Late Wisconsin Glacier Surface Elevations and Flow Directions from the Beartooth Plateau to the Clarks Fork Valley, Wyoming

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Bachelor of Science in Geology

by
Katharine C. Adams

Department of Geology
Washington and Lee University
Lexington, VA 24450
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LATE WISCONSIN GLACIER SURFACE ELEVATIONS AND FLOW DIRECTIONS FROM THE BEARTOOTH PLATEAU TO THE CLARKS FORK VALLEY, WYOMING

By KATHARINE C. ADAMS

ABSTRACT

During the Pinedale glaciation of the Late Wisconsin, an ice cap covered most of the land comprising the Beartooth Plateau and the highlands to the south, stretching to the Clarks Fork Valley in northwestern Wyoming. The upper limits or this icemass and flow directions during full glacial conditions were determined from field observation of nunataks, highest glaciated uplands, glacial striations, and streamlined bedrock forms. On the southern portion of the Beartooth Plateau, ice possibly reached as far as 11,100 ft (3,383 m), with its surface elevation gradually falling to 8,800 ft (2,682 m) upon reaching the Clark's Fork outlet glacier. Ice flow from the Beartooth Plateau followed a radial pattern heading southwest in the western portion of the field area, south in the central portion, and southeast in the eastern portion. Some glacial striations indicated valley glacier flow constrained by topography in contrast to ice cap flow during full-glacial conditions. Basal shear stress calculations support the constructed model, demonstrating it as physically possible. Averaging 0.7 bars, the values for most of the icemass fell within basal shear stress limits of modern glaciers.

INTRODUCTION

At its maximum, Late Wisconsin glaciation of North America resulted in ice coverage stretching from the Atlantic to the Pacific. The continuous span of ice was divided into two glaciers: the Laurentide Ice Sheet, which covered the northeastern and north-central portions of North America, and the Cordilleran Glacier Complex which occupied the northwest (Flint, 1971). The Cordilleran Glacier Complex only reached as far south as the Columbia Plateau in Washington, but south of this limit of continuous ice, at least 75 separate glacial areas were centered on highlands (Flint, 1971). These highlands, the high alpine ranges in the western United States, included the near-coastal Pacific Mountain system and the Interior Rocky Mountain system (Porter and others, 1983). Instead of forming massive ice sheets, the glaciers situated in these mountains and plateaus were either valley glaciers, groups of valley glaciers, complexes, or ice caps.

Of the total volume of ice situated in the Rocky mountain glacial complexes, most was contained in large ice caps and ice fields, rather than in valley glaciers. The mountains in northwestern Wyoming and adjacent states were home to one such glacial complex. These mountains and plateaus comprise the Yellowstone - Teton - Wind River highlands. Within them lies "the largest single area of former glaciation" in the western United States (Pierce, 1979).

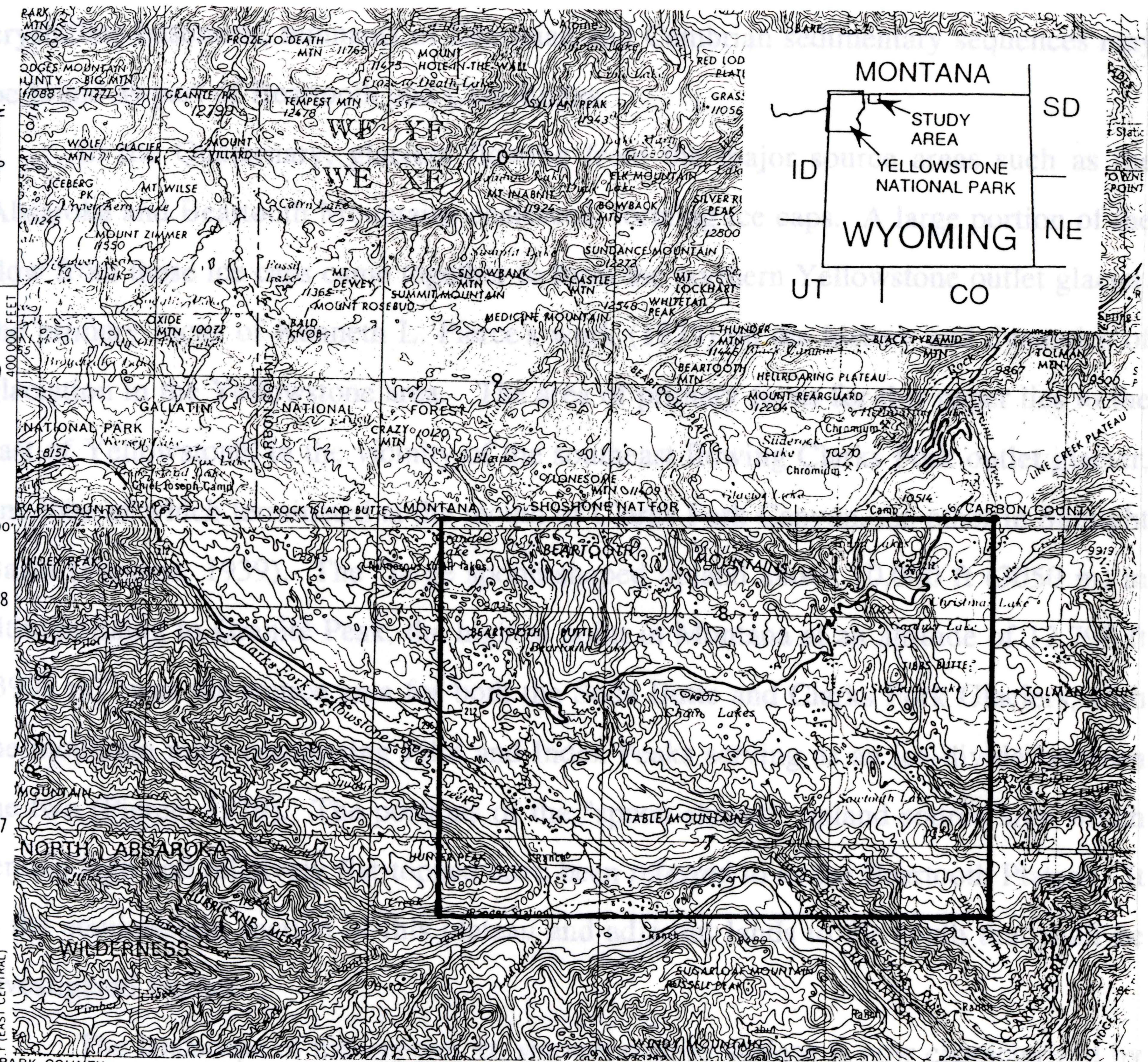

Figure 1. Index Map of Wyoming showing study area.

REGIONAL GEOLOGY AND TOPOGRAPHY

The Beartooth Plateau is a part of these highlands which had an ice cap of its own during the Late Wisconsin (Flint, 1971). Located on the border of Montana and Wyoming, northeast of Yellowstone National Park, the plateau is part of the Beartooth uplift, which is "a large, relatively coherent structural and geographic block" formed during the Laramide orogeny (Pierce, 1979). The block is composed of Archean granite and gneiss, bounded by faulting and folding on the north and east edges. Overlying the crystalline basement, remnants of Devonian and Cambrian sedimentary sequences may be found as part of Beartooth and Clay Buttes.

In the Quaternary, glaciers flowing from the major source areas such as the Absaroka and Beartooth Mountains coalesced, forming ice caps. A large portion of the flow from these ice caps came together to form the northern Yellowstone outlet glacier, the primary focus of Kenneth L. Pierce's work (1979) on the history and dynamics of glaciation in the Yellowstone area. The area of primary focus for this paper lies to the east of Yellowstone, in the vicinity of the southeast flowing Clarks Fork outlet glacier, one lobe of which terminated at the mouth of Clarks Fork Canyon, the other in Sunlight Basin (Parsons, 1939). The "large spoon-shaped upland above 10,000 ft (3050 m) in altitude rising to Granite Peak, the highest point in Montana at an altitude of 12,799 ft (3901 m)" was the source area for both the Soda Butte and Clarks Fork Glaciers, with the mountain mass comprising Pilot and Index Peaks serving as an ice-divide between the two (Pierce, 1979). The portions of the "spoon-shaped" upland source area which remain high and relatively flat-topped have been referred to as the Beartooth Plateau. It is the southeastern section of this plateau and adjacent lands to the south that shall be investigated further.

The landscape of the uppermost, smooth plateau surface is now characterized by periglacial landforms such as tors and patterned ground, and is virtually devoid of any glacial erosional forms. In sharp contrast to this topography, the northeastern edge of the plateau ends abruptly (see Plate 1 or Figure 3). Carved out by Rock Creek ice and glaciers originating in the deep cirques along the plateau edge, the elevation of the land plunges approximately 2600 ft (790 m) over a distance of about 1 kilometer to the Rock Creek valley floor. On the southeastern side, the change is not so drastic. The elevation decreases by 4600 ft (1402 m) over a distance of about 16 kilometers from the high edge of the plateau to a portion of the Clarks Fork valley floor. This broad, glacially scoured expanse from the Beartooth Plateau to the Clarks Fork constitutes the bulk of the study area. Bordering the plateau's uppermost surface, it is riddled with cirques, tarns, U-shaped valleys, roches moutonnées, glacially polished bedrock, striations, and ice-transported boulders and erratics.

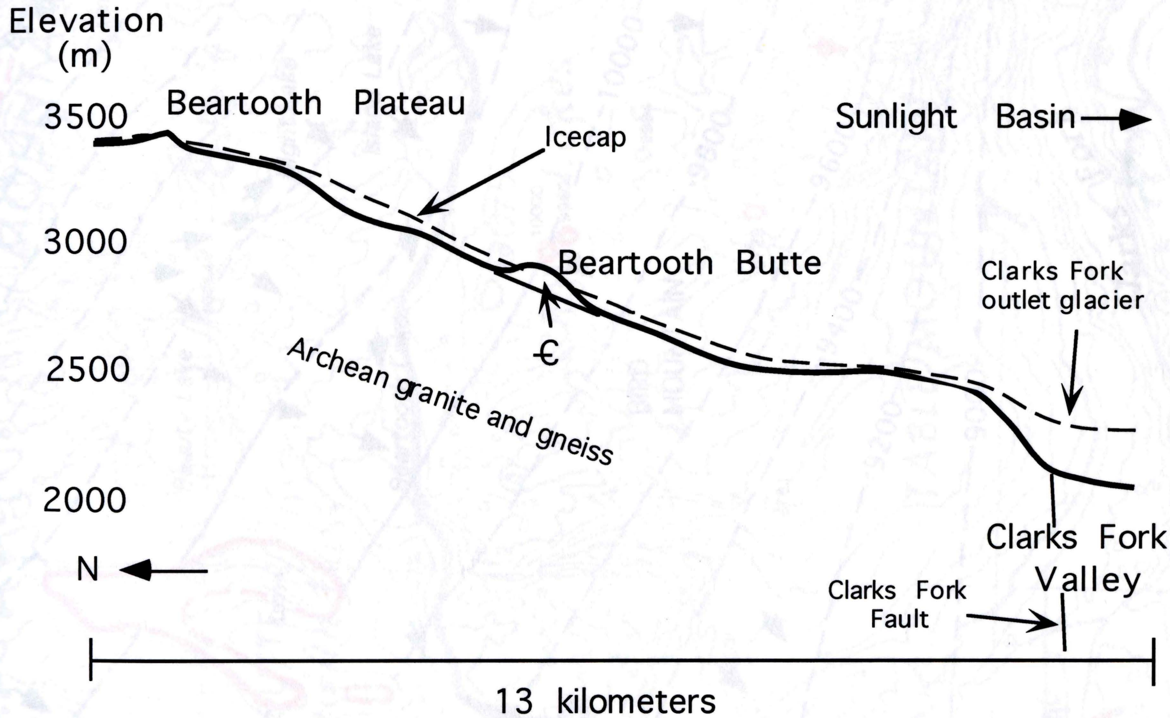

Figure 2. Diagrammatic Cross Section of the Clarks Fork Region.

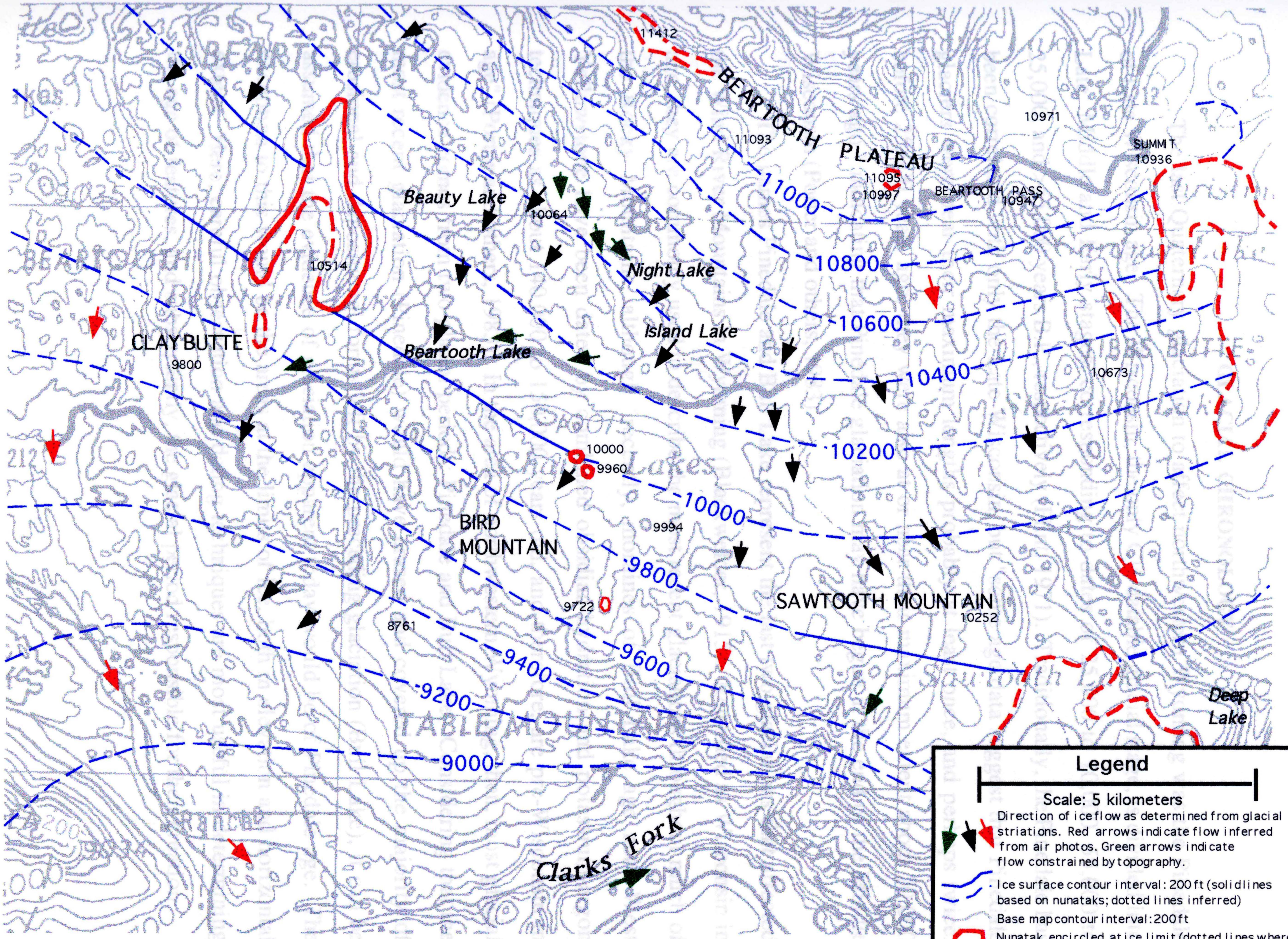

Figure 3. Pinedale Icemass Reconstructed at Full-Glacial Conditions, from the Beartooth Plateau to the Clarks Fork

GLACIAL CHRONOLOGY

The last Quaternary glaciation to affect northwestern Wyoming was the Pinedale of Late Wisconsin age. The Wisconsin, the only C^{14} datable glacial stage, is divided into Late, Middle and Early stages, corresponding to dates of 25,000 to 10,000, 55,000 to 25,000, and >55,000 years B.P., respectively (Flint, 1971). Traditionally, the Pinedale has been correlated with only the Late Wisconsin, but "some age data suggest that Pinedale glaciers were near their maximum lengths probably at least once, and perhaps twice, before 25,000 years ago; they were also at or near their maximum extent about 20,000 years ago," indicating that Pinedale glaciation may have begun at the end of the Middle Wisconsin (Porter and others, 1983).

The names Pinedale and Bull Lake represent the last two glaciations that affected the Wind River Mountains in Wyoming (Blackwelder, 1915). Regarded as Wisconsin in age, they are commonly used to describe the most recent glaciations throughout much of the Rocky Mountains, although "recently, moraines of Bull Lake age near West Yellowstone have been dated by combined obsidian-hydration and potassium-argon techniques as about 140,000 to 150,000 years old, and therefore, are pre-Wisconsin in age" (Porter and others, 1983). In the area of northwestern Wyoming, then, most glacial evidence comes from two glacial periods: Pinedale and Bull Lake. Considering that the most recent glaciation (Pinedale) would have erased and covered up the majority of erosional and depositional forms left by the preceding glaciation (Bull Lake), it can be assumed that the evidence of glaciation apparent today would be of Pinedale age. The exception to this would be the areas at the limits of ice, both in elevation and horizontal extent. Here, careful observation and dating techniques are most useful in determining the age of features, and subsequently, the glacial period responsible for their creation.

DISCUSSION OF FIELD METHODS

DETERMINING DIRECTIONS OF GLACIAL FLOW

Ice flow directions may be used to determine whether ice flow is topographically constrained or unconstrained. Topographically constrained ice is usually representative of valley glaciers, whereas unconstrained flow is a distinguishing characteristic of sheet flow of ice caps. Depending on location and elevation, differing flow directions could show the transition from sheet flow to stream flow in an ice cap complex.

The following paragraphs describe methods by which former ice flow directions may be determined. The list is not meant to be comprehensive, it only describes the methods utilized for the purposes of this paper.

Glacial Striations: Glacial striations are scratches and grooves on bedrock protuberances formed as a result of the sliding action of a glacier over the rock surface. They are found in previously glaciated areas on bedrock surfaces and clasts, especially those protected from the effects of weathering. Well-preserved striations, consequently, are often found protected beneath perched boulders. Striations may indicate the line of flow, but not necessarily the direction of ice movement.

Since striations may show both regional and local flow patterns, other factors must be considered when determining which pattern is being exhibited. Striations found on high peaks and ridges most likely reflect regional ice flow and also parallel the slope of the ice surface, whereas ones located in valleys and oriented parallel to valley walls are a result of ice flow constrained by topography (Pierce, 1979). When these two types of striations are found at high and low elevations of the same locale, "the higher striations indicate an earlier flow direction and the lower ones a later direction" (Pierce, 1979). Thus, they may indicate regional flow during a glacial maximum, followed by waning flow confined to valley walls.

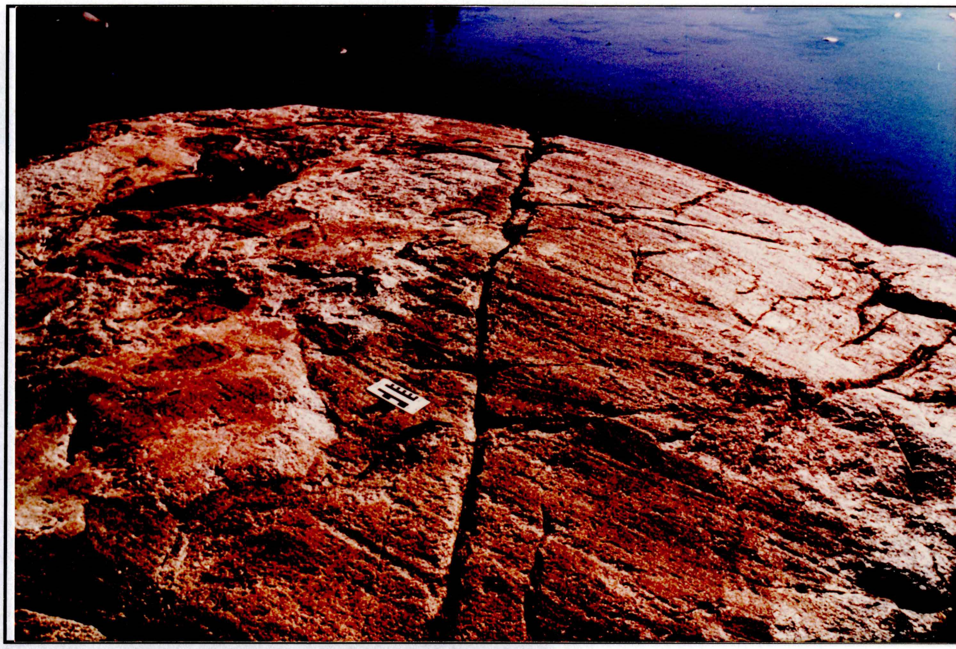

Figure 4. Glacial Polish and Striations on a Granite Outcrop near Long Lake.

Roches Moutonnées: Roches moutonnées "comprise a family of partly streamlined forms generally regarded as the hallmark of glacial erosion" (Sugden, 1976). The term describes a streamlined, asymmetrical hill, that is molded on the upstream side, and craggy and broken off on the lee side. As glaciers pass over small hills or outcrops of bedrock, the basal ice tends to melt on the upstream side of the hill due to the increase in pressure. The ice re-freezes on the lee side of the hill, and in the process, plucks pieces of rock off the downstream side of the bedrock protuberance, causing the asymmetrical shape. "Most descriptions of roches moutonnées in the literature suggest that the features are best developed on well jointed granites and other crystalline rocks" (Sugden, 1976). Long after the glacial polish and striations wear away, the ice flow direction may still be inferred from the orientation of the long axis of a roches moutonnée.

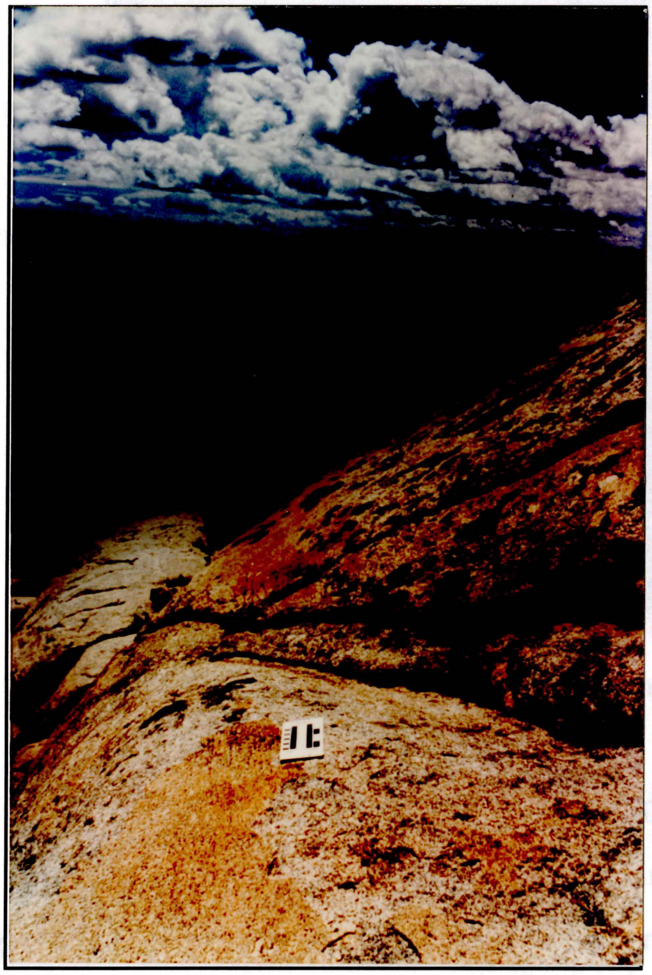

Figure 5. Streamlined roches moutonnée northwest of Hauser Lake (in background).

Ice Transported Boulders: Ice transported boulders can be any clast that has been transported and deposited by moving ice. An erratic is an ice transported boulder of a different lithology than the underlying bedrock. Ice transported boulders are used only to "indicate net transport direction and are not used to indicate flow direction during any specific stage of glaciation unless supported by other evidence" (Pierce, 1979). Therefore, they are most useful where found strewn in long lines, or trains, and where they have been plucked from a known and limited source area. In these positions, "boulder travel" takes on a linear dimension and shows the direction of ice flow.

Figure 6. Ice transported boulders scattered across the valley near the Chain Lakes.

RECONSTRUCTION OF THE ICE SURFACE

Ice surface elevations are helpful in studying the erosional ability of an ice mass, as glacial erosion is a function of ice thickness and its basal characteristics. Most areas inundated by ice during the last glaciation "are defined by end moraines, lateral moraines, erratics on glacially scoured terrain, large-scale glacial scour features (apparent on aerial photographs and on the ground), and fresh glacial striations and polish" (Pierce, 1979). To determine the ice surface altitude, however, the upper limits of these features must be located.

If an area has been glaciated, the landscape that results is enough proof that ice once covered the area. Questionable terrain would include, high peaks, buttes, and ridges that rise significantly above the surrounding topography, and jagged, rough, extremely weathered knobs and hilltops. Such areas could have been high enough to escape inundation by ice. If so, the elevation of the ice surface at this point must have been somewhere below the summit elevation of the feature. Examination of a lower, glaciated area, upslope to the higher, unglaciated one may provide an elevation at which the effects of glacial flow and erosion no longer appear. These unglaciated high points, or nunataks, provide the best evidence for ice surface elevations. As a portion of land that remained above the ice surface during the last glaciation, a nunatak should show no signs of recent glaciation above the limit of ice.

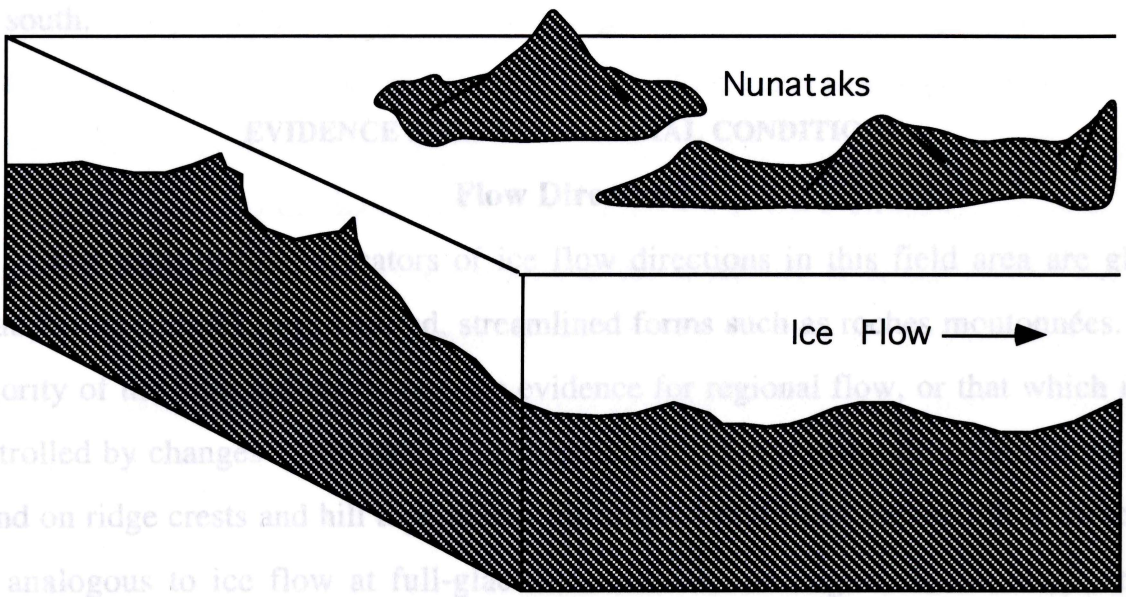

Figure 7. Nunataks Piercing the Surface of an Ice Sheet

On the other hand, some elevated areas *do* exhibit evidence of glacial scouring all the way up to their summits. This way of approximating ice surface elevation notes the elevations of high areas that the ice did cover. In such a case, the minimum ice surface elevation would be at least as high as the highest glaciated summit.

FIELD OBSERVATIONS

The primary objective of this study is to determine former flow directions, surface elevation, and thickness of the Beartooth Plateau ice as it radiated from the plateau toward the Clarks Fork of the Yellowstone River. With this information, the former glacial system of the area can be reconstructed. The study area is located in the Muddy Creek, Beartooth Butte, and Deep Lake 7.5 minute quadrangles of Wyoming.

Combined, these include the southernmost portion of the Beartooth Plateau and much of the glaciated highlands which slope south toward the Clarks Fork (see Figure 3). Specific attention was placed on the area between Beartooth Butte in the west to Tibbs Butte and Sawtooth Mountain in the east and southeast, respectively, and from the Montana-Wyoming border in the north, to the northern edge of Clarks Fork valley in the south.

EVIDENCE FOR FULL-GLACIAL CONDITIONS

Flow Directions

The most useful indicators of ice flow directions in this field area are glacial striations on bedrock, and scoured, streamlined forms such as roches moutonnées. The majority of the flow indicators provide evidence for regional flow, or that which is not controlled by changes in the underlying topography. Since many of the striations were found on ridge crests and hill tops, it may be inferred that these regional flow directions are analogous to ice flow at full-glacial conditions. In Figure 3, the black arrows represent the locations and orientations of glacial striations. The red arrows indicate the same flow type, but their orientations have been determined by streamlined features apparent on both aerial photographs and topographic maps.

The striations representative of full-glacial flow radiate from the Beartooth Plateau, ranging from S45˚W in the west, due south in the central portion, and S44˚E in the eastern section of the map (Plate 1 or Figure 3). In the central to eastern portion of the study area, the orientations of glacial striations match those of the roches moutonnées and bedrock knolls that constitute much of the evident aerial scouring. Over much of the field area, including this area from Long Lake to Fantan Lake, striation orientations and long axis alignments of topographic features coincide.

The striations located on top of the valley ridge, southwest of Night and Island Lakes (about 5 kilometers east of Beartooth Butte), do not coincide with the topographic trend (see Plate 1 or Figure 3). Instead, they occur perpendicular to the valley orientation. Therefore, it appears that the ice flowing down from the plateau continued its southwestern path even as it was bisected by a northwest trending valley and ridge. This example of flow at maximum glacial conditions is made more credible by the striations found north of Night Lake on the valley floor. Their orientations parallel that of the valley walls, and are indicative of glacial flow governed by the shape of the ground. In the field area, other striations occur which parallel topographic trends, but none so striking as those near Night Lake. The reason being that many of the other valleys and ridges already trend in the same direction as the full-glacial ice flow. The only other location where flow directions differ markedly is in the broad east-west trending valley in the central portion of the study area. About 2 kilometers east of Beartooth Lake, these striations are oriented east-west, indicating flow in a different direction, and thus at a later time than that of the regional flow to the southwest (Pierce, 1979).

The unmistakable flow of ice down the Clarks Fork Valley is also worth mentioning, although it is not directly investigated here. The field area involved in this project stretches only as far as the cliffs that make up the northern wall of the Clarks Fork Valley, but it is doubtless that the ice flowing in this valley affected the dynamics and behavior of the ice flowing from the north. The Clarks Fork outlet glacier forked from the Soda Butte glacier at the culmination of Pilot and Index Peaks in the west (Pierce, 1979). From there, the glacier headed approximately southeast and then east while adjacent to the southern portion of the study area. Little of the Clarks Fork may be seen on Plate 1, and only a slightly larger section on the map in Figure 3, but the glacier's previous ice flow direction is indicated. The majority of the ice flowing down from the plateau , then, must have curved from south-southwest flow to east-southeast upon meeting the Clarks Fork outlet glacier (see Figure 8).

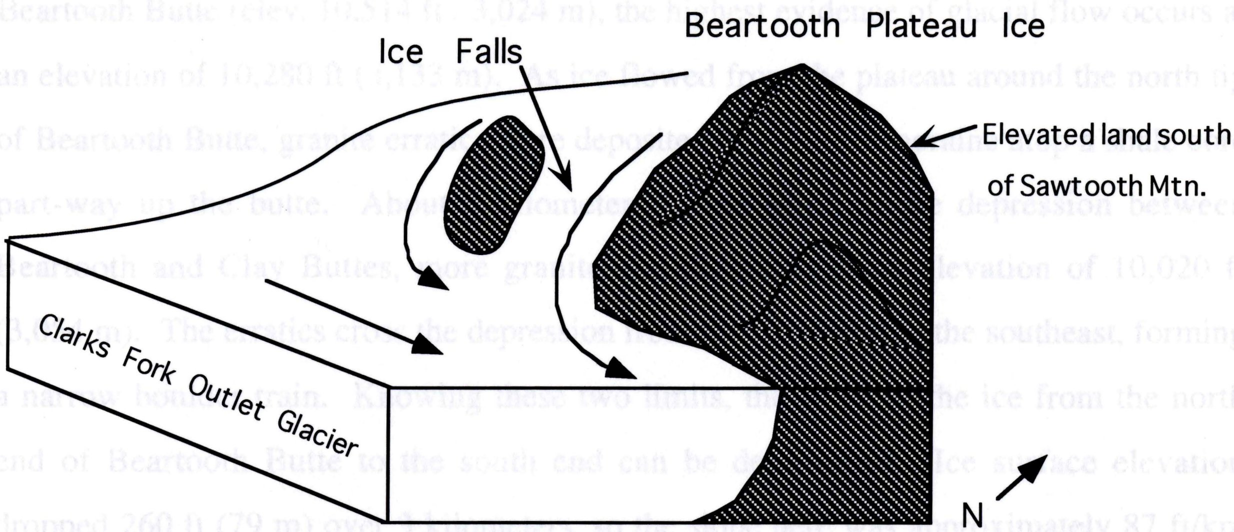

Figure 8. Ice Flow from the Beartooth Plateau at its confluence with the Clarks Fork Outlet Glacier.

Ice Surface Altitude

The determination of ice surface altitude during full glacial time, and the subsequent construction of ice surface contours relies upon what little data may be obtained from field observation combined with knowledge of ice flow directions, the direction of ice surface slope, and a great deal of lateral projection of known ice surface elevations. Unfortunately, much of the past evidence of glaciation on possible nunataks has been removed by mass wasting of glacially oversteepened slopes. This is particularly apparent on Beartooth Butte, where sedimentary rock is less resistant to weathering and erosion, and is still happening today. At first glance, many of these craggy-topped mountains appear to have escaped the last glacial inundation, but actual evidence indicative of the extent of ice is less conspicuous. Therefore, estimates must be made based on lateral projections from known ice surface elevations in conjunction with other indicators of flow characteristics.

Limits at Beartooth Butte

Although the landscape surrounding Beartooth and Clay Buttes bears signs of glacial erosion and aerial scour, the buttes themselves appear to have remained free of ice flowing from the Beartooth Plateau (see Figure 9). Off the northernmost point of

Beartooth Butte (elev. 10,514 ft / 3,024 m), the highest evidence of glacial flow occurs at an elevation of 10,280 ft (3,133 m). As ice flowed from the plateau around the north tip of Beartooth Butte, granite erratics were deposited in a medial moraine atop a shale core part-way up the butte. About 3 kilometers to the south in the depression between Beartooth and Clay Buttes, more granite erratics occur at an elevation of 10,020 ft (3,054 m). The erratics cross the depression from the northwest to the southeast, forming a narrow boulder train. Knowing these two limits, the slope of the ice from the north end of Beartooth Butte to the south end can be determined. Ice surface elevation dropped 260 ft (79 m) over 3 kilometers, so the slope here was approximately 87 ft/km (26m/km).

Figure 9. Beartooth Butte as a nunatak.

Figure 10. A granite erratic (foreground) atop Table Mountain with Beartooth Butte to the northwest.

Seen clearly on the map in Plate 1 and Figure 3, Beartooth Butte also contains its own interior valley. Although the ice from the plateau does not appear to have breached the outer walls , it is probable that the valley contained a glacier or ice field of its own). The valley is broad and U-shaped, containing a stream and marshy areas to the east and northeast of Clay Butte. Portions of the valley floor are lined with exposed limestone and the valley walls exhibit active slumps and landslides. This type of mass wasting commonly occurs after slopes are oversteepened by glaciation. Though, without concrete evidence of glaciation, we can only assume that at the very least, a snow or ice field accumulated in this valley.

Beartooth Butte was originally approached with the notion that the ice surface elevation was going to be much higher. A map of ice surface altitude contours was prepared by K.L. Pierce (1979) to show ice at full-glacial conditions in the Yellowstone area to the west. By projecting the contours of that map further to the east (to Beartooth Butte), it was initially presumed that the ice surface would reach approximately 10,500 ft (3,200 m) around the Butte. Since no traces of glacial erosion or deposition were found higher than 10,280 ft (3,133 m), this experiment proves that wherever possible, it is best to obtain actual data from the field. Unfortunately, attempting to reconstruct the ice surface over about 300 square kilometers of land undoubtedly requires a good bit of lateral projection and estimation. Hopefully, evidence such as flow directions at the glacial maximum and glaciated high points that indicate minimum ice surface altitude, provide enough supplementary information to reduce the error inherent in estimation.

Limits at Sawtooth Mountain

In the southeast portion of the field area, ice limits occur around Sawtooth Mountain (elev. 10,252 ft / 3,125 m) and the hill to the southwest (elev. 9,919 ft / 3,023 m). The ragged, rocky spires of Sawtooth Mountain provide evidence that ice did not overtop the mountain. Rarely do such pinnacled features survive inundation by ice. The altitude of the ice surface was estimated to be 9,900 ft (2,957 m) on Sawtooth Mountain and about 9,700 ft (2,957 m) on the hill southwest of Sawtooth Mountain. Between these two mountains, the ice surface altitude decreases by 200 ft (60 m) over a distance of 2 kilometers. Therefore, the former ice surface slope about Sawtooth Mountain was approximately 100 ft/km (30 m/km). On Plate 1 and Figure 3, the 9,800 foot contour lies approximately 100 feet above the elevation of the land in the valley between Sawtooth Mountain and the hill to the southwest. Here, a peat bog lies in the depression as well as a train of ice transported boulders trailing south from the east side of Sawtooth Mountain, indicating that some ice did find its way up into the valley, possibly pushed around the mountain from the northeast. It also appears as though no ice covered the elevated land in the southeastern portion of the map area.

Other Possible Limits and Ice-Covered High Points

Other nunataks occur in the central portion of the field area, approximately 1 kilometer north of Bird Mountain. These granite knobs appear ice free because they are deeply weathered and crumbling, lacking the fresh, pink, scoured and polished appearance of bedrock on nearby, less-elevated terrain (see Figure 11). Ice surface elevation on these hills was barely below the 9,960 ft (3,035 m) and 10,000 ft (3,048 m) summits of these hills. Mere meters of elevation separate the freshly glaciated terrain from the outcrops of apparently old and unmodified granite, suggesting that the ice altitude here was very close to the top of the knolls. Interestingly, these knolls, weathered as they are, still exhibit a streamlined shape, and an asymmetrical profile. The south-facing slope of these and other hills in the area are steeper than the north facing ones because the ice has plucked chunks of granite from the down-ice side, resulting in a steeper, blocky surface of joint and fracture planes. The highly weathered knolls still exhibit this general shape suggesting that they may have been glaciated at an earlier time (possibly Bull Lake?).

Figure 11. Highly weathered, dark granite knob (foreground) shown in contrast with fresh, pink glaciated granite (background).

Minimum ice surface altitudes were determined at locations where the highest features show evidence of recent glaciation. About 3 kilometers directly east of

Beartooth Butte, a cliff stretches upward from the eastern shore of Beauty Lake to an elevation of 10,064 ft (3,068 m). At the top of this cliff is a pink, granite outcrop that was rounded and smoothly polished. More polish and glacial striations occur further southeast. On the summits of Bird Mountain (9,924 ft / 3,026 m) and Table Mountain (8,761 ft / 2,671 m), erratics and ice-transported boulders provide evidence of ice coverage. Granitic ice-transported boulders occur atop Bird Mountain along with erratic pieces of reddish sandstone. One of the smooth, rounded boulders on Bird Mountain is about 10 meters long, 4 meters wide, and 4 meters high! Granitic erratics also occur on the limestone summit of Table Mountain, located in the southwestern portion of the field area. Here, large chunks of limestone from the point of Table Mountain appear to have been plucked from the point and moved several meters to the southeast by an ice mass passing over the summit.

DATA INTERPRETATION

CONSTRUCTION OF ICE SURFACE CONTOURS

"The orientation of contours on the ice surface and the direction of surface slope can be determined if appropriate glacial scour or other features are observed" (Pierce, 1979). The ice surface elevation contours on Plate 1 and Figure 3 were constructed perpendicular to the flow of ice at full-glacial conditions, using striations found on summits and ridge crests. Using the known ice surface elevations determined at the nunataks (shown encircled by solid red lines on Plate 1 and Figure 3), the flow directions aided in determining the correct lateral projection of the contours into areas of unknown ice surface altitude. The contours lay perpendicular to the ice flow direction and generally concur with the regional topographic trends. The final map of contours and probable nunataks shows a possible reconstruction of the icemass flowing down from the plateau.

BASAL SHEAR STRESS OF RECONSTRUCTED ICEMASS

"Calculation of basal shear stress is a powerful tool in evaluating the plausibility of reconstructions of former ice masses" (Pierce, 1979). This is because the two primary variables that determine basal shear stress are ice thickness and surface slope. A cross section of the reconstructed icemass (Figure 12) show these parameters, making it possible to calculate the basal shear stress from the formula $\tau_b = \rho g h \sin\alpha \bullet F$ (Nye, 1952). In the equation, τ_b = basal shear stress in bars, ρ = ice density (0.9 g/cm^3), g = acceleration of gravity (980 cm/sec^2), h = the ice thickness (in centimeters), α = slope of the ice surface, and F is the shape factor, assumed here to be 0.7 to account for drag (Pierce, 1979). The majority of modern glaciers have basal shear stress values ranging from 0.5 and 1.5 bars, with the reasons for the this narrow range quoted here by W.H. Matthews (1967, in Pierce, 1979):

> From laboratory tests it has been inferred that if the ice thickness or surface slope of an ice sheet somehow became so great that basal shear stress greatly exceeded this fixed value, the basal ice could be deformed so easily that the sheet would very rapidly spread out until its thickness or slope was significantly reduced. Conversely, if because of combination of thinness and flatness in an ice sheet the basal shear stress was very low, movement would virtually cease until accumulation of more ice increased thickness or slopes or both, and these in turn increased the basal shear stress.

Basal shear stresses calculated from the reconstructed icemass stretching from the Beartooth Plateau to the Clarks Fork Valley average 0.7 bars, as determined by several longitudinal cross sections radiating from the ice sheet crest on the plateau. One cross section trending northeast – southwest along flow lines is shown in Figure 12. The transect is divided into four sections based on average ice thickness, surface slope, and land surface gradient which allow calculation of basal shear stress values representative of different uniform flow types. The values range from 0.75 bars between the ice crest at approximately 11,100 ft (3,383 m) and the head of Little Bear Creek where the ice surface is approximately 10,800 ft (3,292 m), 0.62 bars from this creek to the east–west trending valley south of Little Bear Lake, 0.34 bars from this valley to the cliffs at the top of the northern Clarks Fork valley wall, and 1.6 bars south along the cliff face into the valley where plateau ice presumably met with the east–flowing Clarks Fork ice.

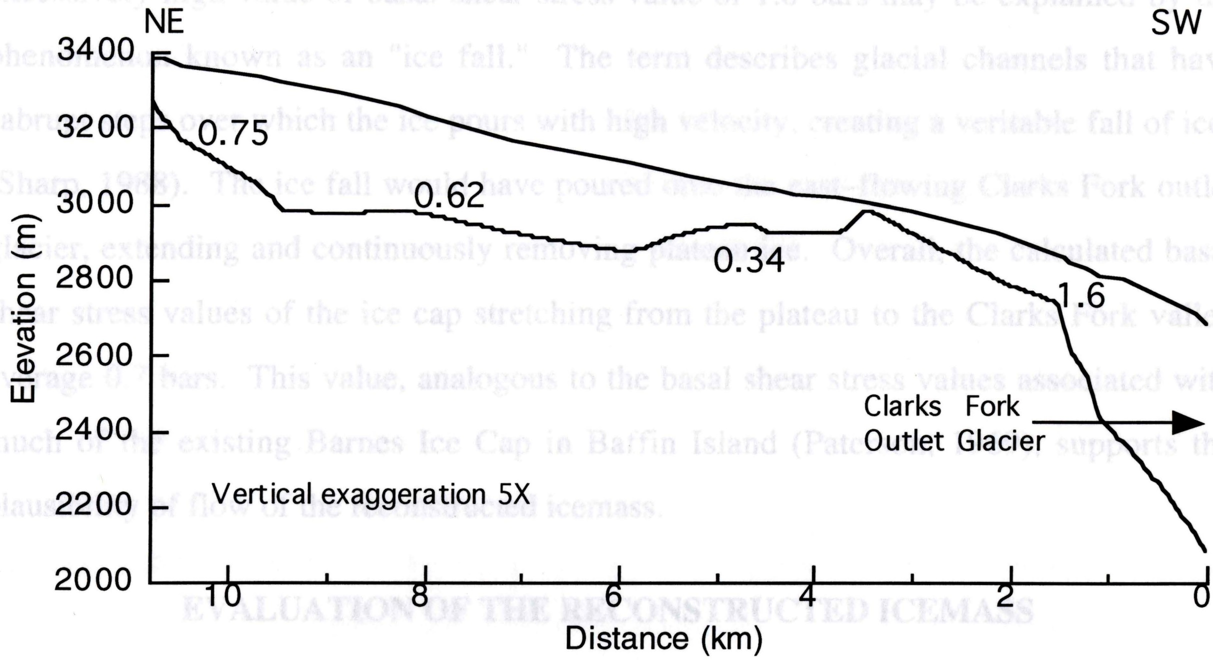

Figure 12. Longitudinal cross section of ice from the Plateau to the Clarks Fork Valley showing basal shear stress values

Table 1. Values for basal shear stress calculations from cross section in Figure 12.

Section	Thickness, h (m)	Drop(m)/ Distance (km)	slope (α) in degrees	Basal Shear Stress (bars)
1	304.8	91.44 /2.29	2.29	0.75
2	249.9	182.88/4.57	2.29	0.62
3	172.2	121.92/3.81	1.83	0.34
4	396.2	274.32/4.19	3.75	1.6

The different basal shear stress values calculated for the 14.9 kilometer cross section in Figure 12 represent changing land surface gradients beneath the ice, subsequent changes in ice surface elevation, and ice thickness values. Where the basal shear stress reduces to 0.34 bars, a value falling below the range stipulated for modern glaciers, the ice is actually flowing uphill. Striations indicate that ice flow during the glacial maximum continued southwest across the valley containing the Chain Lakes, erosion. This develops where "ice erosion has been concentrated in a trough or series of troughs and has left the intervening slopes or plateau unmodified" (Sugden, 1976). This type of landscape develops beneath ice sheets, leaving the intervening plateau areas "regolith covered and devoid of glacial erosional forms," with "fragile rock remnants like tors of possible pre-glacial age surviving on the interfluves" (Sugden, 1976). This leads to the idea of erosion only occurring where there is basal sliding. In turn, basal sliding only occurs where the ice is thick enough for the pressure melting point to be reached. "Landscapes of selective linear erosion form an intermediate category where basal ice is at the pressure melting point only in troughs," remaining "below the pressure melting point over the sites of intervening interfluves" (Sugden, 1976). The unusual thing about the erosional pattern of the Beartooth Plateau ice cap is that it did not form large troughs, rather, it scoured vast expanses of varying topography to the south of the southern edge of the plateau while leaving the plateau surface unmodified. It is not unlikely, however, that a combination of selective linear erosion and aerial scouring may have produced a landscape resulting from "selective aerial scouring." Basal melting leads to ice flow and glacial erosion, and "since ice temperatures rise with depth, conditions are more favorable for basal melting where ice is thicker" (Sugden, 1976). This may explain the juxtaposition of two vastly different glacial landscapes in the area of the Beartooth Plateau if, in fact, the plateau ice was thin and cold-based, with warm-based ice flowing from the plateau edge to the Clarks Fork.

Another alternative is that ice only reached as far as the cirque headwalls that circumvent the highest part of the plateau. This would leave the upper plateau devoid of ice. The approximate elevation of the top of the cirque headwalls ranges from 11,000 feet (3,354 m) in the north-central portion to 10,600 feet (3,232 m) in the northeastern portion of the field area. However, reducing the ice surface elevation to the level of the highest cirque headwalls would cause the rest of the ice surface contours to be altered as well, reducing the thickness of the ice to the south. Considering the values of basal shear stress with the model as it is now, reducing the ice thickness, h, would yield unreasonably lower values of basal shear stress, indicating that the ice would not be able to flow. However, the extensive glacial erosion apparent from the plateau to the Clarks Fork valley is proof enough that ice did indeed flow there. Also, small glaciers are presently located in the north facing cirques south of Emerald Lake, where the ice, even now, stretches as high as 11,200 feet (3,415 m). It is difficult to comprehend how ice may have only reached as high as the top of the cirque headwalls during the Late Wisconsin glacial period, when even today, glaciers exist at this elevation. The other alternative, therefore, presents the possibility that there may have been ice, thin ice, or at least a big pile of snow on the plateau, supported by the capacity of periglacial landforms to remain intact beneath thin, cold-based ice sheets.

Locating nunataks makes it possible to calculate some average ice surface slopes both in the western and eastern portions of the map area. The altitudes of the contours on the ice surface are "based on nunataks and maximum altitudes of glaciated uplands" (Porter and others, 1983) They show that the ice was considerably thinner on the plateau (thin or absent depending on the theory), and thicker to the south (up to 3,000 feet or 915 meters in the Clarks Fork Valley!). This evidence suggests that the plateau ice was either absent or cold-based and frozen to its bed allowing no flow, and therefore no glacial erosion on the plateau.

DISCUSSION OF FLOW PATTERNS

The two main sets of flow directions shown in Figure 3 indicate signs of both valley glacier forms and ice sheet forms, which appear to have evolved at different times. The flow directions which radiate from the Beartooth plateau from S45°W to S45°E represent a phase of ice cap flow. These directions, found at some of the highest elevations, indicate flow in the upper part of the ice during maximum glaciation. The other set of flow directions indicates a later period when topography constrained the glaciers. These glaciers were confined to valleys, as indicated by the striations that they etched in directions parallel to valley orientations. The presence of cirques along the edge of the upper Beartooth Plateau also supports the evolution of a glacial phase other than ice cap growth. "Since there is no reason to suppose that cirques form beneath ice sheets, such situations have long been accepted as representing different phases of mountain and ice sheet glaciation" (Sugden, 1976). Pinedale glaciers, named to represent the last glaciation occurring in this area, were near their maximums about 20,000 years ago (Porter and others, 1983) and were characterized by ice cap forms. Over the next 10,000 years, the ice caps diminished in size and scope yielding complexes of valley glaciers, evident from the striations found oriented in the same directions as the valley walls.

SUMMARY

The model of the glacial system from the Beartooth Plateau to the Clarks Fork valley was constructed on the basis of observed ice flow direction, known and estimated ice surface altitude, and ice surface slope. Basal shear stress calculations of the resulting model yield an average value of 0.7 bars, demonstrating the reconstruction to be physically reasonable. Falling between 0.5 to 1.5 bars, this average basal shear stress lies within the "general limits observed on modern glaciers" (Pierce, 1979). In the northern Yellowstone area, the "basal shear stress for areas with strongly converging flow lines averages 1.2 bars, but the basal shear stress for those with strongly diverging flow lines averages 0.8 bars, thereby providing an explanation for most of the differences in calculated basal shear stress" (Porter and others, 1983). Similarly, the basal shear stress values calculated for much of the Beartooth Plateau ice cap fluctuate locally, yet maintain an average of 0.7 bars. This is indicative of the radiating, divergent flow lines of ice moving southward toward the Clarks Fork valley.

The presence of periglacial landforms and lack of glacial scour on the Beartooth Plateau gives the impression that ice did not reach the plateau surface during the last glacial maximum. Although features such as tors and patterned ground do not usually survive glacial inundation, isolated cases have been found in places like the Cairngorm Mountains in Scotland (Sugden, 1968). Selective glacial erosion has been attributed to ice being thin and placed on a low gradient, yielding unreasonably low basal shear stress values which allow no basal melting, and therefore no flow. Ice like this would be termed cold-based, or frozen to its bed. It is possible that features below thin, cold-based ice could survive. The other argument is that ice did not reach as far as the plateau surface. Reducing the ice surface elevation like this would reduce ice thickness, and consequently reduce the basal shear stress values for much of the ice flowing down from the Beartooth Plateau. With the values already near the low end of the accepted range for modern glaciers, lowering the ice surface elevation further would yield even lower values of basal shear stress, possibly eliminating all basal sliding.

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aided most likely by compression from upstream ice as well as extension from downstream ice flowing over the lip of the northern Clarks Fork valley wall. Here, the excessively high value of basal shear stress value of 1.6 bars may be explained by the phenomenon known as an "ice fall." The term describes glacial channels that have "abrupt steps over which the ice pours with high velocity, creating a veritable fall of ice" (Sharp, 1988). The ice fall would have poured onto the east–flowing Clarks Fork outlet glacier, extending and continuously removing plateau ice. Overall, the calculated basal shear stress values of the ice cap stretching from the plateau to the Clarks Fork valley average 0.7 bars. This value, analogous to the basal shear stress values associated with much of the existing Barnes Ice Cap in Baffin Island (Paterson, 1969), supports the plausibility of flow of the reconstructed icemass.

EVALUATION OF THE RECONSTRUCTED ICEMASS

PERIGLACIAL LANDFORM SURVIVAL ON THE BEARTOOTH PLATEAU

While attempting to reconstruct the icemass flowing from the Beartooth Plateau to the Clarks Fork Valley, several problems were encountered. The first problem, mentioned in previous sections, concerned the estimation necessary to construct a model of ice surface altitudes. After combining the ice flow data and known elevations with the results of previous studies in nearby areas, it appears that the resulting model is plausible. However, this model calls for ice to be present on a large portion of the plateau surface. The Beartooth Plateau surface, however, shows no signs of glaciation. Rather, it is representative of a periglacial environment exhibiting patterned ground, tors, and markedly different terrain than that lying to the south of the plateau. Several explanations exist for this discrepancy.

First, glacial erosion may be selective under ice. The sharp contrast between the smooth, relatively flat surface of the highest portion of the Beartooth Plateau and the modified, carved landscape to the south is similar to a landscape of selective linear

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