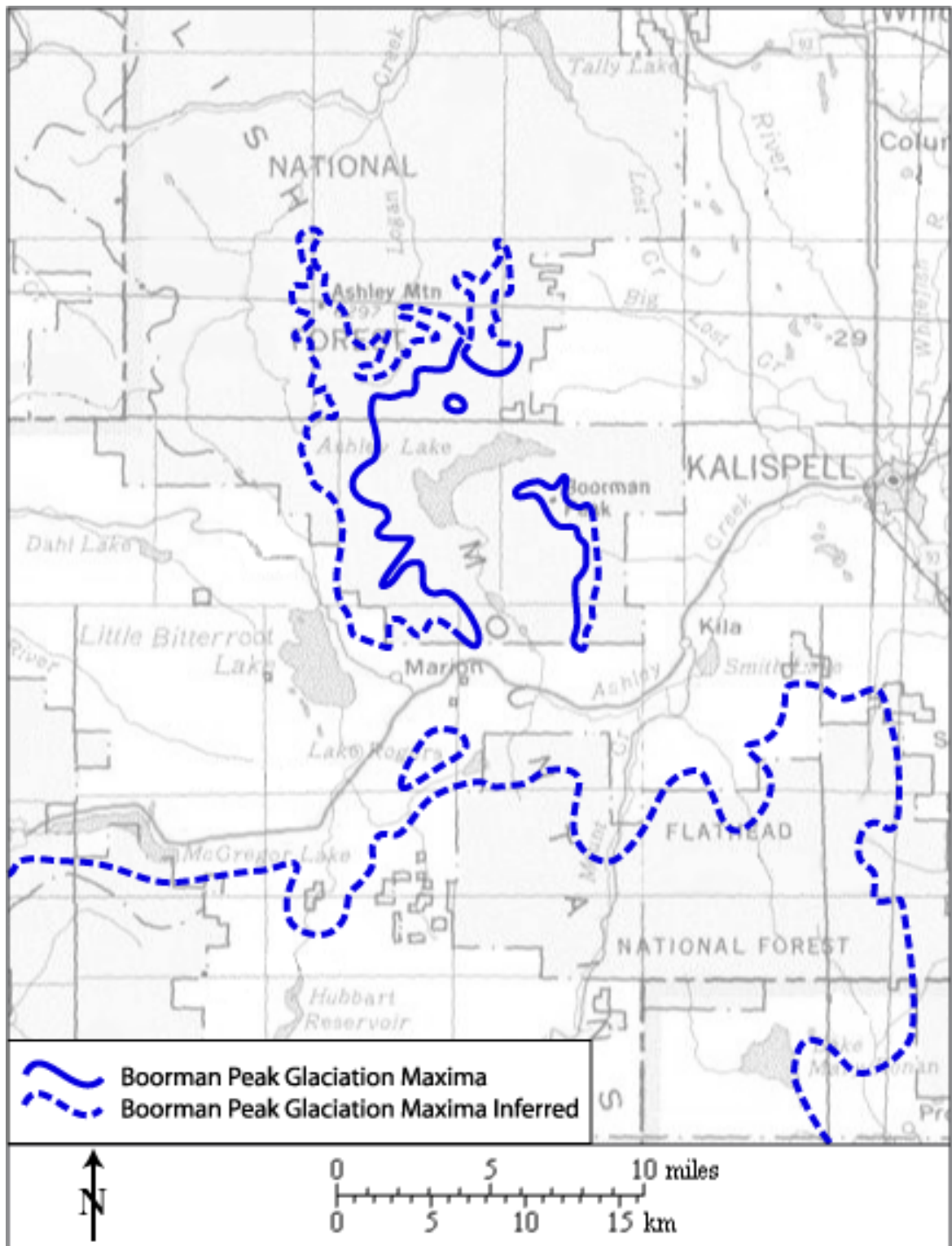


Figure 25 – Inferred regional Boorman Peak Glaciation maxima

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APPENDICES

APPENDIX A

WEATHERING RIND STATISTICS

Ashley Kame

<u>Clast #</u>	<u>Mode(mm)</u>	<u>Argillite</u>		<u>Gray Quartzite</u>	<u>Purple Quartzite</u>
		<u>Gray Limestone</u>	<u>Gray Dolostone</u>		
1	5.1	X			
2	0.1			X	
3	0.1	X			
4	0.5	X			
5	4.4	X			
6	1.6	X			
7	2.5		X		
8	0.1	X			
9	0.1			X	
10	1.6	X			
11	1.9			X	
12	0.1	X			
13	0.1			X	
14	0.1			X	
15	2.2		X		
16	0.1			X	
17	0.1			X	
18	1.5		X		
19	2.0		X		
20	0.3	X			
21	0.1				X
22	0.1		X		
23	2.5			X	
24	0.1				X
25	1.1			X	
26	0.1			X	
27	0.5	X			
28	0.1	X			
29	0.8	X			
30	0.4			X	
31	0.1	X			
32	0.1	X			
33	0.1				X
34	0.1	X			
35	1.6	X			
36	0.3		X		
37	3.9		X		
38	1.1	X			
39	1.1			X	
40	0.6	X			
41	0.1	X			
42	0.1			X	
43	0.7	X			

Brist Kame

<u>Clast #</u>	<u>Mode(mm)</u>	<u>Argillite</u>		<u>Gray</u>	<u>Green</u>	<u>Gray</u> <u>Quartzite</u>	<u>Purple</u> <u>Quartzite</u>
		<u>Gray</u> <u>Limestone</u>	<u>Gray</u> <u>Dolostone</u>				
1	1.7		X				
2	0.2			X			
3	1.0	X					
4	2.8	X					
5	2.0	X					
6	0.1				X		
7	0.1			X			
8	4.5		X				
9	0.1	X					
10	0.1	X					
11	1.7		X				
12	0.1			X			
13	0.1	X					
14	0.8	X					
15	0.1	X					
16	0.1			X			
17	0.0						X
18	0.1	X					
19	0.4	X					
20	1.0		X				
21	0.1			X			
22	0.1	X					
23	2.1	X					
24	0.1	X					
25	0.1	X					
26	0.2	X					
27	0.1	X					
28	0.1	X					
29	0.1			X			
30	2.6			X			
31	2.7			X			
32	0.1	X					
33	0.1	X					
34	1.7	X					
35	0.1	X					
36	0.1						X
37	0.1	X					
38	0.1	X					
39	1.1			X			
40	1.1		X				
41	2.4	X					
42	1.9		X				
43	0.1	X					
44	0.1	X					
45	0.1		X				
46	0.8		X				
47	0.1	X					

Gamma Moraine

<u>Clast #</u>	<u>Mode(mm)</u>	<u>Gray Limestone</u>	<u>Gray Dolostone</u>	<u>Argillite</u>		<u>Gray Quartzite</u>	<u>Purple Quartzite</u>
				<u>Gray</u>	<u>Green</u>		
1	0.1	X					
2	0.1					X	
3	0.2	X					
4	0.1	X					
5	0.2	X					
6	0.2	X					
7	0.3	X					
8	0.7	X					
9	0.4		X				
10	2.2	X					
11	0.2	X					
12	0.3	X					
13	2.2		X				
14	3.8	X					
15	0.8			X			
16	0.1			X			
17	0.5	X					
18	2.1					X	
19	2.3	X					
20	0.0			X			
21	1.0	X					
22	0.0			X			
23	1.9	X					
24	0.1			X			
25	4.9		X				
26	1.6			X			
27	1.3	X					
28	6.5	X					
29	4.9			X			
30	0.1						X
31	0.1					X	
32	0.1						X
33	3.9			X			
34	0.0				X		
35	4.9	X					
36	0.0					X	
37	1.6	X					
38	0.1	X					
39	0.1	X					
40	0.1	X					
41	0.1				X		
42	4.4			X			
43	1.1			X			
44	6.3	X					
45	4.2	X					
46	5.2	X					
47	0.4	X					
48	1.8		X				
49	1.0		X				
50	0.1	X					

Hammerhead Moraine

<u>Clast #</u>	<u>Mode(mm)</u>	<u>Gray Limestone</u>	<u>Gray Dolostone</u>	<u>Argillite</u>		<u>Brown Quartzite</u>	<u>Brown Siltstone</u>
				<u>Gray</u>	<u>Green</u>		
1	2.0	X					
2	4.4			X			
3	2.0			X			
4	1.7					X	
5	5.4		X				
6	0.1				X		
7	0.1				X		
8	5.5	X					
9	4.6					X	
10	4.3					X	
11	4.2	X					
12	2.5		X				
13	0.6					X	
14	1.9					X	
15	0.0					X	
16	2.9			X			
17	2.2			X			
18	6.7					X	
19	3.4			X			
20	5.3			X			
21	0.0				X		
22	5.3		X				
23	3.8						X
24	8.2						X
25	0.1				X		
26	8.2	X					
27	0.1				X		
28	3.1					X	
29	4.0			X			
30	5.2						X
31	3.8			X			

Hodge Moraine

<u>Clast #</u>	<u>Min(mm)</u>	<u>Mode(mm)</u>	<u>Max(mm)</u>	<u>Gray Limestone</u>	<u>Gray Dolostone</u>	<u>Brown siltstone</u>
1	0.4	2.0	2.4		X	
2	0.5	3.3	3.9		X	
3	1.2	1.8	2.0		X	
4	7	12.0	16.6		X	
5	2.4	10.8	15.9		X	
6	7.2	10.0	12.3		X	
7	6.2	6.4	7.8		X	
8	9.1	10.1	11.9	X		
9	2.5	3.1	3.3	X		
10	4	5.3	6.5		X	
11	3	7.0	8.2			X
12	3.2	5.1	6.2	X		
13	0.8	1.2	1.8	X		
14	8.5	13.8	19.9	X		

Monroe Moraine

<u>Clast #</u>	<u>Mode(mm)</u>	<u>Gray Limestone</u>	<u>Gray Dolostone</u>	<u>Argillite</u>		<u>Gray Quartzite</u>	<u>Purple Quartzite</u>
				<u>Gray</u>	<u>Green</u>		
1	2.0	X					
2	0.1				X		
3	0.1					X	
4	0.1				X		
5	0.1				X		
6	2.4	X					
7	0.1				X		
8	1.8	X					
9	1.0	X					
10	2.5	X					
11	0.1				X		
12	0.1				X		
13	0.1				X		
14	0.1				X		
15	0.1				X		
16	2.2	X					
17	0.1						X
18	0.1					X	
19	0.9				X		
20	0.1					X	
21	0.7				X		
22	0.1				X		
23	3.4		Did not indicate				
24	0.1					X	
25	3.9			X			
26	0.1					X	
27	0.1				X		
28	0.1						X
29	3.5			X			
30	0.1				X		
31	5.7			X			
32	0.1			X			
33	0.1					X	
34	0.1				X		
35	0.1						X
36	0.1					X	
37	0.1				X		
38	0.1				X		
39	0.1					X	
40	0.1				X		
41	0.1						X
42	0.1				X		
43	0.6	X					
44	1.5	X					
45	0.1				X		
46	0.1				X		
47	0.1				X		
48	0.1					X	
49	0.1			X			
50	0.1			X			
51	0.1				X		
52	0.1			X			
53	0.8				X		
54	0.1			X			
55	0.1				X		
56	0.1			X			

Rand Moraine

<u>Clast</u> <u>#</u>	<u>Mode</u> <u>(mm)</u>	<u>Gray</u> <u>Limestone</u>	<u>Gray</u> <u>Dolostone</u>	<u>Argillite</u>		<u>Gray</u> <u>Quartzite</u>	<u>Purple</u> <u>Quartzite</u>	<u>Brown</u> <u>Quartzite</u>
				<u>Gray</u>	<u>Green</u>			
1	1.5		X					
2	3.9					X		
3	0.1					X		
4	0.1						X	
5	4.9					X		
6	0.0						X	
7	7.6	X						
8	1.1							X
9	13.0	X						
10	0.1			X				
11	7.6	X						
12	4.0	X						
13	0.1						X	
14	7.3	X						
15	2.5			X				
16	2.8	X						
17	5.1	X						
18	0.1						X	
19	0.1			X				
20	0.1				X			
21	0.1			X				
22	0.1					X		
23	4.8		X					
24	5.5	X						
25	0.1				X			
26	14.4	X						
27	0.1					X		

APPENDIX B

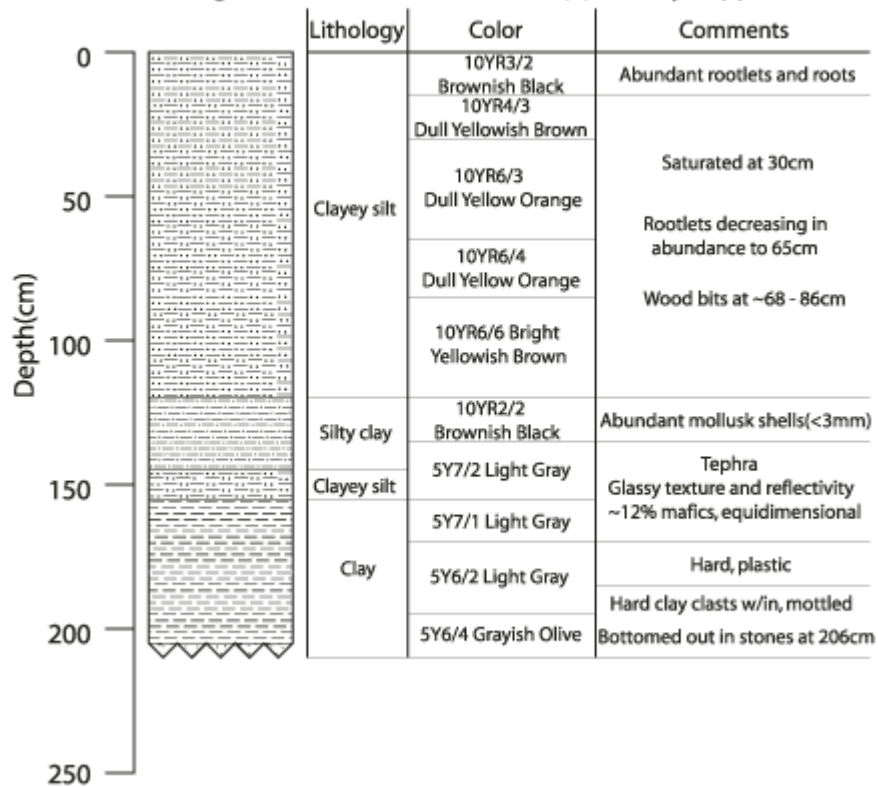
AUGERING LOGS

Augering: 1

Date: 07/06/02

Location: Hodge Creek

Worker(s): Denny Capps, Dr. Locke



Augering: 2

Date: 07/12/02

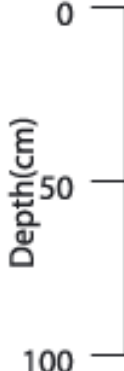
Location: Gamma Depression

Worker(s): Denny Capps

Depth(cm)	Lithology	Color	Comments
0	Clayey Silt	10YR5/6 Yellowish Brown	Roots, rootlets to ~57cm
50		10YR6/4 Dull Yellow Orange	Tephra Becoming lighter in color with some sand
100		10YR5/4 Dull Yellowish Brown	Light gray silt lens at ~73cm 1cm light gray silt lens at ~ 78cm Water saturated at 111cm
150		10YR3/4 Dark Brown	Damp, packs well through 163cm Less clay at 163cm
200	Silt	2.5Y7/3 Light Yellow	Tephra, saturated, grades into above, very plastic
	Clayey Silt	2.5Y5/3 Yellow Brown	Somewhat plastic
250	Till	2.5Y6/3 Dull Yellow	Sandy silt matrix

Augering: 3
Location: Twin Bogs

Date: 07/16/02
Worker(s): Denny Capps

Depth(cm)			Lithology	Color	Comments
			Silt	10YR2/3 Brownish Black	Full of rootlets Abrupt transition at 35cm
				10YR4/6 Brown	Sandy silt with rootlets
				7.5Y8/1 Light gray	Tephra, saturated, rootlets
			Till	7.5Y2/1 Black	Stones in sandy silt matrix

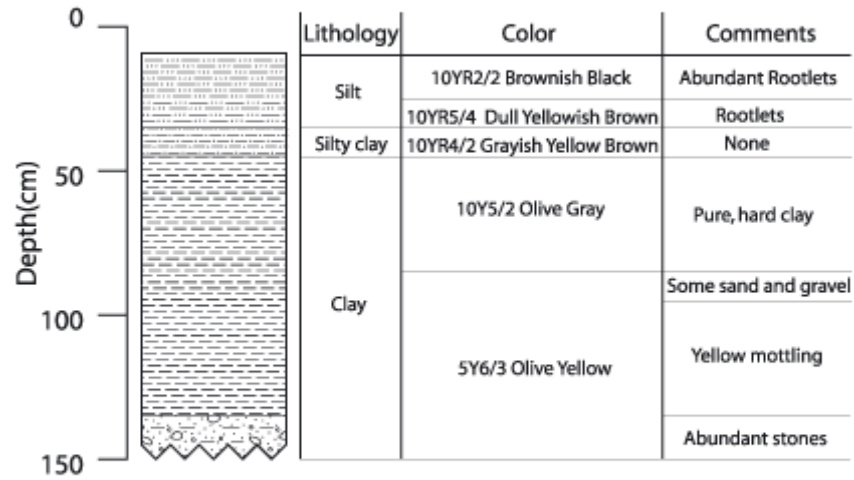
Augering: 4
Location: Drainage Exit Bog

Date: 07/17/02
Worker(s): Denny Capps

Depth (cm)	Lithology	Color	Comments
0	Silt	7.5YR2/2 Brownish Black	Rootlets! Abundant organic matter
50	Clayey silt		
100	Silt		Rootlets, decayed organic matter
150	Silty clay	10YR3/2 Brownish Black	Abundant mollusk shells
		10YR3/1 Brownish Black	
200	Silt	2.5GY6/1 Olive gray	Mazama tephra Very pure silt
250	Silty clay	10YR4/2 Grayish Yellow Brown	Mollusk shells abundant
		10YR4/1 Brownish Gray	Almost pure clay at top. Significantly more mollusks shells in lowest 3cm
300		10YR5/1 Brownish Gray	Some mollusk shells
350		7.5GY6/1 Greenish Gray	Varved
400		10GY5/1 Greenish Gray	
450		10GY6/1 Greenish Gray	
		7.5GY6/1 Greenish Gray	Mottled with darker green
500		10GY5/1 Greenish Gray	With abundant stones

Augering: 5
Location: Gamma Moraine

Date: 07/19/02
Worker(s): Denny Capps
Scott Larson



Augering: 6
Location: Twin Section 36

Date: 08/03/02
Worker(s): Denny Capps

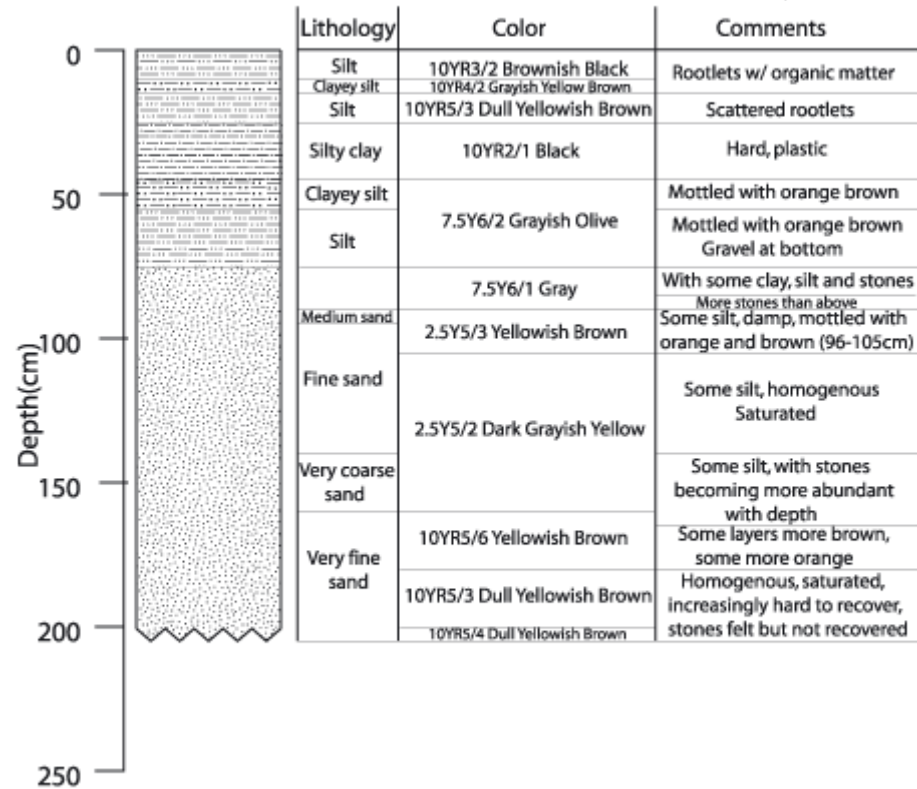
Depth(cm)	Lithology	Color	Comments
0	Silt	10YR3/1 Brownish Black	Abundant roots and rootlets
			Rootlets
			Abrupt transition at bottom
		10YR7/3 Dull Yellow Orange	Orange mottling, dry, crumbly
		10YR6/4 Dull Yellow Orange	Rootlets
50	Clayey silt	10YR6/3 Dull Yellow Orange	None
		10YR3/3 Dark Brown	Damp
		10YR4/2 Dark Brown	Damp, more clay towards bottom
100	Silt	7.5Y8/1 Light Gray	Tephra, grades into above
150	Clayey silt	10Y6/1 Gray	Abrupt transition with above
			Mottled
			Clasts up to 1.5cm
			Mottled with orange
			Mottled with brown orange
			More intense mottling
			More intense mottling
200	Silty clay	7.5Y6/1 Gray	Slightly more clay
			Less mottling
			None
			Mottled, more clay at bottom
			some small gravels
250	Sand	5Y6/2 Grayish Olive	Mottled brown orange
			Brown orange layers
			Silty, saturated
300	Silt	2.5Y5/3 Yellowish Brown	Fine sand and clay, Tephra 362-363cm? Less solid
350	Clayey silt	5Y6/1 Gray	Variable colors, some sand
			Brown orange mottled, Stones in last 2cm
400	Sand	5Y4/3 Dark Olive	Some mottling
450	Till	5Y3/1 Olive Black	Heterolithic, sub-angular to sub-rounded
500			Dolomite stones up to 3cm, sub-angular, saturated

Augering: 7

Date: 08/06/02

Location: South Lake Tip

Worker(s): Denny Capps



APPENDIX C

TEPHRA CHARACTERIZATION WORKSHEETS

Tephra characterization worksheet

Site: Augering # 2

Date: 07/12/02

Category	Property	Observation
Field Characteristics	Thickness	1 cm
	Bedding characteristics	Two lenses 5 cm apart, sampled lowest
	Grain size	Silt
	Color	Light Gray
	Stratigraphic context	73 and 78 cm, 229 cm total depth
	Grittiness	Very sharp and angular
	Reflectivity	None noted
Lab Characteristics	Percentage mafic crystals	~2-3%
	Degree of hydration	Low
	Morphology of glass shards	Very needle-like with abundant bubble-wall and bubble-wall junctions

Probable Tephra - Mazama

Tephra characterization worksheet

Site: Augering # 2

Date: 07/12/02

Category	Property	Observation
Field Characteristics	Thickness	17 cm
	Bedding characteristics	Gradational upper boundary
	Grain size	Silt
	Color	Light Yellow (2.5Y7/3)
	Stratigraphic context	186-203 cm, 229 cm total depth
	Grittiness	None noted
	Reflectivity	Glassy
Lab Characteristics	Percentage mafic crystals	~15%
	Degree of hydration	Medium-High
	Morphology of glass shards	Mostly equidimensional, some elongated forms

Probable Tephra - Glacier Peak G

Tephra characterization worksheet

Site Location: Augering # 1

Date: 07/06/02

Category	Property	Observation
Field Characteristics	Thickness	9 cm
	Bedding characteristics	No special bedding noted
	Grain size	Clayey-silt
	Color	Light Gray (5Y7/2)
	Stratigraphic context	146-155 cm interval, 206 total depth
	Grittiness	Very sharp and angular
	Reflectivity	Sparkled
Lab Characteristics	Percentage mafic crystals	~12%
	Degree of hydration	Medium
	Morphology of glass shards	Mostly equidimensional, some bubble-wall and bubble-wall junctions, scattered cylinders

Probable Tephra - Glacier Peak G

Tephra characterization worksheet

Site Location: Augering #4

Date: 07/17/02

Category	Property	Observation
Field Characteristics	Thickness	50 cm
	Bedding characteristics	Grain size change at upper and lower boundary
	Grain size	Silt
	Color	Olive Gray (2.5GY6/1)
	Stratigraphic context	156-206 cm interval, 467cm total depth
	Grittiness	Very sharp and angular
	Reflectivity	None noted
Lab Characteristics	Percentage mafic crystals	~2%
	Degree of hydration	Low-medium
	Morphology of glass shards	Commonly needle-like, bubble-wall and bubble-wall junctions

Probable Tephra - Mazama

Tephra characterization worksheet

Site Location: Augering # 3 **Date:** 07/16/02

Category	Property	Observation
Field Characteristics	Thickness	19 cm
	Bedding characteristics	No special bedding noted
	Grain size	Silt
	Color	Light Gray (7.5Y2/1)
	Stratigraphic context	48-67 cm, total depth 68 cm
	Grittiness	None noted
	Reflectivity	None noted
Lab Characteristics	Percentage mafic crystals	~7%
	Degree of hydration	Low
	Morphology of glass shards	Equidimensional and elongate form common, bubble-wall and bubble-wall junctions

Note: Very little material left after sieving, may have not been representative

Probable Tephra - Mazama?

Tephra characterization worksheet

Site Location: Augering # 6 **Date:** 08/03/02

Category	Property	Observation
Field Characteristics	Thickness	8 cm
	Bedding characteristics	Abrupt transition at bottom, gradational at top
	Grain size	Silt
	Color	Light Gray (7.5Y8/1)
	Stratigraphic context	122-130 cm interval, 461 cm total depth
	Grittiness	Very sharp and angular
	Reflectivity	None noted
Lab Characteristics	Percentage mafic crystals	~15%
	Degree of hydration	Medium
	Morphology of glass shards	Mostly equidimensional, some elongate

Probable Tephra - Glacier Peak G