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Overdeepened glaciation, past and present:

Lake Pend Oreille, northern Idaho and Taku Glacier, southeastern Alaska

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ABSTRACT

Seismic surveys show that fiord-forming glaciation has overdeepened valleys to depths hundreds of meters below sea level both at former interior Cordilleran ice margins as well as at present glaciated coastlines in the Pacific Northwest. At Lake Pend Oreille, near the southernmost extent of Cordilleran glaciation in Idaho, the bedrock surface has been cut by ice to depths over 300 m below present-day sea level. Deposited in the lake are over 600 m of glacial sediments with a stratigraphy remarkably similar to that of the overdeepened lake basins of southern British Columbia. Pleistocene ice filled Lake Pend Oreille, formed the Clark Fork ice dam, and impounded glacial Lake Missoula. The latest of the catastrophic Missoula Floods, caused by repeated failures of the ice dam, scoured Lake Pend Oreille bottom sediments, resulting in a U-shaped unconformity and the impressive 350 m depth of the lake. The Taku Glacier, the largest on the Juneau Icefield in southeastern Alaska, extends as deep as 800 m below sea level. Cross sections of the glacier show overdeepened channels carved into former subglacial surfaces. In the last 100 years, fiord-forming glacial processes have gouged 140 m of subglacial sediments at sub-sea level elevations from beneath this glacier. We propose that the Taku Glacier is a modern analog for the glacial overdeepening that occurred in Lake Pend Oreille and other lake basins at the edge of Cordilleran ice margins in North America.

INTRODUCTION

Glaciers, under gravity flow conditions, are an efficient means of erosion down to sea level. However, an increasing body of seismic evidence indicates that many glaciated valleys in North America and Europe were cut into bedrock to depths hundreds of meters *below* sea level (Mullins and Hinchey, 1989; Finckh et al. 1984). Although tectonic subsidence is commonly posed for such overdeepened bedrock depressions, we believe fiord-forming glaciation in the presence of proglacial lakes is a more likely explanation. By this mechanism, the bouyant force on the floating ice near the terminus reshapes the ice into a wedge that thickens upglacier with maximum basal erosion at the junction of floating and grounded ice, resulting in overdeepening below water level.

Lake Pend Oreille in northern Idaho and the Taku Glacier in southeastern Alaska (Figure 1) occupy long, linear, deep valleys, preferentially located along pre-existing structural features, at the margins of former or present ice sheets. Seismic studies of the bedrock morphology and sedimentary facies in similar fiord-like valleys at the edge of former ice sheets, such as the Okanagan and Kalamalka valleys in British Columbia, have revealed bedrock erosion to depths well below sea level and subsequent rapid infilling by sediment (Mullins et al., 1990; Eyles et al. 1990; Fulton and Smith, 1978). These inland valleys, which contain substantial sediment thicknesses, owe their morphology to large-scale glacial erosion at the edge of Cordilleran ice sheets and are thought to be remnants of much larger proglacial lakes.

In the first part of this paper, we present data from seismic reflection surveys of Lake Pend Oreille. These data, obtained from the U.S. Navy, show overdeepening of the bedrock surface to below sea level and a substantial section of glacial sediment. With this information, we provide new insight into the glacial history of the Purcell Trench and suggest a genetic link between Lake Pend Oreille and the fiord-like Pleistocene lakes of British Columbia. We also synthesize the seismic data with the glacial morphology of the southernmost extent of Cordilleran glaciation in Idaho.

In the second part of this paper, we present new seismic reflection data from the Taku Glacier. This glacier, the main outlet of the Juneau Icefield of the Alaska-Canada Boundary Range between Juneau, Alaska and Atlin, British Columbia, is the deepest temperate glacier yet discovered, with ice extending as deep as 800 m below sea level. This glacier represents a modern analog of fiord formation at the edge of an ice sheet, a process that may also explain the overdeepened bedrock basins beneath Lake Pend Oreille and the fiord-like lake valleys of British Columbia.

FIORD-FORMING GLACIATION

A mechanism for glacial fiord formation arose from a study of Skelton inlet in Antarctica, a fiord with ice as deep as 1933 m below sea level (Crary, 1966). In this model, which critically depends on floating the ice at the terminus, the ice erodes bed material in a completely different manner than purely grounded glaciers. To a first approximation, the bouyant force on floating ice near a terminus reshapes the ice into a wedge that increases in thickness up valley according to:

$$11.17\frac{dZ}{dx}=\frac{\sigma_{xY}}{Y} \quad [1]$$

where x is the distance directed downstream (km), Z is ice thickness (km), Y is the half-width of the valley (km), σ_{xY} is the shear stress against the valley wall in bars (Crary, 1966, eq. 4). Typically, σ_{xY} ranges from -0.5 to -1.5 bars (Nye, 1952). The floating ice thickens upglacier with maximum basal erosion at the junction between floating and grounded ice, resulting in overdeepening below sea or proglacial lake level (Figure 2). As the glacier recedes, the zone of deepest erosion moves landward creating fiord basins.

As the floating ice moves down the inlet it retains its wedge shape in the water; the movement at the bottom has an upward component (Figure 2). Crary (1966) compared the mechanism to an endless belt or chain saw with direct erosive action taking place near the junction of the floating and grounded ice. Once floating ice is formed in a valley, erosion is focused in a zone near the junction with the grounded ice, continuously gouging up bed material. Unlike purely grounded ice that deforms at its base and is most efficient in clearing out obstacles to gravity flow to sea level, fiord-forming ice has little basal deformation and will in time erode a valley that increases in depth inland.

LAKE PEND OREILLE

Lake Pend Oreille, the largest lake in Idaho, is located about 80 km south of the British Columbia border in the Purcell Trench close to Sandpoint, Idaho (Figure 3). The lake level in Lake Pend Oreille is 629 m above sea level, with the surrounding terrain as high as 1830 m. The maximum depth of the lake is an impressive 350 m, the deepest lake, by far, in the region.

A review of the geologic history of the Lake Pend Oreille area is given by Savage (1965), summaries of Cordilleran ice sheet incursions into the region are provided by Waitt and Thorson (1983) and Richmond (1986), and a review of regional tectonics in light of recent COCORP traverses is given by Potter et al. (1986). The Purcell Trench, a major topographic depression in British Columbia and northern Idaho, is incised into the margin of a metamorphic and granitic complex, formed during Mesozoic convergent tectonics, but later subjected to Eocene extensional forces. The trench is bounded on the west by the Selkirk Range and on the east by the Cabinet Mountains. Prior to the Miocene Epoch, the ancestral Rathdrum River flowed within the Purcell trench from the Canadian border south to the Coeur d'Alene area in a meandering pattern following the least resistant rock exposures and fault zones. This pattern is still apparent in the present day meandering shape of Lake Pend Oreille. During Miocene time, a lava dam formed at the south end of present-day Lake Pend Oreille resulting in Tertiary Lake Rathdrum, with water levels up as high as 800 meters above sea level. However, the basalt dam eventually eroded, restoring the southern outlet of the lake, which persisted into Pleistocene time when great lobes of glacial ice moved south from Canada, once again damming the south end of the lake, this time with glacial debris. The present outlet of the lake is at the northwest end of the lake into the Pend Oreille River which flows to Washington and then north into British Columbia.

In eastern Washington, early ice lobes dammed the Columbia and its tributary drainages forming Pleistocene Lake Columbia (Waitt and Thorson, 1983) with water up to 792 m above sea level, thereby flooding the Rathdrum valley and creating proglacial waters in front of the ice sheets advancing down the Purcell Trench, a condition we believe was critical for fiord-forming glacial processes that overdeepened the Lake Pend Oreille basin to depths hundreds of meters below sea level.

Cordilleran ice advancing into Lake Pend Oreille formed ice dams at the northern end of the lake, repeatedly blocking the Clark Fork drainage and forming glacial Lake Missoula. Repeated failures of the ice dam caused catastrophic flooding known as the Missoula Floods and formed the Channelled Scabland of Washington. The terminus of the latest glacial advance extended to the southern end of the lake and upon failure of the ice dam, much of the flood discharge was directed through the lake basin and into the Rathdrum Prairie. This scouring of bottom sediment by the great floods is the reason for the great depth of the lake compared to other lakes in the region.

Method

The seismic data, obtained using towed acoustic sources and a towed hydrophone receiver, were recorded graphically at different times using high frequency (about 1000 Hz) and low frequency (500 Hz) seismic reflection systems, respectively. The higher frequency data, recorded in 1966, provided good vertical resolution of the shallow sediments, whereas the lower frequency records, recorded in 1988, were capable of resolving apparent reflections from bedrock as deep as 1 km below the lake surface. On most of the profiles, the instrument settings were selected to enhance reflections from the Navy's target of interest: the shallowmost sediments, not the bedrock surface. We believe a water velocity of 1472 m/s was used to set the depth scale on the 1988 data. Based on data from Okanagan Valley lake bottom sediments (MacAulay and Hobson, 1972), the actual sediment velocity is probably about 1600 m/s, resulting in a possible overestimation of up to 50 m in the deepest sub-bottom depths shown.

Results and Interpretation

The 1966 higher frequency record section across the lake is shown in Figure 4. We recognize four seismic stratigraphic units. The lake sub-bottom includes a section of layered sediments consisting of an upper unit (I) of thin-bedded silts and clays and a lower unit (II) containing several packages of coarser sediment. The net maximum thickness of stratified units I and II is 50 m, each unit being about 25 m thick in the center of the lake. The boundary between units I and II is highly reflective, suggesting a change in sediment type at that interface. A reasonable interpretation is that the uppermost unit I represents fine-grained lacustrine sediment deposited in Holocene time, and the lower unit II represents latest Pleistocene post-Cordilleran alpine glaciation. Alpine moraines, which occupied upper valleys in the Selkirk and Cabinet Ranges, post-date the recession of the Cordilleran Purcell Trench lobe. Although these glaciers did not reach the Lake Pend Oreille basin, they were a source of post-flood coarse sediments. If this interpretation is correct, the depositional rate of fine grained lacustral sediments in Lake Pend Oreille since alpine deglaciation has been about 2.5 mm per year. The U-shaped unconformity separating units II and III represents scouring by the latest flood episode. Unit III consists of poorly sorted glacial sediments, and the contact on the eastern margin with the bedrock unit (IV) is the side of the glacially-cut valley.

In an earlier interpretation of this section, Breckenridge (1989a, 1989b) considered unit III to be bedrock, with a U-shape surface cut by the glacier and that the persistent linear trend signature on the eastern margin of the basin was a major fault separating unit III from unit IV, another bedrock unit. In light of the newer more extensive seismic reflection profile available (Figure 5), we now recognize unit III as poorly sorted glacial sediment, and the contact on the eastern margin as a glacial valley wall matching a similar wall on the west.

The bathymetry of the profile shows that lake (elev 624 m) has steep bedrock walls that continue to depths near 200 m where the lake floor sediment is encountered lapping up against the bedrock surface. The lake floor is in the U-shape with a near flat center at a depth of 350 m. On the eastern edge, a bench occurs near a depth of 240 m, probably supported by an underlying glacial bedrock berm. A similar bench occurs above the water surface on the eastern shore of the lake at an elevation of 700 m, about 76 m above the present lake level. The western edge of the lake floor has a more gradual slope, without any evidence of bedrock benches. Submarine landslide deposits create a hummock appearance on the eastern slope of the lake floor.

Unit III, which represents over 99% of the total sediment volume in the lake basin, is homogeneous, without internal structure or bedding. On the higher frequency sections, the seismic wavelength is about 1.6 m in the sediments, and the vertical resolution is less than a meter. Thus, any bedding within unit III of anomalous composition and thickness of a meter or greater would be apparent on the seismic section.

The 500 Hz seismic reflection profile across the lake (Figure 5), though not recorded with settings to enhance the deeper reflections, nonetheless shows bedrock reflections on both sides of the glacial valley as deep as 650 m, well below sea level. Three line-drawings taken directly from other 500 Hz seismic reflection profiles A-A', B-B', and C-C' across the lake are shown in Figure 6. Only profile B-B' had control points where settings were appropriate to see deep bedrock reflections at the base of the glacial valley, but the similarity of the profiles suggests that sediments extend down to depths over a kilometer, more than 400 meters below sea level on all the profiles.

Discussion

Considering that Lake Pend Oreille is located immediately below the Lake Missoula ice dams, one obvious question is whether the deep bedrock depression beneath the lake could have been cut by the Missoula floods. In 1923, J Harlen Bretz of the University of Chicago began a series of research papers explaining the origin of the Channelled Scabland in Washington. He attributed the system of dry channels, coulees, and falls to an episode of flooding on a scale larger than geologists had ever recognized. His "outrageous hypothesis" was disputed by prominent geologists. The resulting controversy is one of the most famous in geology. Bretz's ideas for such large scale flooding challenged the classical uniformitarian doctrine then ruling the science of geology. Considering that the Missoula floods cut Grand Coulee, it is certainly not outrageous to consider if the floods might have cut the Pend Oreille basin as well.

The cause of the immense Missoula Floods were attributed to the sudden drainage of glacial Lake Missoula (Pardee, 1942). A great lobe of the Cordilleran ice sheet in Canada had advanced south down the Purcell Trench and dammed the Clark Fork Valley. The resulting glacial lake covered about 8000 km^2 of western Montana and held over 2200 km^3 of water. The failure of the ice dam, the cause of the flood, released as much as 17 km^3/s of floodwater from the Clark Fork Valley into the immediately adjacent Lake Pend Oreille basin. The terminus of the latest glacial advances extended to the southern end of the lake and upon failure of the ice dam, much of the flood discharge was directed through the lake basin and into the

Rathdrum Prairie (Figure 3). Although the floods were particularly effective in moving glacial material, we find it difficult to accept that even the great floods could have eroded 700 m of bedrock to depths so far below sea level in only one location. Bedrock is exposed in the flood channel near Athol, Idaho, within a few kilometers south of the lake, and seismic profile D-D' (Figure 6) directly in the flood path in the Rathdrum Prairie indicates a bedrock elevation of 243 m above sea level, 600 m above the bedrock elevation in the Pend Oreille Lake basin. Furthermore, the extreme similarity of the seismic profiles across Lake Pend Oreille to other large glacial lakes where great floods did not occur (British Columbia, New York, the Alps) leads us to seek another explanation for the overdeepening. However, as previously stated, we do believe that the extraordinary present-day lake depth and the U-shaped unconformity between units (II) and (III) are the result of flood scouring.

Another reasonable question to ask is whether the basin is tectonic in origin, perhaps an Eocene graben or half graben. There is no question that the Purcell Trench is a major structural feature. However, the route of the Purcell Lobe within the Purcell trench was controlled by the pre-existing ancestral Rathdrum River drainage. The incised stream valleys provided routes for rapid ice flow from southern British Columbia into northern Idaho. The meandering shape of the lake seems to rule out its origin as a direct structural feature. Furthermore, although many ancient thrust and more recent normal faults exist in the vicinity of the lake, none have been mapped with a geometry that would allow the shorelines of the lake to represent fault surfaces.

We believe the Lake Pend Oreille basin to be glacial in origin. As previously mentioned, the sub-bottom profiles of Lake Pend Oreille are remarkably similar to seismic profiles of other large glacial fiord-like lakes in British Columbia that were also at the edge of the Cordilleran ice sheet. Eyles et al. (1990) noted the extreme thickness of the sub-bottom sediment in Okanagan Lake and suggested that the large volume of glacio-lacustrine sediment may be typical of such lakes. Similar results were found in Kalamalka Lake (Mullins et al. 1990) and in lakes of the southern Alps by Finckh et al. (1984). The Pleistocene geological environment at Lake Pend Oreille would seem to have been ideal for fiord-forming glacial processes. The lake was at the edge of the Cordilleran ice sheet and a major proglacial lake was present to float the glacier at its terminus.

THE TAKU GLACIER

The Taku, the largest glacier system on the Juneau Icefield (Figure 1), embraces over 550 km^2 of area. The Taku Glacier is one of the most extensively studied temperate glaciers in the world, being the primary field area of the Juneau Icefield Research Program (JIRP) for the last 50 years. Miller (1996) provides a comprehensive bibliography of JIRP research. The mass balance of the Taku Glacier system is discussed by Pelto and Miller (1990).

Seismic experiments have been carried out regularly on the Juneau Icefield since the renowned exploration seismologist Charles Poulter sledged early seismic equipment up the Taku Glacier in 1949. Since 1992, the availability of portable high dynamic range, light-weight instrumentation on the icefield has resulted in high quality seismic records and revolutionary insight into ice thicknesses on the Taku Glacier. Like the Lake Pend Oreille basin and the fiord-lake basins of central British Columbia, the bedrock basin beneath the Taku Glacier apparently extends to depths hundreds of meters below sea level, establishing, according to Nolan et al. (1995), the Taku Glacier as the deepest and thickest temperate glacier yet measured.

Method

We have carried out six seismic reflection profiles along six profile lines across the main Taku glacier and its major tributaries (Figure 7). Surveying of shot and receiver locations and elevations were determined by Global Positioning System (GPS) measurements accurate to a few centimeters. We used a Bison 9000 series seismograph with geophones spaced 10 m apart along survey lines across the glacier. Explosive charges (one to ten sticks of dynamite) were loaded into shallow auger holes for each shot. Shot-receiver distances ranged from 30 m to as far as 1.5 km. Reflections from the base of the glacier were generally clear and easy to recognize on the records by their frequency, character, and distinctive moveout times (Figure 8). Cross-sections of the glacier were drawn by migrating the bottom reflections to their correct subsurface positions using a constant ice velocity of 3660 m/s, a value determined from P-wave first arrival times. The migration process was based on the mathematical assumption that the glacial cross sections were two-dimensional, perpendicular to the seismic lines.

Results

A digital image of a continuous seismic record section across the center of the Taku Glacier near Camp 10 is shown in Figure 9. This section has been corrected for normal moveout, and the time scale has been converted to depth, but the reflections have not been migrated. The reflection package from the base of the glacier has been digitally enhanced by a coherency filter. The inverted reflections, apparently from below the glacier, represent geometrical focusing as a result of the sharp curvature of the base of the glacier. This "bow-tie" effect, a common occurrence on seismic transects above convex-downward structures was observed on all the seismic profiles on the Taku Glacier. The mathematical migration process that we used to create the cross-sections of Figure 10 shifts the flanks of the glacier to their true positions and collapses the inverted diffraction arc to a point at the base of the glacier, which for the case of this profile is at a depth of 1450 m, 400 m below sea level (Figure 10, Profile IV).

Cross sections based on the seismic results of the six seismic profiles are shown in Figure 10. Because the glacier does not calve at present, the base of the ice is known to be near sea level at its terminus. At Profile I, about 5 km up-glacier from the terminus, the base of the ice already extends as deep as 250 m below sea level. Prior to this century's glacial advance, this transect was free of ice, and the maximum bathymetric depth in the inlet at this location was 108 m (Poulter, 1949). Since then, about 142 m of sub-glacial material has been eroded by glaciation in the deepest part of the valley. The bench on the left side Profile I clearly represents the pre-advance surface. We believe this behavior is typical, that of renewed fiord-forming glaciation gouging into former ice-cut surfaces. Profile II, located 21 km up from the terminus, shows the thickest and deepest ice found to date on the icefield. The ice is 1500 m thick and extends to 700 m below sea level. Like profile I, profile II shows bench development just above sea level, suggesting that renewed fiord-forming glaciation at some time in the past has drastically deepened a pre-existing glacial valley. Profile III, about 29 km up the northeast branch of the glacier shows about 1300 m of ice as deep as 300 m below sea level with a bench at about 400 m elevation on the southeast side. Profile IV, about 32 km up from the terminus near Camp 10, shows a cross section with well-developed benches at about 420 m above sea level. The maximum ice thickness is 1450 m, extending as deep as 320 m below sea level. Profile V on the currently slow-moving southwest branch of the glacier has about 1100 m of ice extending to only 50 m below sea level; this is the only profile we obtained on the glacial system to date that shows a simple U-shape glacial cross-section without bench development. Profile VI, about 42 km from the terminus, shows 1450 m of ice extending as deep as 175 m below sea level with a bench developed at about 100 m above sea level.

Discussion

Profiles across the main Taku Glacier show ice to depths hundreds of meters below sea level. Figure 11 shows the subglacial topography of the Taku Valley. When compared with Figure 7, the oversteepened walls and overdeepened valley form a spectacular picture. Comparisons of present ice thicknesses near the terminus with actual bathymetric soundings from a hundred years ago when the ice had retreated, shows that glacial erosion with an upward component has been occurring even with the most recent glacial advance. At Profile I, we estimate a bed erosion rate of about 1.4 m/yr since 1890. This process has gouged a new glacial channel into the old glacial bed leaving a characteristic bench.

CONCLUSION

Seismic reflection profiles of Lake Pend Oreille in northern Idaho and on the Taku Glacier in southeastern Alaska suggest that the bedrock surface has been cut in both localities by glaciation to hundreds of meters below present-day sea level. These results show that overdeepening has occurred both at glaciated coastlines and at interior Cordilleran ice margins. The 1.4 m/yr erosion of sub-glacial material at profile I of the Taku Glacier since 1890 is a dramatic modern analog for fiord-forming glaciation with valley depth increasing upstream.

Lake Pend Oreille, like the large lakes of southern British Columbia, appears to be a fiord lake. It is steep-walled, elongate, and contains a trough overdeepened to below sea level. For fiord lakes to form inland, the presence of a large proglacial lake is necessary and the ice must be sufficiently thin relative to the lake depth to float and the ice must be moving very quickly. Assuming the same lake level as today, to erode the deepest portion of the Pend Oreille basin, the ice would have had to have been only about 1.1 km thick, extending upward to an elevation only 100 m above the present lake level.

Most of the Taku Glacier profiles show bench development. The benches are at elevations above sea level and are remnants of former subglacial surfaces. At times of thinner ice and rapid ice movement into the inlet, the fiord-forming process gouged deeper channels inset into the older subglacial surfaces. It is possible that the bedrock benches present on the margins of Lake Pend Oreille represent the same process.

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References

Breckenridge, R.M. 1989a. Pleistocene ice dams and glacial Lake Missoula floods in northern Idaho and adjacent areas, *in* V.E. Chamberlain , R.M. Breckenridge, and Bill Bonnichsen, eds., Guidebook to the Geology of Northern and Western Idaho and Surrounding area. Idaho Geological Survey Bulletin **28**: 5-16.

Breckenridge, R.M., 1989b. Lower glacial lakes Missoula and Clark Fork Ice dams, *in* R.M. Breckenridge, editor, Glacial Lake Missoula and the Channeled Scabland, American Geophysical Union, 28th International Geological Congress, Field Trip Guidebook T310, pp. 13-23.

Crary, A.P. 1966. Mechanisms for fiord formation indicated by studies of an ice covered inlet. Geological Society of America Bulletin, **77**: 911-929.

Eyles, N., Mullins, H.T., and Hine, H.C. 1990. Thick and fast: Sedimentation in a Pleistocene fiord lake of British Columbia, Canada. Geology, **18**: 1153-1157.

Finckh, P., Kelts, K., and Lambert, A. 1984. Seismic stratigraphy and bedrock forms in perialpine lakes. Geological Society of America Bulletin, **95**: 1118-1128.

Fulton, R.J., and Smith, G.W. 1978. Late Pleistocene stratigraphy of south-central British Columbia. Canadian Journal of Earth Sciences, **15**: 971-980.

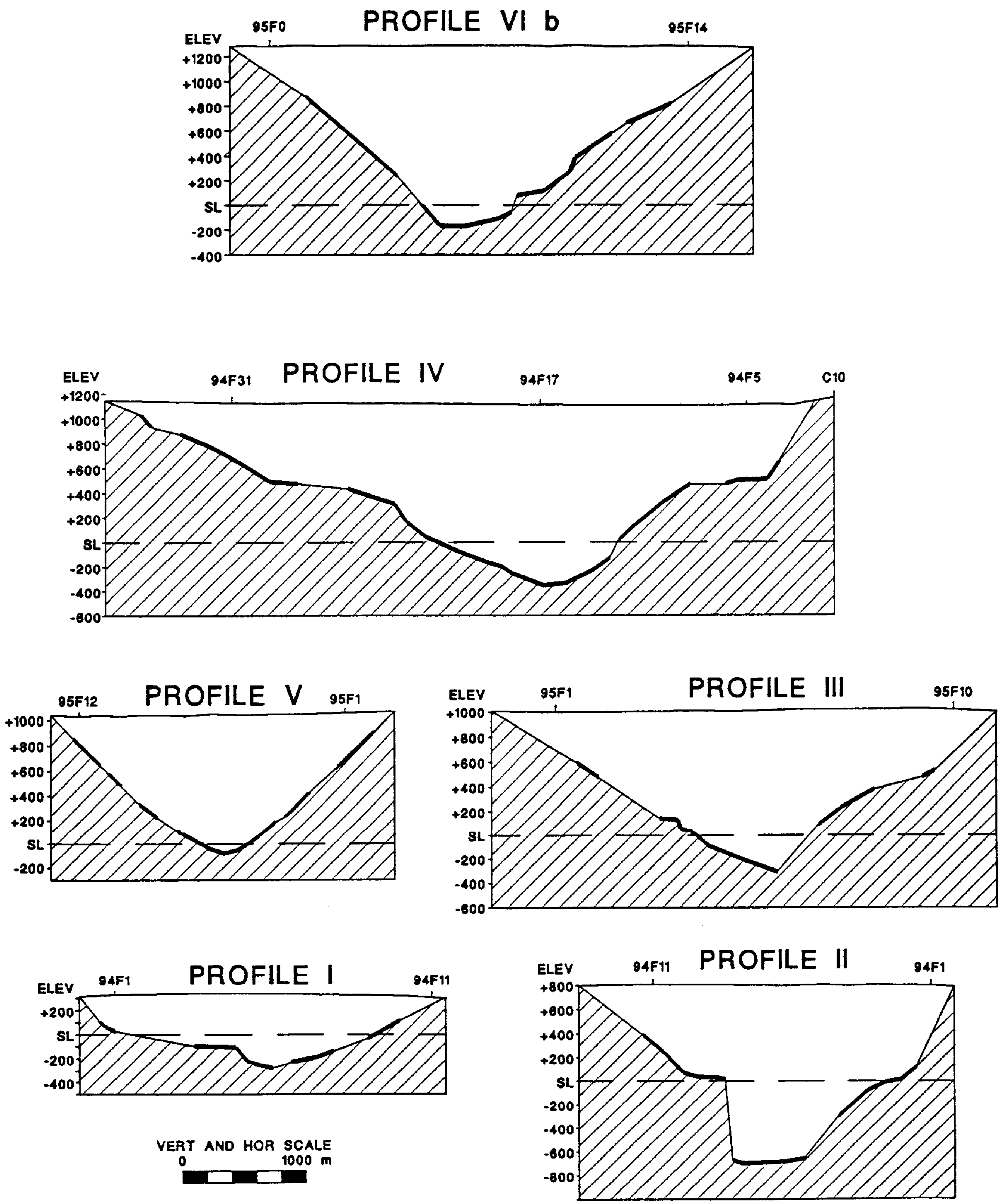
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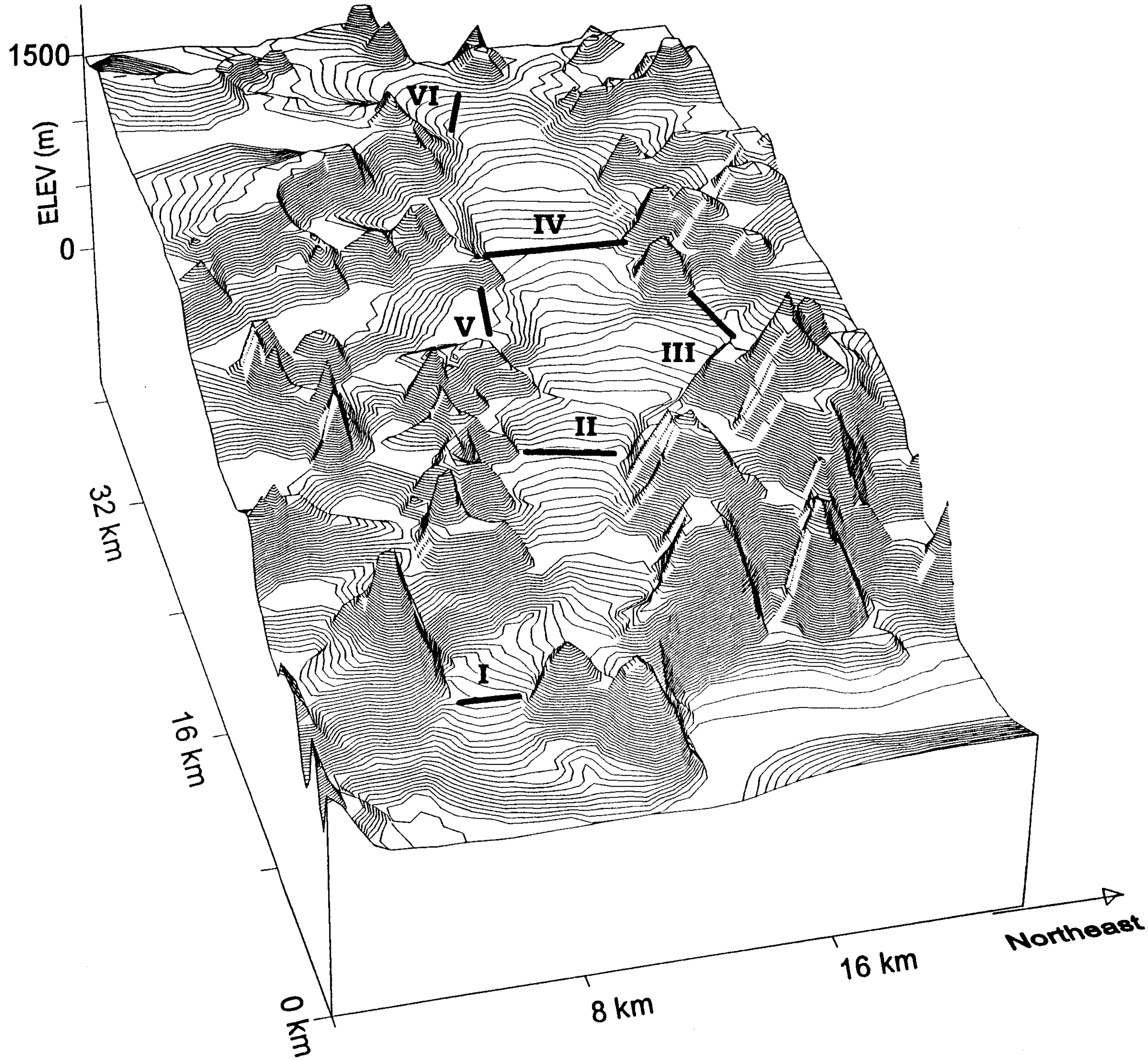
Fig 10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III
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V
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MacCaulay, H.A., and Hobson, G.D. 1972. A seismic refraction survey of the North Okanagan and South Shuswap Valleys. Geological Survey of Canada, Paper 72-8, Part A, pp. 1-8.

Mullins, H.T., and Hinchey, E.J. 1989. Erosion and infill of New York Finger Lakes: Implications for Laurentide ice sheet deglaciation. Geology, **17**: 622-625.

Mullins, H.T., Eyles, N., and Hinchey, E.J. 1990. Seismic reflection investigation of Kalamalka Lake: a "fiord lake" on the Interior Plateau of southern British Columbia. Canadian Journal of Earth Sciences, **27**: 1225-1235.

Newcomb, R.C. 1953. Seismic cross sections across the Spokane River Valley and the Hillyard Trough, Idaho and Washington: U.S. Geological Survey Ground Water Branch, Open File Report 53-199, 16 pp.

Nolan, M., Motyka, R.J., Echelmeyer, K., and Trabant, D.C. 1995. Ice thickness measurements of Taku Glacier, Alaska, USA, and their relevance to its recent behavior. Journal of Glaciology, **41**: 541.

Nye, J.F. 1952. The mechanics of glacier flow. Journal of Glaciology, **2**: 82-93.

Pardee, J.T. 1942. Unusual currents in glacial Lake Missoula, Montana. Geological Society of America Bulletin, **53**: 1569-1599.

Pelto, M.S., and Miller, M.M. 1990. Mass balance of the Taku Glacier, Alaska. Northwest science, **64**: 121-130.

Potter, C.J., Sanford, W.E., Yoos, T.R., Prussen, E.I., Keach, R.W., Oliver, J.E., Kaufman, S., and Brown, L.D. 1986. COCORP deep seismic reflection traverse of the interior of the North American Cordillera, Washington and Idaho: Implications for orogenic evolution. Tectonics, **5**: 1007-1025.

Poulter, T.C., Allen, C.F., and Miller, S.W. 1949. Seismic Measurements on the Taku Glacier. Stanford Research Institute.

Richmond, G.R. 1986. Tentative Correlation of deposits of the Cordilleran Ice-Sheet in the Northern Rocky Mountains, *in* G.M. Richmond and D.S. Fullerton, eds. Quaternary Glaciations in the United States of America: Quaternary Science Reviews, v.5 Quaternary Glaciations in the Northern Hemisphere, pp. 99-128.

Savage, C.N. 1965. Geologic History of Pend Oreille Lake Region in North Idaho, Idaho Bureau of Mines and Geology, Pamphlet 134, 17 pp.

Waitt, R.B. Jr, and Thorson, R.M. 1983. The Cordilleran ice sheet in Washington, Idaho, and Montana, *in* The Late Pleistocene; of Late-Quaternary environments of the United States, V.1, Minneapolis, Minnesota, Univ of Minnesota Press, pp.53-71.

LIST OF CAPTIONS

Fig 1. Location of the Juneau Icefield and Lake Pend Oreille.

Fig. 2. The geometry of fiord formation. The floating glacial ice is reshaped into a wedge by bouyant forces. Maximum erosion occurs where the floating ice, thickening upstream, becomes grounded. As the glacier recedes, the zone of maximum erosion moves upglacier.

Fig. 3. Map of north Idaho and vicinity showing the setting of Lake Pend Oreille. The hachured line represents the late Pleistocene extent of the Cordillerian ice sheet in Idaho. The cross hatched area shows the extent of glacial Lake Missoula and the arrows show the major routes of the Lake Missoula floods. The location of the seismic cross-section D-D' across Rathdrum Prairie is shown. Locations of the Lake Pend Oreille profiles A-A',B-B', C-C' and W-E are shown on the inset map.

Fig. 4. The 1966 seismic record section across Lake Pend Oreille basin. The location is shown as W-E on Figure 3. Roman numerals refer to seismic stratigraphic units discussed in the text. Horizontal scale approximately 5 kilometers.

Fig. 5. A 500 Hz seismic record section across Lake Pend Oreille basin in the vicinity of cross-section B-B'. Roman numerals refer to seismic stratigraphic units discussed in the text. Horizontal scale approximately 5 kilometers.

Fig. 6. Cross-sections derived from seismic data across Lake Pend Oreille (A-A', B-B', and C-C') and the Rathdrum Valley (D-D'). Profile D-D' modified from Newcomb, 1953.

Fig. 7. The Taku Glacier system and the locations of the seismic lines, identified by Roman numerals. Camp 10, the main JIRP facility, is at the northeast end of Profile IV.

Fig. 8. A typical seismic field record collected on the Taku Glacier.

Fig. 9. Digital image of a near 100% subsurface coverage seismic record section across the Taku Glacier along the center of profile IV near Camp 10.

Fig. 10. Cross sections of the Taku Glacier based on seismic results. See Figure 7 for the profile locations.

Fig. 11. The valley of the Taku Glacier as it would appear without ice or water.

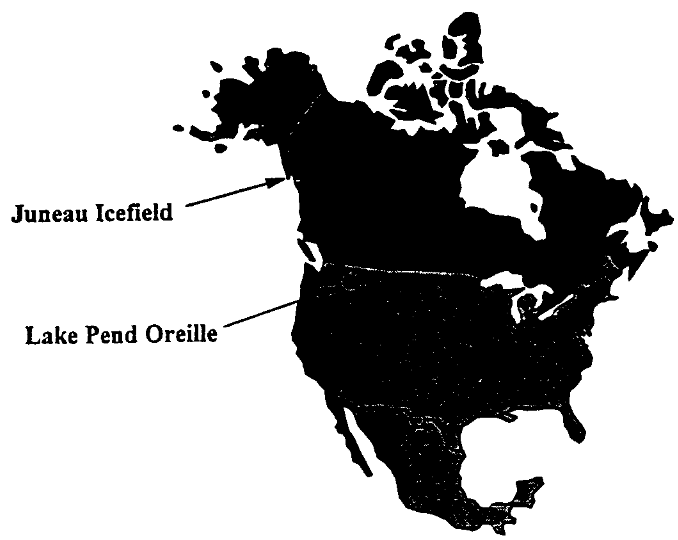
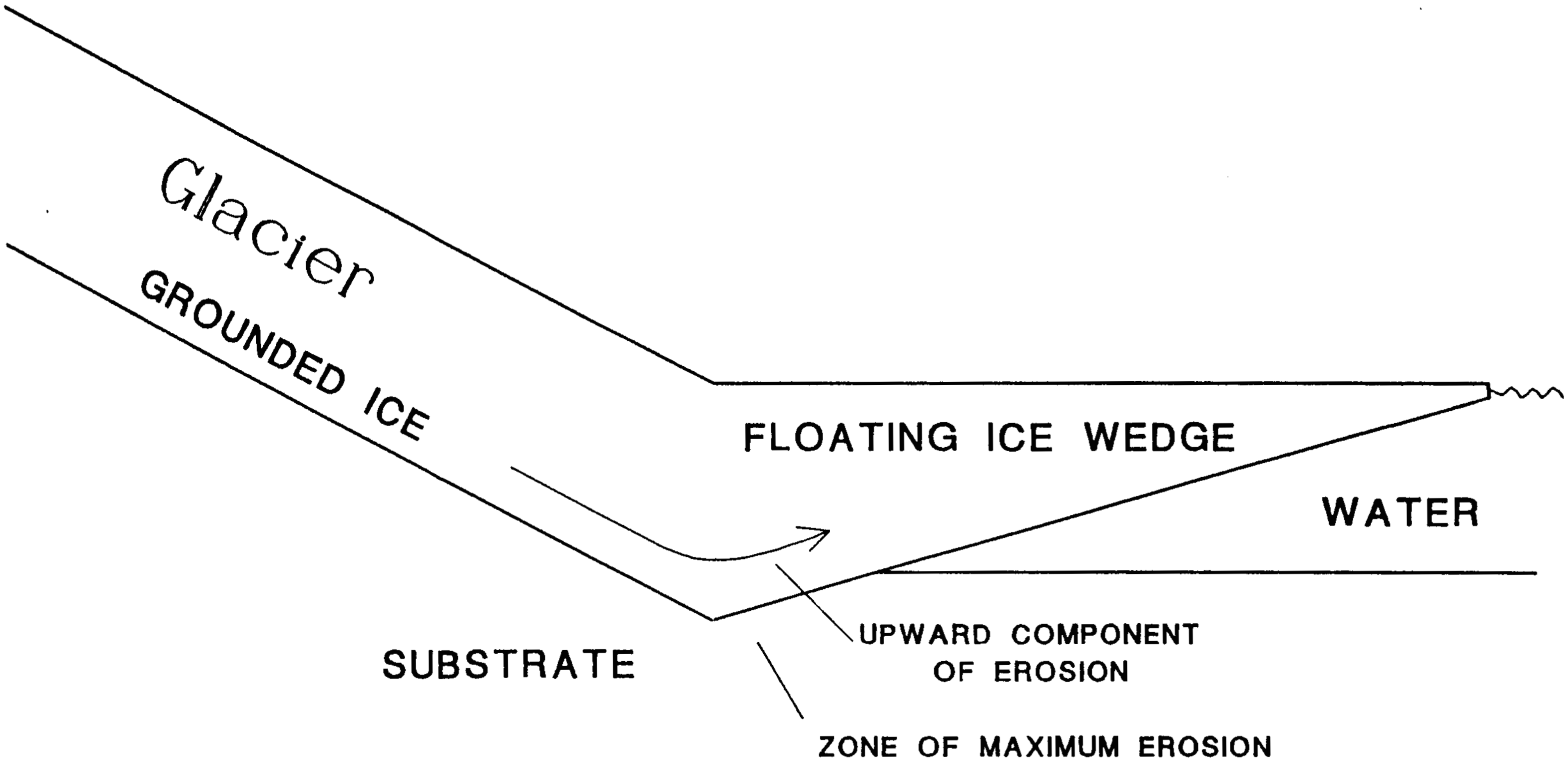


Fig 1



Glacier
GROUNDED ICE
FLOATING ICE WEDGE
WATER
SUBSTRATE
UPWARD COMPONENT OF EROSION
ZONE OF MAXIMUM EROSION

Fig 2

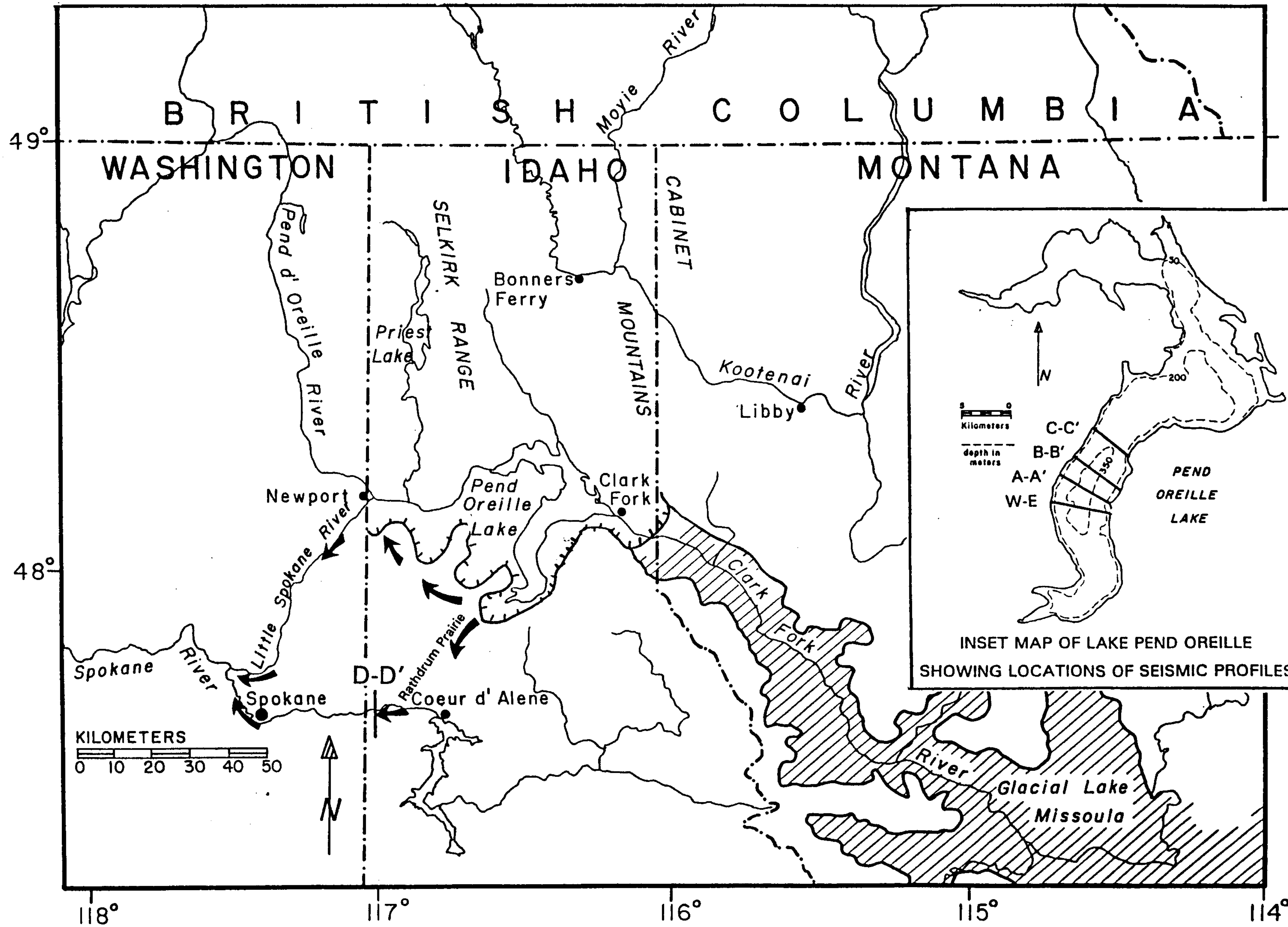

B R I T I S H C O L U M B I A
49°
48°
WASHINGTON
IDAHO
MONTANA
Moyie River
Pend d' Oreille River
SELKIRK RANGE
CABINET MOUNTAINS
Bonners Ferry
Priest Lake
Kootenai River
Libby
Clark Fork
Newport
Pend Oreille Lake
Little Spokane River
Rathdrum Prairie
Clark Fork River
Glacial Lake Missoula
Spokane River
D-D'
Spokane
Coeur d' Alene
KILOMETERS
0 10 20 30 40 50
N
118°
117°
116°
115°
114°
Kilometers
depth in meters
C-C'
B-B'
A-A'
W-E
30
200
350
PEND OREILLE LAKE
INSET MAP OF LAKE PEND OREILLE
SHOWING LOCATIONS OF SEISMIC PROFILES

Fig 3

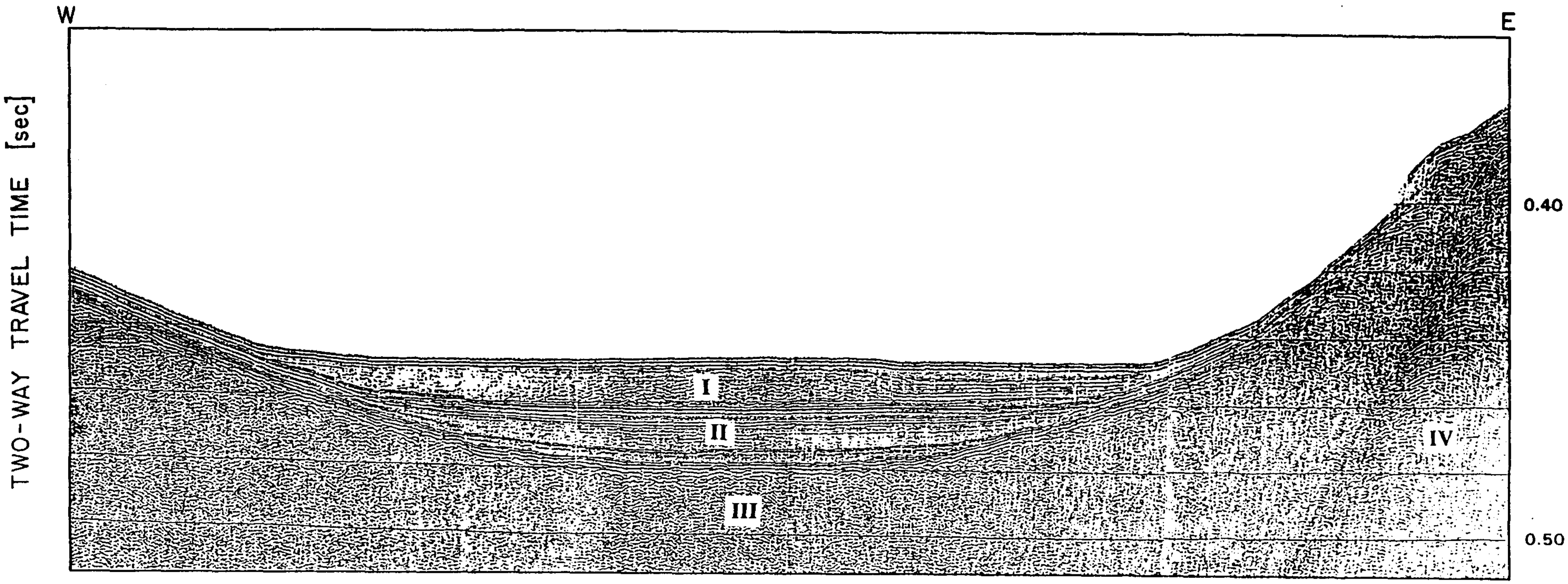


Fig 4

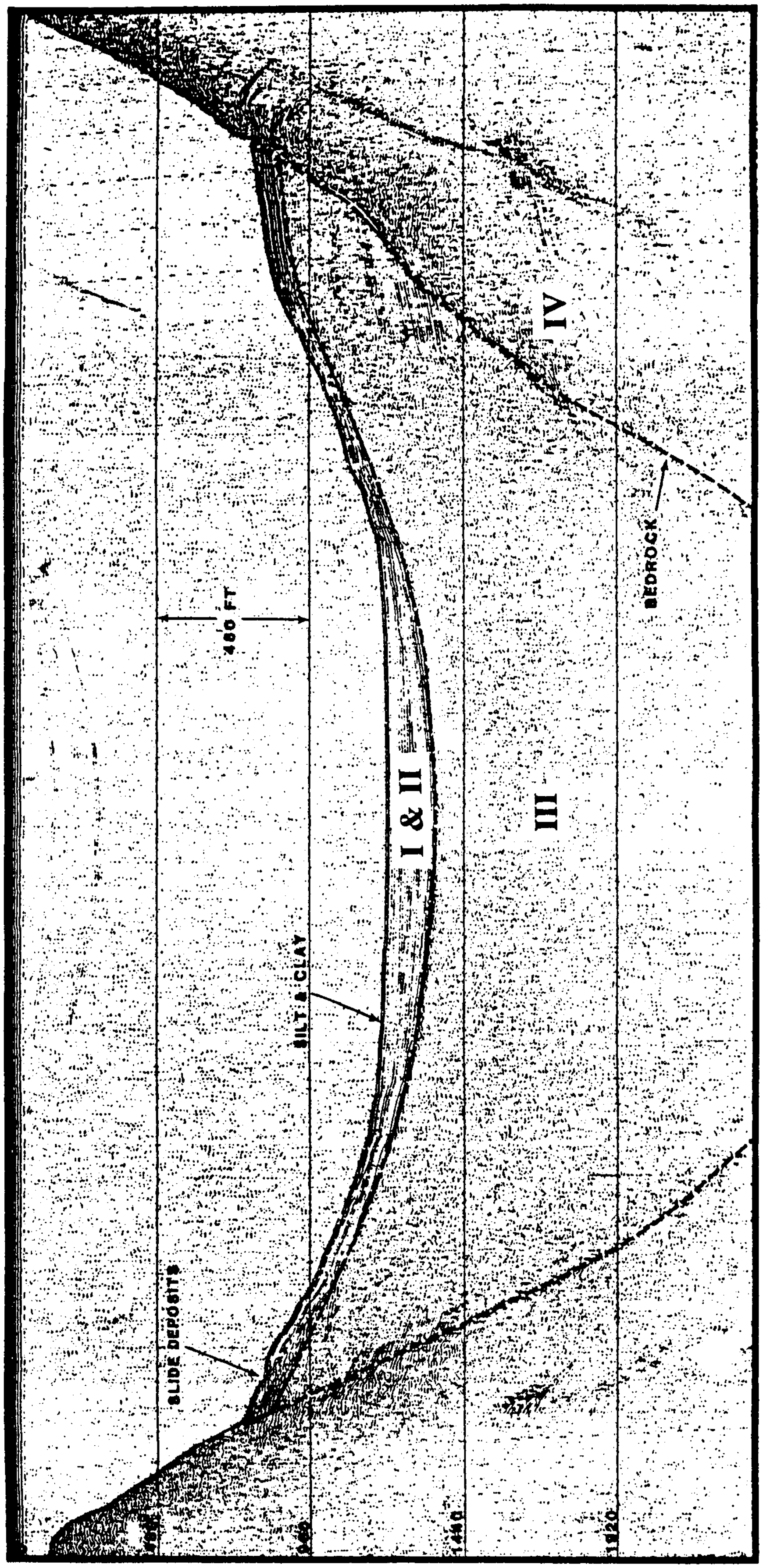
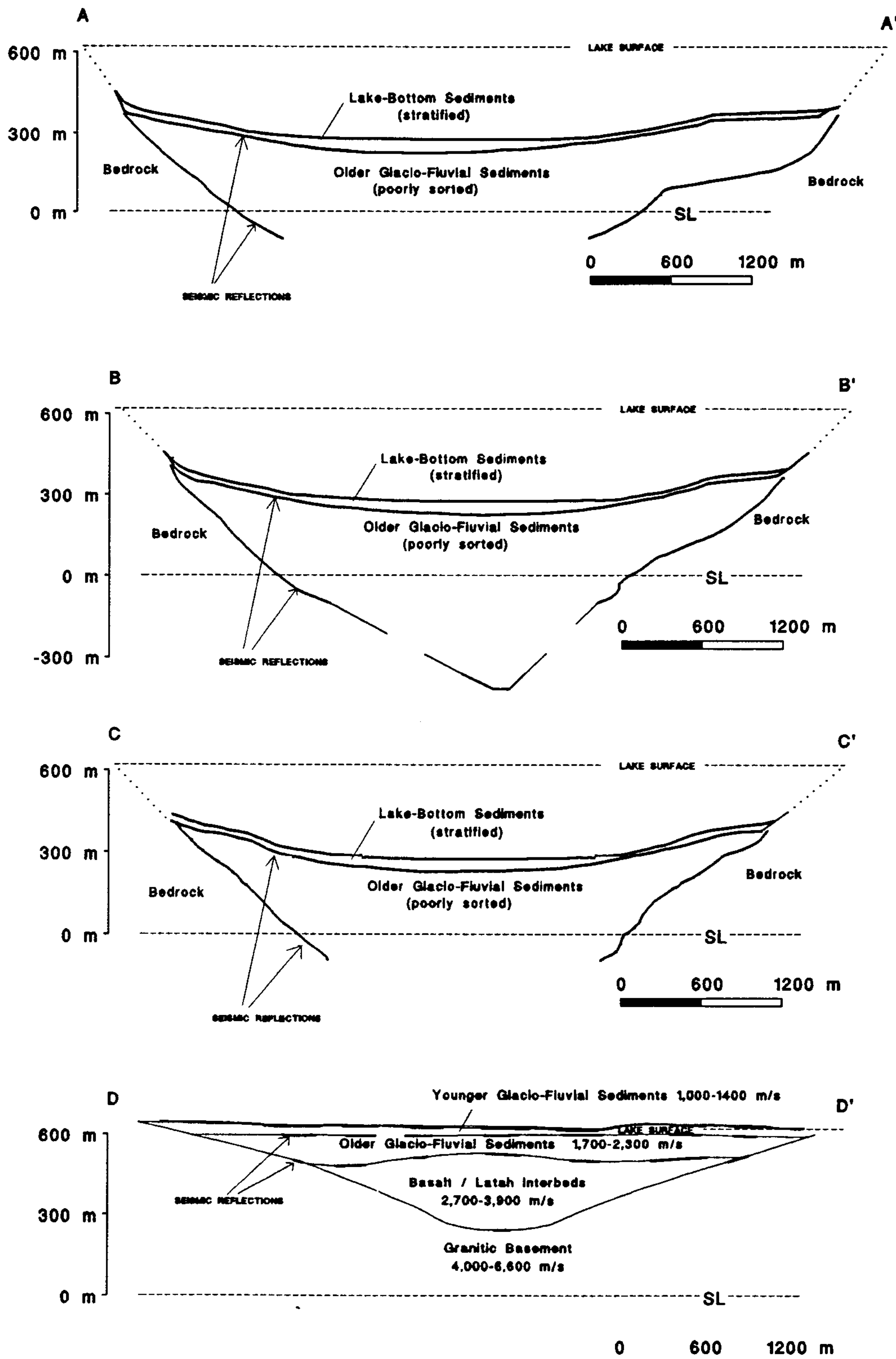


Fig 5



A
A'
600 m
300 m
0 m
LAKE SURFACE
Lake-Bottom Sediments
(stratified)
Bedrock
Older Glacio-Fluvial Sediments
(poorly sorted)
Bedrock
SL
0
600
1200 m
SEISMIC REFLECTIONS
B
B'
600 m
300 m
0 m
-300 m
LAKE SURFACE
Lake-Bottom Sediments
(stratified)
Older Glacio-Fluvial Sediments
(poorly sorted)
Bedrock
Bedrock
SL
0
600
1200 m
SEISMIC REFLECTIONS
C
C'
600 m
300 m
0 m
LAKE SURFACE
Lake-Bottom Sediments
(stratified)
Older Glacio-Fluvial Sediments
(poorly sorted)
Bedrock
Bedrock
SL
0
600
1200 m
SEISMIC REFLECTIONS
D
D'
Younger Glacio-Fluvial Sediments 1,000-1400 m/s
LAKE SURFACE
600 m
300 m
0 m
Older Glacio-Fluvial Sediments 1,700-2,300 m/s
SEISMIC REFLECTIONS
Basalt / Latah Interbeds
2,700-3,900 m/s
Granitic Basement
4,000-6,600 m/s
SL
0
600
1200 m

Fig 6

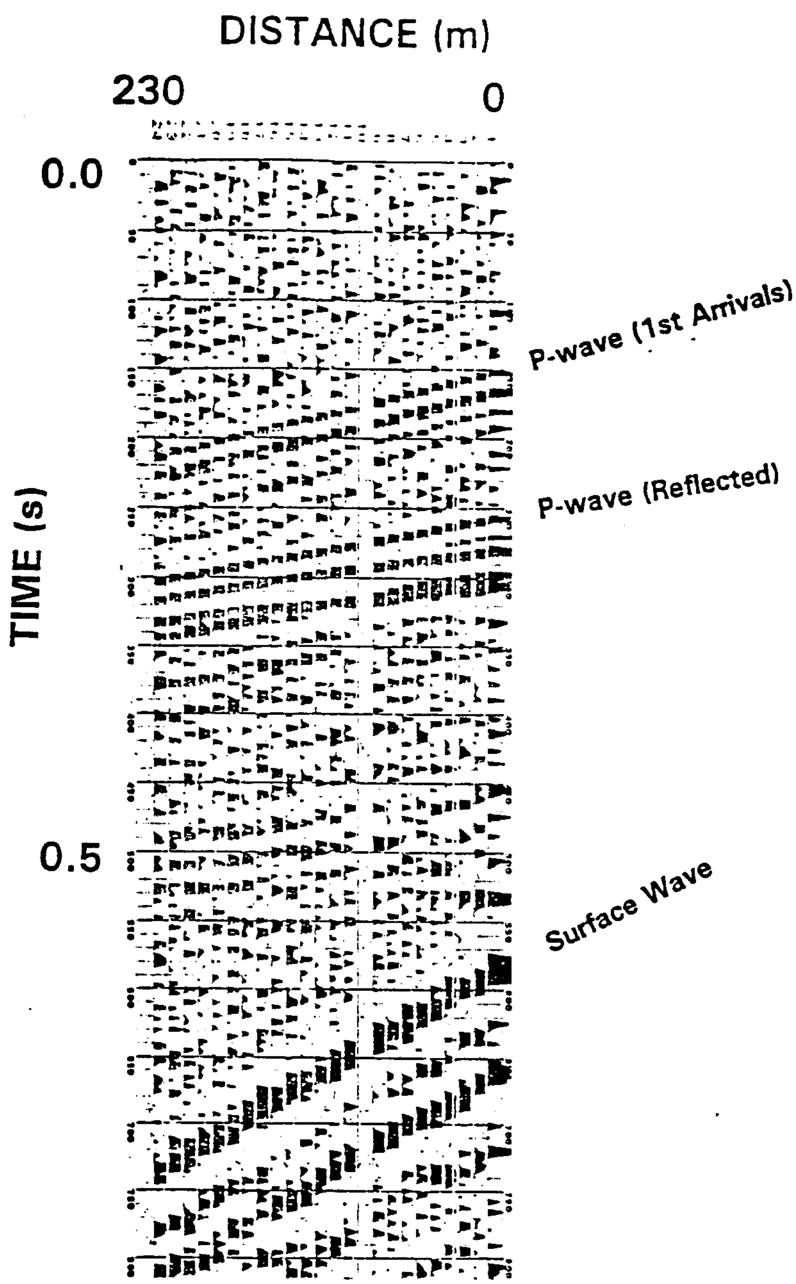

DISTANCE (m)
230
0
0.0
0.5
TIME (s)
P-wave (1st Arrivals)
P-wave (Reflected)
Surface Wave

Fig 8

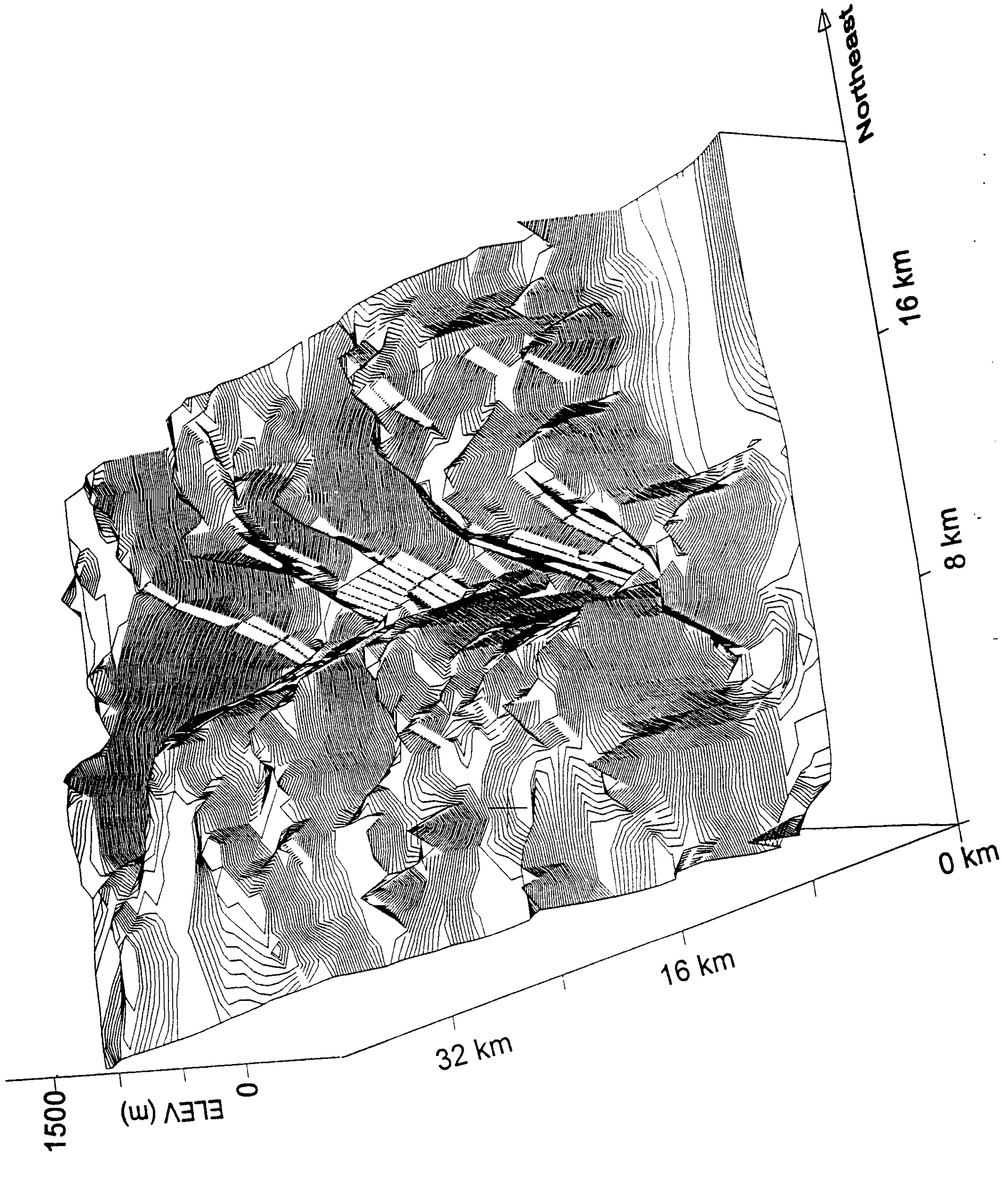
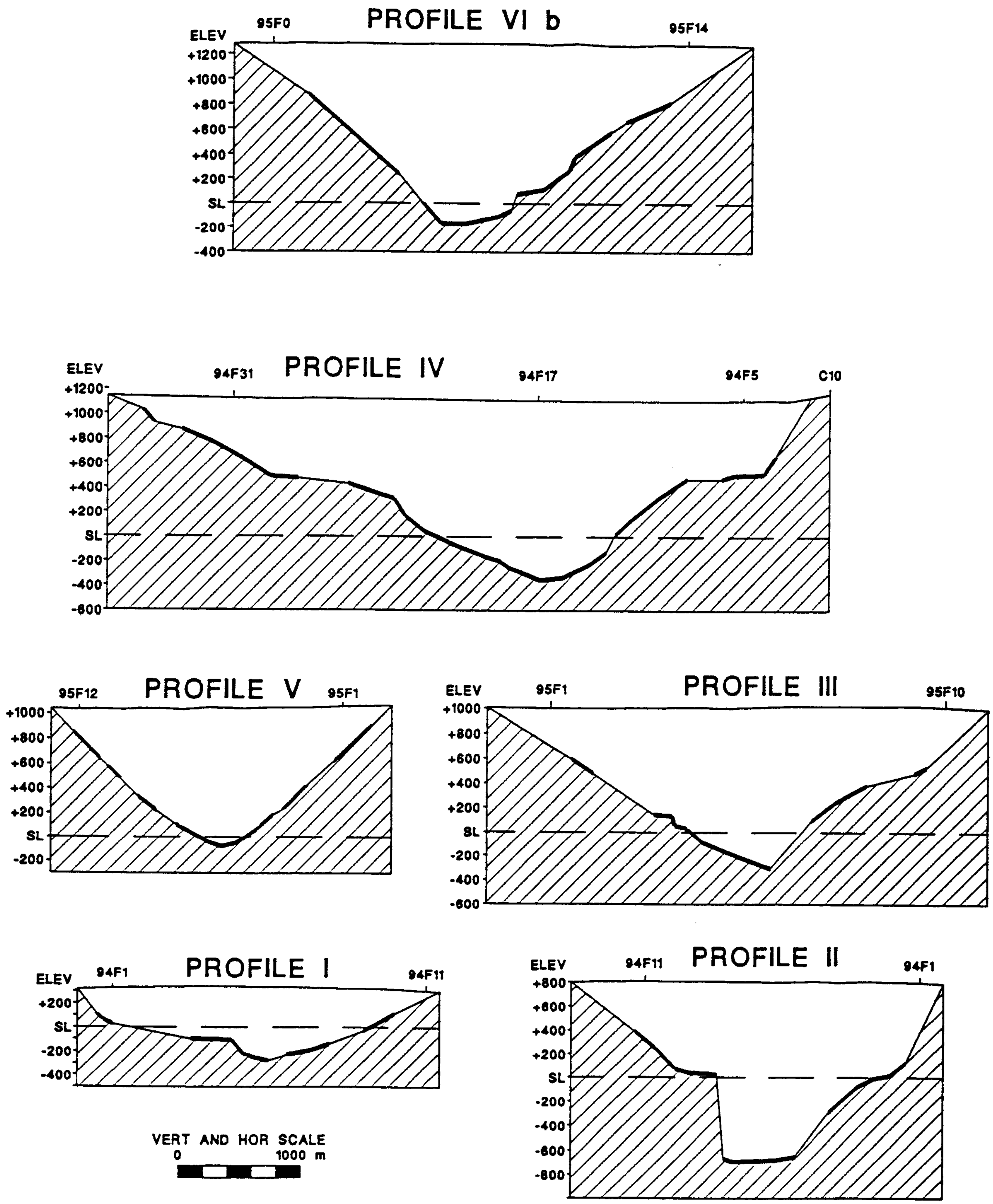
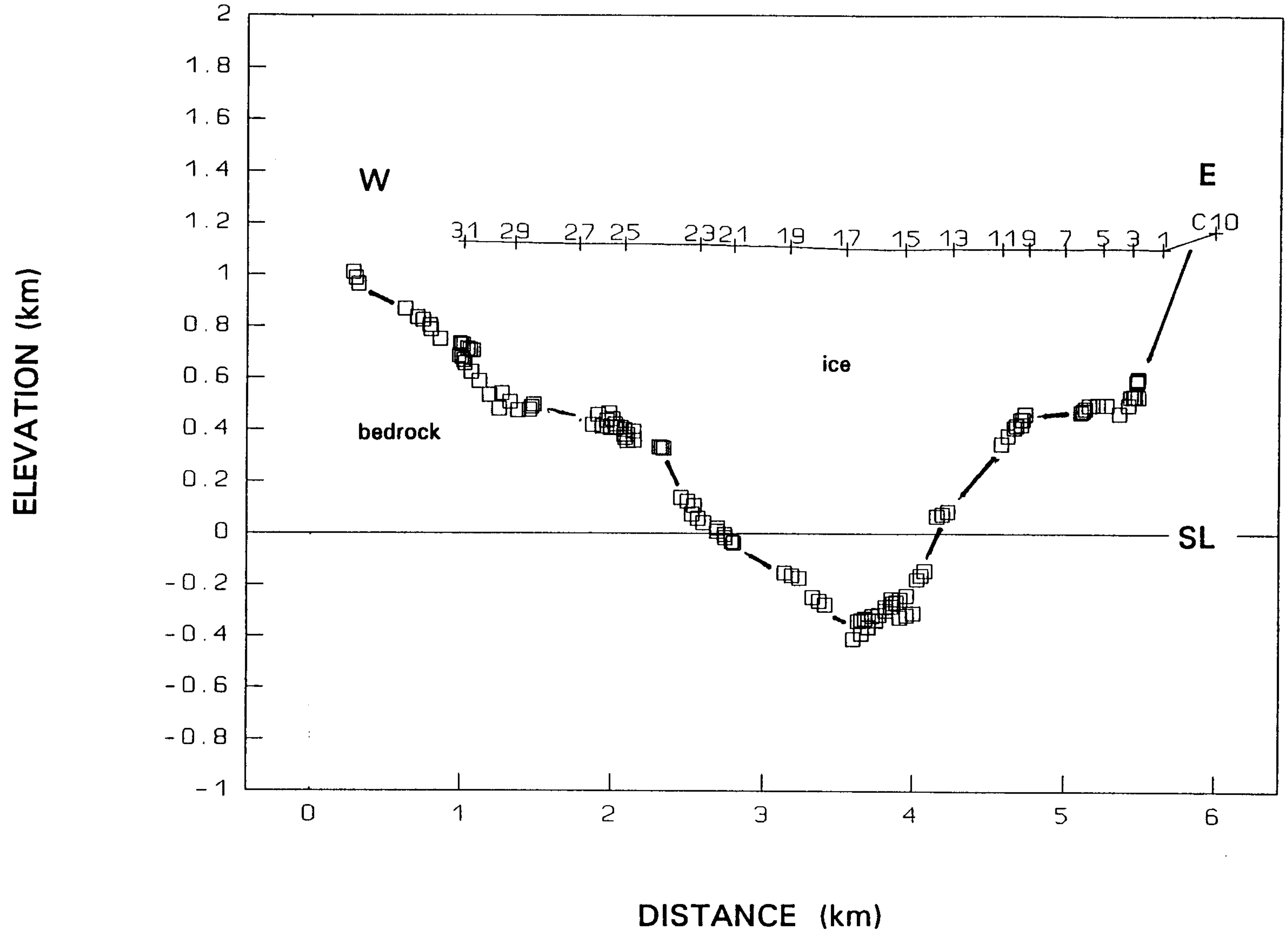
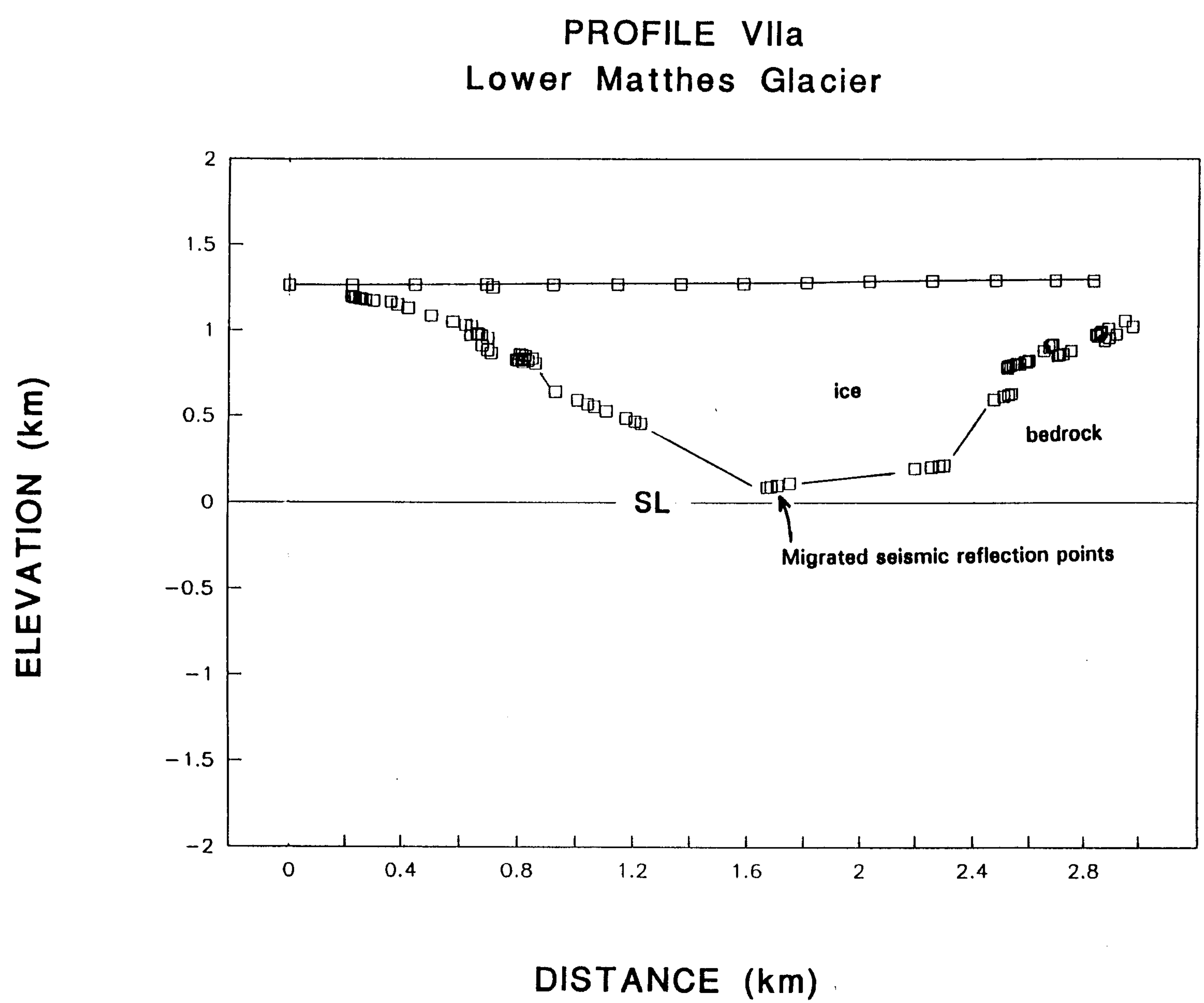


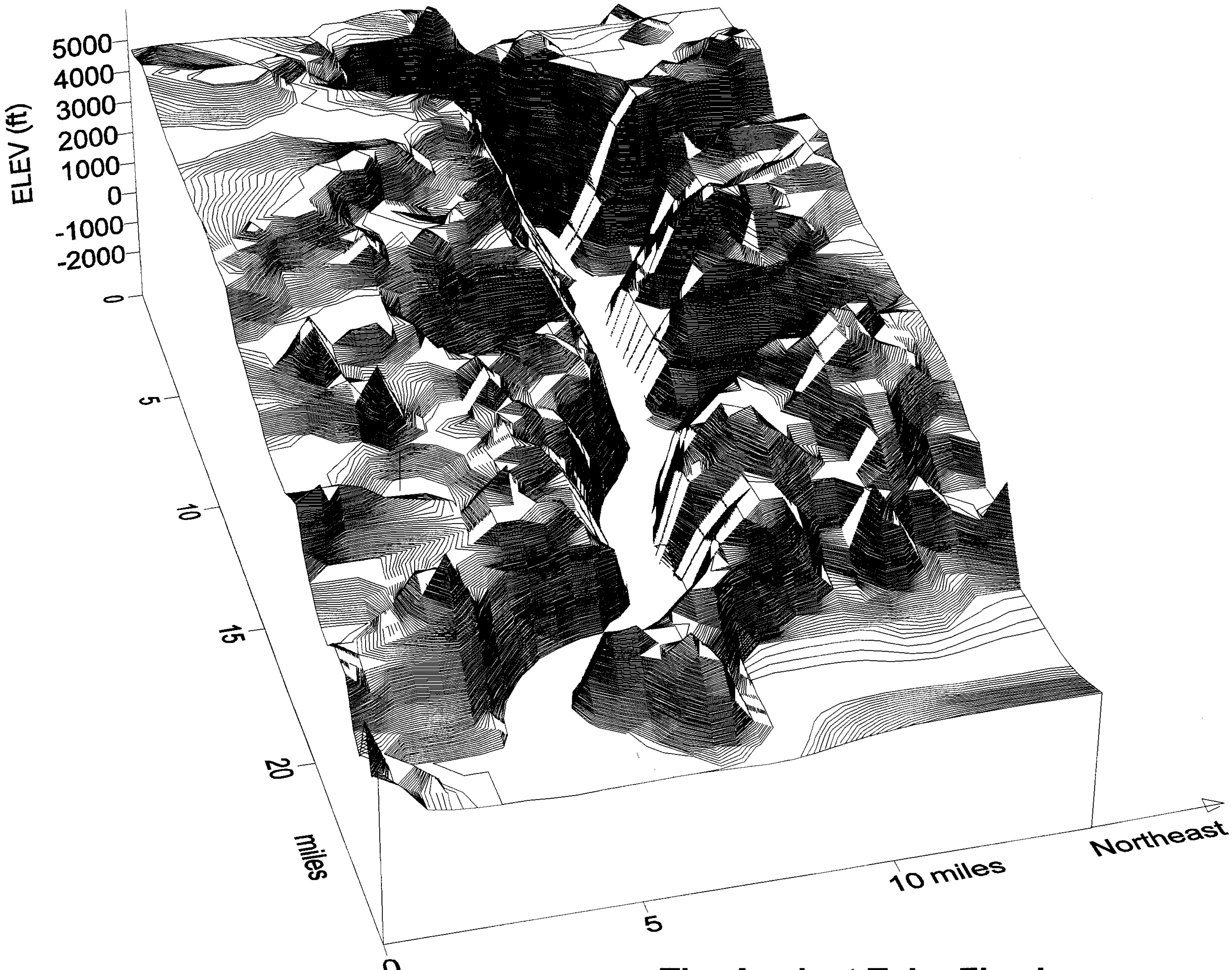
Fig 11


PROFILE VI b
ELEV
95F0
95F14
+1200
+1000
+800
+600
+400
+200
SL
-200
-400
PROFILE IV
ELEV
94F31
94F17
94F5
C10
+1200
+1000
+800
+600
+400
+200
SL
-200
-400
-600
PROFILE V
95F12
95F1
+1000
+800
+600
+400
+200
SL
-200
PROFILE III
ELEV
95F1
95F10
+1000
+800
+600
+400
+200
SL
-200
-400
-600
PROFILE I
ELEV
94F1
94F11
+200
SL
-200
-400
PROFILE II
ELEV
94F11
94F1
+800
+600
+400
+200
SL
-200
-400
-600
-800
VERT AND HOR SCALE
0
1000 m

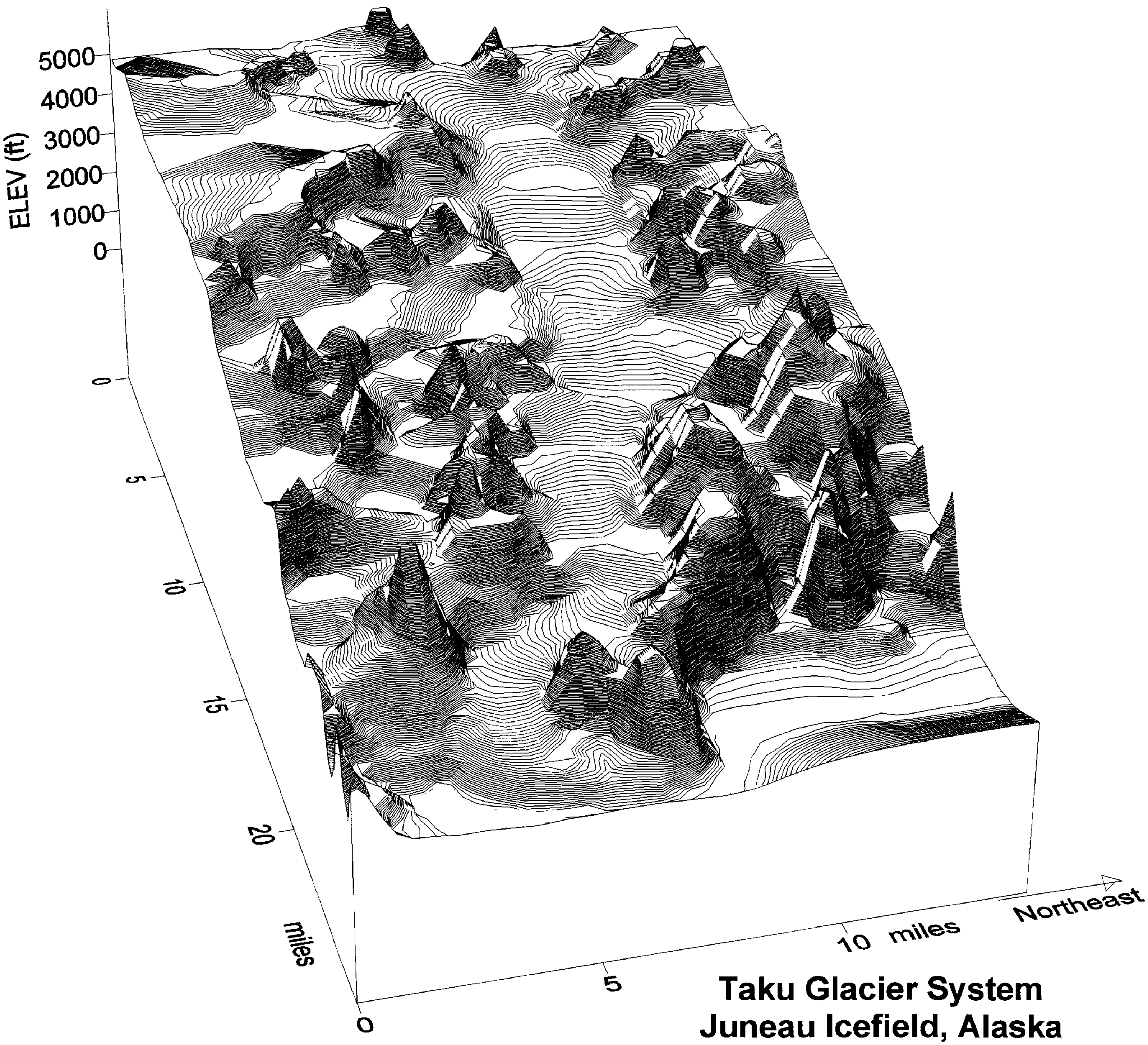
PROFILE IV
Taku Glacier at C-10



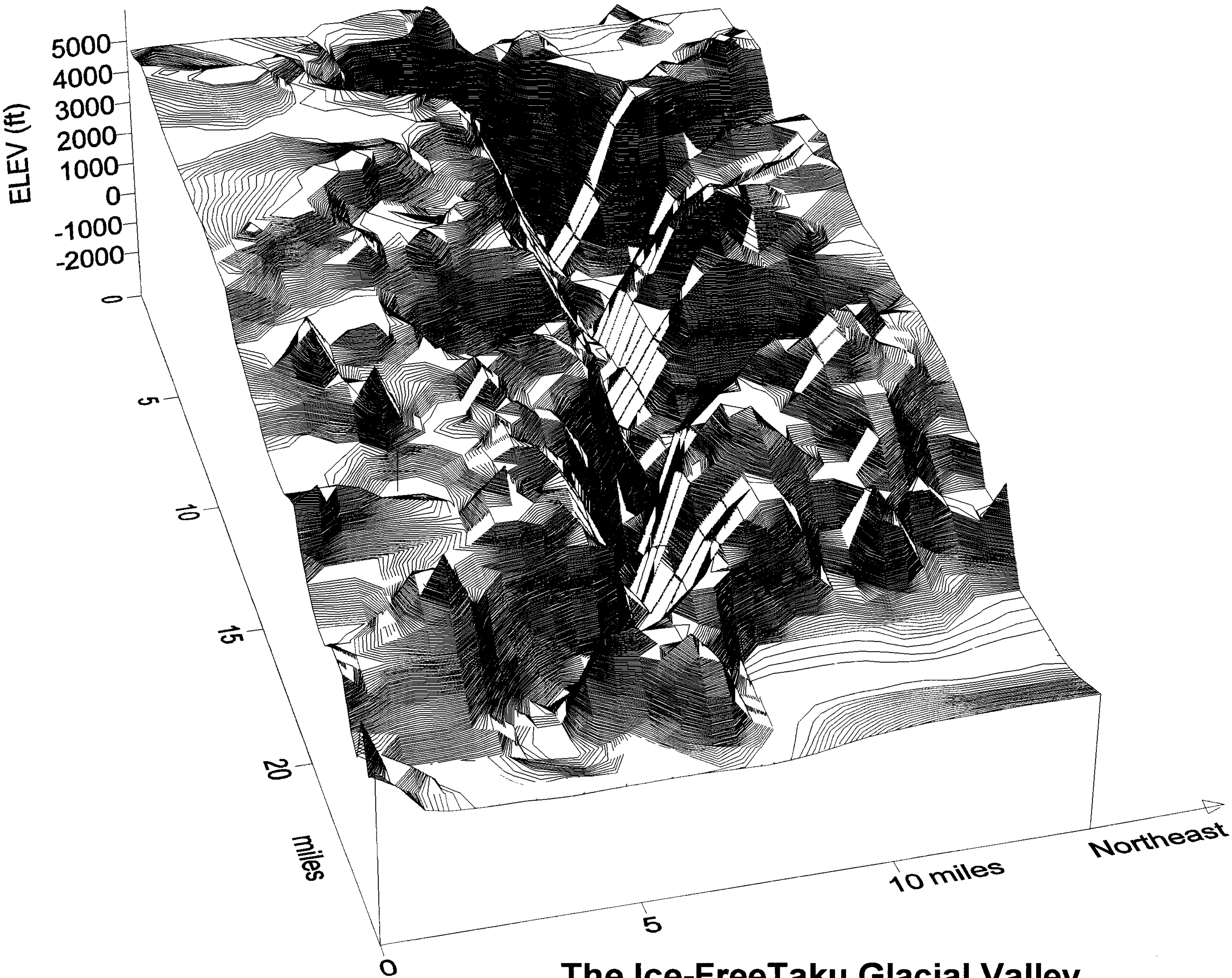

PROFILE VIIa
Lower Matthes Glacier
ELEVATION (km)
2
1.5
1
0.5
0
-0.5
-1
-1.5
-2
0
0.4
0.8
1.2
1.6
2
2.4
2.8
ice
bedrock
SL
Migrated seismic reflection points
DISTANCE (km)



The Ancient Taku Fjord



Taku Glacier System
Juneau Icefield, Alaska



The Ice-FreeTaku Glacial Valley