GEOMORPHOLOGY AND GLACIAL GEOLOGY OF THE METHOW DRAINAGE BASIN, EASTERN NORTH CASCADE RANGE, WASHINGTON

by

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A dissertation submitted in partial fulfillment

of the requirements for the degree of

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We have carefully read the dissertation entitled Geomorphology and Glacial Geology of the Methow Drainage Basin, Eastern North Cascade Range, Washington submitted by Richard Brown Waitt, Jr. in partial fulfillment of the requirements of the degree of Doctor of Philosophy and recommend its acceptance. In support of this recommendation we present the following joint statement of evaluation to be filed with the dissertation.

Waitt's careful and detailed study of the glacial record preserved in this major North Cascade drainage has led to some significant reinterpretations of the history of the last glaciation in Washington. He has convincingly reconstructed the extent and surface gradient of the massive Cordilleran Ice Sheet which entered Washington from Canada some 20,000 years ago, delineated its southern extent in considerably greater detail than was previously possible, and given us a clear picture of the sequence of events that were involved in deglaciation of this region. Among the more important of his contributions was the conclusion that the thick ice sheet in the northeastern Cascades disappeared largely by downwasting, rather than by backwasting, the mode of retreat previously inferred. Furthermore, he found evidence to show that residual alpine glaciers were not present in this part of the range following disappearance of the ice sheet, nor were they regenerated in the upper parts of alpine valleys during the closing phases of deglaciation. Waitt has been able to reconstruct the approximate limit of alpine glaciers prior to inundation by the Cordilleran Ice Sheet, making possible a comparison with alpine glacier systems farther south in the range that were not affected by the ice sheet. Because the glacial record of the Methow Valley is unique among the numerous drainage basins of the Cascade Range, important insights are offered into regional variations and interrelationships of glacier systems. The gathering of evidence required extensive foot travel into remote areas, some at high altitudes, and a proper assessment and evaluation of regional geologic relationships during the course of the study. Consequently, the study represents a systematic and well-reasoned investigation that has resulted in a body of data having varied and important implications to the study of Quaternary glaciation in North America and to future investigations into the environmental history of northern Washington.

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Abstract

GEOMORPHOLOGY AND GLACIAL GEOLOGY OF THE METHOW DRAINAGE BASIN, EASTERN NORTH CASCADE RANGE, WASHINGTON

by Richard Brown Waitt, Jr.

Chairman of Supervisory Committee: Professor Stephen C. Porter
Department of Geology

The Methow Valley, excavated in upturned Mesozoic sedimentary and volcanic rocks that have been downfaulted between crystalline blocks, is ideally positioned to have originated as a consequent stream on the floor of an early Tertiary graben. The boundary scarps, however, vary from resequent to obsequent, depending on the erosional resistance of rocks juxtaposed across the faults. The Methow-Pasayten lowland is therefore of erosional origin--a rift-block valley in the sense of Johnson, not a graben as defined by Suess.

Cirques, aretes, and U-shaped hanging troughs characteristic of alpine glaciation dominate the northern Methow landscape. However, the surface drift is ubiquitously of northern provenance, and it mantles high ridgecrests as well as valley floors. Clearly, ice-sheet glaciation postdated the most recent alpine advance. Because alpine drift and depositional landforms are absent in the Methow region, apparently having been eroded by the ice sheet, the maximum extent of alpine glaciers must be inferred wholly from such erosional evidence as cirque morphology and limits of U-shaped

cross-valley profiles. Thus reconstructed, alpine glaciers ranged from small, steep, independent glaciers in tributary valleys to long, gently inclined trunk glaciers in the main valleys.

Beveled-off cols and ridges, striations, drift, and hummocky topography at high altitudes are the primary modifications of the Methow uplands by the ice sheet. The freshness of the drift indicates that the last glaciation of the region approximately correlates with the Fraser Glaciation of western Washington and British Columbia. Ice-flow directional indicators pointing regionally south-southeast show that the ice sheet in the Methow valley was a sector of the Cordilleran Ice Sheet.

Striations, stoss-and-lee asymmetry, and erratics of northern provenance indicate that Cordilleran ice entered the Methow drainage basin not only through the Harts Pass area, but over cols at the heads of <u>all</u> major tributaries. Moreover, stoss-and-lee surfaces denoting the overflow of ice from Chelan tributaries into Methow tributaries reveal unequivocally that the ice sheet innundated the Chelan trough as well. The upper limit of glacial evidence, a surface sloping from 8000 ft in the northern Methow region to 3000 ft in Columbia Valley, indicates that the ice sheet over the northern areas was almost continuous, and broken only by a few scattered nunataks. The Methow Mountains, by contrast, were a continuous nunatak which separated the Methow and Chelan sectors of the ice sheet.

Deglaciation of the Methow region did not produce end moraines; instead, kame terraces, associated ice-marginal channels kettles, eskers, and other ice-contact landforms indicate that even the mountainous terrain deglaciated largely by downwasting and regional stagnation. Granodiorite erratics on the cirque floor at the head of the Pasayten West Fork show that the most recent glacier ice ascended the cirque. Striations on two cirque head-walls 20 mi apart further indicate that alpine glaciers were not significantly regenerated after the Cordilleran Ice Sheet melted. The absence of looped valley moraines throughout the region is complimentary evidence that a late-glacial alpine phase was lacking.

Odd as it is that the Cordilleran Ice Sheet dwindled without a rebirth of alpine glaciers. Goldthwait inferred a similar glacial sequence for the New Hampshire uplands. That identical relations between alpine and ice-sheet glaciations obtained <u>simultaneously</u> on opposite sides of the continent, regions affected by independent ice sheets and nourished by different sources of moisture, indicates that glaciers fluctuated not in response to glacier dynamics, but to changes of climate. Apparently an abrupt, popular, and "final" amelioration of climate struck the 40° to 50° latitudes in the Western Hemisphere about 14,000 years ago.

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Acknowledgements

I wish to acknowledge the generous time and assistance each of the following persons has given me during this study. Prof. Stephen C. Porter suggested and supervised the project. His advise on the content and presentation are manifest on most of the following pages. Prof. Julian D. Barksdale, patriarch of Methow Valley geology, was of indispensable help in elucidating unpublished results of his many years of study of the bedrock geology. These two gentlemen and Dr. John T. Whetten read and criticized the manuscript, which has thus benefited greatly. From time to time I have benefited by discussions related to my project with Drs. E. C. Cheney, L. G. Hanson, V. S. Mallory, R. J. Stewart, and H. E. Wheeler, and with graduate students M. Cole, T. C. Pierson, and M. E. Tennyson, the latter three of whom were occasional companions during the field study. To my wife, Cynthia, I am perhaps not thankful enough for her many patient hours of typing and help with illustrations; to her family, the Vernon Wagners, I am thankful for everything.

Financial aid for this project was received from the Penrose Bequest of the Geological Society of America, from the Society of the Sigma Xi, and from the Corporation Fund of the Department of Geological Sciences, University of Washington. The U. S. Forest Service in Winthrop was generous in supplying maps, information on weather and on condition of trails, as well as making available space for occasional office work and for storage.

All in a golden afternoon
Full leisurely we glide;
For both our oars, with little skill
By little arms are plied,
While little hands make vain pretence
Our wanderings to guide.

—Lewis Carroll

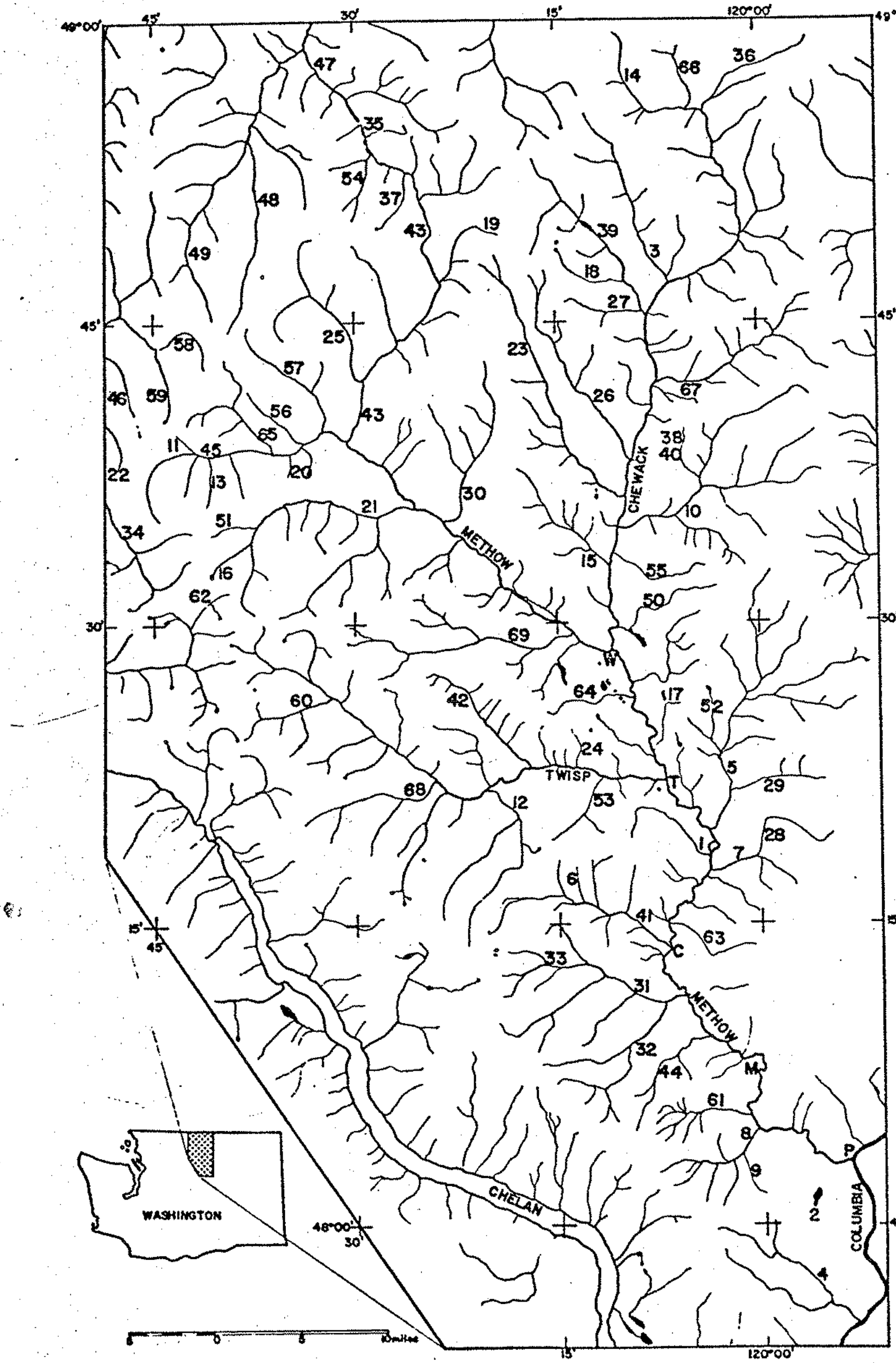
INTRODUCTION

Statement of Problems

As a questioned dashed line on Crandell's compilation (1965, Fig. 2) indicates, the southern and upper limits of the Late Wisconsin Cordilleran Ice Sheet, though well known in the Puget Lowland and Columbia Plateau (Bretz, 1913; Carson, 1970; Crandell, 1965; Flint, 1935, 1937; Richmond, *et al.*, 1965), remain conjectural in the northern Cascade Range.

Cary and Carlston (1937), Mackin (1941), and Vance (1957, p. 292-299) demonstrated that alpine glaciers in the western Cascades dwindled or disappeared prior to the arrival of the Puget Lobe of the Cordilleran Ice Sheet. In the eastern Cascades, to the contrary, Waters (1933) implied synchroneity of valley ice descending Chelan trough and a tongue of the Okanogan Lobe ascending the trough from the east, a thesis subscribed to by Whetten (1967, p. 259). The conflict rests unresolved; spatial and temporal relations between alpine and the Ice Sheet glaciations remain unestablished in the northwestern Cascades.

Obviously unsuited to a study of the relationship of alpine and ice-sheet glaciers is the submerged floor of Chelan trough. The Methow Valley (Fig. 1), however, is accessible, has valley-floor moraines, and has been affected by the ice sheet (Barksdale, 1941a) as well as by alpine glaciers. The Methow bedrock includes a greater variety of distinctive lithologies than do other valleys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in the northern Cascades, and because bedrock mapping is well advanced on 1:125,000 to 1:62,500 scales, the provenance of drift stones can be accurately determined.

Deglaciation of northern Washington followed two distinctively different modes. Deglaciation of the eastern Cascades affected strictly by alpine glaciers was by back-wasting of ice tongues that experienced moraine-building stillstands or readvances (Hopkins, 1966; Long, 1951; Merrill, 1966; Page, 1939; Porter, 1969); regions of the Okanogan Valley and Waterville Plateau affected solely by the Cordilleran Ice Sheet apparently deglaciated by down-wasting and stagnation (Flint, 1935, p. 180; 1937; Hanson, 1970, p. 88-91). In the Methow Valley Barksdale (1941a) correlated moraines that appeared to be progressively younger upvalley with moraines described by Page (1939) in Wenatchee Valley near Leavenworth. The flat-topped, sandy gravelly Methow Valley "moraines", however, unlike the nested arcuate moraines in the Leavenworth area, are associated with kame terraces, ice-marginal channels, and with laminated silt deposited in ice-dammed lakes. In contrast, Cascade valleys south of Chelan trough are nearly free of such stagnant-ice deposits.

The three major objectives of the present study of the glacial geology of the Methow Drainage Basin were 1) to determine limits and surface profiles of the Cordilleran Ice Sheet, 2) to ascertain spatial and temporal relations between the ice-sheet and Cascade alpine glaciers, and 3) to examine and compare patterns of deglaciation in the northern Cascades with those of Cascade Valleys farther south and of the Columbia Plateau to the east.

Previous Work

Study of the Cordilleran Ice Sheet in the Okanogan and Columbia Valleys has a long history. In the pioneering report Willis (1887) recognized the terminal moraine on the Waterville Plateau and postulated the Grand Coulee as an ice-deranged former course of the Columbia River. Russell (1893, 1898, 1899) and Dawson (1898) advanced arguments for the late-glacial origin of the Columbia Great Terrace and for an enormous late-glacial Lake Lewis in north-central Washington; Salisbury (1901) roughly traced the drift border east of Chelan Falls.

Working mainly east of the Grand Coulee, Bretz (1923, 1932) elucidated the origin of the Grand Coulee and the Channeled Scabland farther east, invoking a great flood as the agent of excavation. Flint (1935, 1936, 1937, 1938) and Flint and Irwin (1939) deciphered flow patterns, upper limits, and surface profiles of the ice sheet east of the Methow Drainage Basin. Waters (1933) elaborated on the ice-sheet limits, on deglacial patterns, on the distribution of the major ice-marginal coulees and kame terraces in Columbia and Chelan Valleys.

Students of the Canadian Rockies and Coast Ranges have long recognized that the local alpine glaciers eventually coalesced on the high plateau of interior British Columbia to form the central reservoir for the Cordilleran Ice Sheet (Flint, 1971, Fig. 18-1; Prest, 1970, p. 754). Outlet glaciers discharged westward along valleys transecting the Coast Ranges (Armstrong and Brown, 1954; Armstrong and Tipper, 1948) and south along each side of the Cascade Range, thereby invading the Puget and Okanogan lowlands. Although the greatest flux of the Late Wisconsin (Fraser) Cordilleran Ice-Sheet was into the lowlands, Cascade peaks along the 49th parallel were buried to an altitude of 8000 feet (Daly, 1912, p. 594; qualified by Barksdale, 1941a, p. 729). Striations show that this ice sheet flowed generally south into the North Cascades of Washington (Brock, 1902; Rice, 1947, Map 889A).

Major drift sheets separated by nonglacial deposits record at least four advances of the Cordilleran Ice Sheet down the Frazer and Puget lowlands (Armstrong, et al., 1965). In southern British Columbia nonglacial sediments dated between 43,800 and 19,100 radiocarbon years B.P. separate the youngest two drift sheets (Fulton, 1968, 1971; Prest, 1970, p. 704-705). The abundance of stratified drift and complex sequences of ice-dammed lake sediments (Fulton, 1965, 1967, 1969; Mathews, 1944; Nasmith, 1962) indicates that the most recent deglaciation as in the Columbia Plateau was mainly accomplished by downwasting.

In the Methow Valley Russell (1898, Pl. 18) incorrectly placed the terminal position of the Methow Glacier at a conspicuous "moraine" near Winthrop. The only serious glacial geology in the Methow region, however, was Barksdale's (1941a,b) reconnaissance of late-glacial alpine moraines and of ice-sheet invasion of Methow headwaters. Excepting Barker's (1968), brief report on the Chelan area, no glacial geology has been attempted in the northeastern North Cascade Range since Barksdale's study.

Methods of Study

Data for this report were gathered in 1970-1971 during five and a half months of foot traverses concentrated in the Methow headwater regions, along higher ridges and summits to the southeast, and in selected parts of the lower Methow Valley and its tributaries. Field investigations were supplemented by topographic analyses of 15- and $7\frac{1}{2}$- minute topographic sheets and on aerial photographs at scales of 1:40,000, 1:30,000, and 1:15,800. In addition, critical areas were studied on 1:12,000 aerial photographs kindly made available by the U. S. Forest Service in Winthrop. Altitudes were determined with a Paulin altimeter until preliminary $7\frac{1}{2}$-minute topographic sheets became available during the second field session.

General Geologic and Geographic Setting

Crystalline rocks of the northeastern Cascades are interrupted by a 20-mile-wide northwest-trending belt of upturned clastic sedimentary rocks underlying the Methow Valley (Fig. 2). The sedimentary block is generally of lower altitude than the crystalline blocks between which it is inset along near-vertical faults. Most large tributaries head in one of the crystalline blocks and descend to the trunk Methow River, which flows southeast along the general axis of the sedimentary block. The lower 18 mi of the Methow River, having emerged from the sedimentary block, flows over crystalline rocks.

The Methow block is but one of several northwest-trending sedimentary-rock belts in which valley systems are well developed. The Yakima, Wenatchee, and Methow Valleys--each developed mainly in sedimentary rocks--are broader and have more extensive tributary networks than valleys like Chelan and Entiat that are excavated in crystalline rocks. On a regional scale, therefore, major streams of the eastern Cascades streams are remarkably well adjusted to bedrock structure.

BEDROCK GEOLOGY

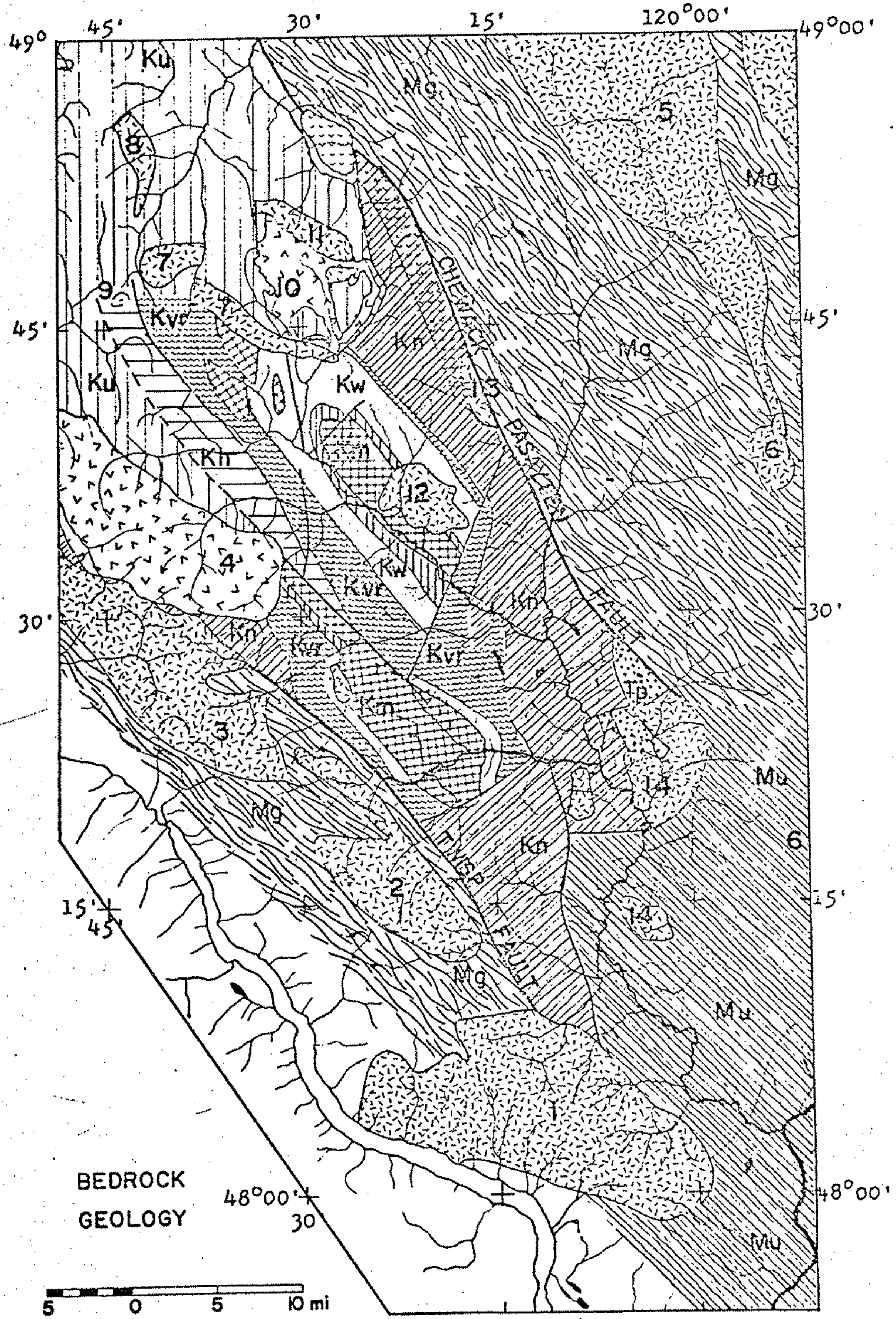
Rock Types and Stratigraphic Relations

CRYSTALLINE ROCKS EAST AND WEST OF THE METHOD SEDIMENTARY BLOCK

Except where intruded by younger plutons (Fig. 2 and Table 2) the upland blocks east and west of the Methow sedimentary block are mostly medium-grained mesocratic quartz-dioritic gneiss that is weakly to conspicuously foliated northwest, parallel to the ambient trend of North Cascade structural axes. The gneiss is generally more melanocratic, and even schistose or phyllitic southwest of the upper Twisp Valley, as it is locally east of the Chewack-Pasayten Fault. A weakly foliated phase of the gneiss in the upper Chewack region contains conspicuous poikiloblastic K-feldspar porphyroblasts.

CRETACEOUS TO PALEOCENE SEDIMENTARY AND VOLCANIC ROCKS OF THE METHOW BLOCK

Within the fault-bounded Methow sedimentary block a 40,000-foot package of andesite, volcanic graywacke, plagioclase arkose, chert litharenite and conglomerate, and black mudstone (Table 1) has been tightly folded into a syncline plunging gently southeast, parallel to the two boundary faults (Fig. 2; Barksdale, 1968; Pitard, 1958). Although rock units change along strike because of unconformities, facies changes, and cross-faults, correlative rocks have been mapped--and different names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applied--at the northwestern end of the sedimentary block in the Manning Park-Hope area of British Columbia (Cairnes, 1944; Coates, 1970; Daly, 1912; Rice, 1947). Fossils establish an Early- to Late-Cretaceous age of the tightly folded section (Barksdale, 1948, 1960, 1968; Coates, 1970; Maurer, 1958; Pierson, 1972), and a Paleocene age (Royce, 1965) for a less-deformed conglomerate member unconformably overlying the Cretaceous rocks (Ryason, 1954). Potassium-argon ages of plutons that have intruded the sedimentary and older rocks range, according to Kulp's (1961) time scale, from Late Cretaceous to Late Eocene (Tabor, et al., 1968, Table 1).

YOUNGER PLUTONIC ROCKS

Quartz diorite to granite plutons variously K-Ar dated as Middle Cretaceous to Late Eocene (Tabor, et al., 1968, Table 1) intrude both the Methow sedimentary block and the adjacent crystalline blocks. All of the plutons are discordant to sedimentary-rock contacts; most have contact-metamorphic aureoles and complex dike systems at the margins (Tabor, et al., 1968, p. 46, 49); some are elongate parallel to the ambient northwest structural grain, have disrupted fold axes in the host rocks, and have caused crowding of folds at the intrusive margins (Tabor, p. 51). The plutons, therefore, intruded both forcefully and by stoping; and they are clearly younger than the upturned Mesozoic rocks.

TABLE 1

Sedimentary rocks of the Methow Inset

(adapted from Barksdale, 1968)

LOWER CRETACEOUS

Newby Group

Description: porphyritic gray andesite and andesite breccia; black mudstone and shale; immature volcanic litharenites; local conglomerate lenses.

Diagnostic lithologies: andesite in drift of Eightmile and Falls Creeks, in drift farther southeast not distinguishable from Midnight Peak andesite.

MIDDLE CRETACEOUS

Goat Creek (lowest), Panther Creek, Harts Pass (highest) Formations

Description: interbedded submature plagioclase arkose and black shale, with rare beds of quartz diorite pebble- and mudstone granule-conglomerate.

UPPER CRETACEOUS

Virginian Ridge Formation

Description: submature chert litharenite and chert-pebble conglomerate; black shale and mudstone.

Diagnostic lithologies: pebble to granule conglomerate of gray to white chert; mature gray chert litharenite; gray pebbly chert litharenites.

Table 1 cont.

Winthrop Formation

Description: gray mature plagioclase arkose, tightly cemented.

Diagnostic lithology: arkose is characteristic, not diagnostic.

Ventura Formation

Description: red-purple to red-brown sandstone and mudstone; intermittent beds of granule to small-pebble conglomerate.

Diagnostic lithologies: red sandstone; red granule to small-pebble chert conglomerate; particolored red, green, and white sandy granule chert conglomerate.

Midnight Peak Formation

Description: gray, greenish-gray, and purplish-gray porphyritic andesite and andesite breccia.

Diagnostic lithologies: porphyritic andesite and andesite breccia.

TERTIARY--PALEOCENE

Pipestone Canyon Formation

Description: basal sandy conglomerate of quartz-diorite pebbles and cobbles grades upward to brown arkose with shale interbeds; massive to foliated quartz diorite pebbles similar to adjacent crystalline block; some pebbles of gray chert, andesite, and dacite in upper conglomerate layers.

TABLE 2

Plutonic Rocks

METHOW-CHELAN MOUNTAINS

1. Cooper Mountain massive leuco-mesocratic* biotite granodiorite (Barksdale, 1968).**

\# 2. Oval Peak foliated mesocratic hornblende quartz diorite (Adams, 1961).

3. Black Peak massive mesocratic biotite-hornblende quartz diorite ((Late Cretaceous))*** (Misch, 1966).

\# 4. Golden Horn massive leucocratic biotite-hornblende miarolitic megacrystic (K-feldspar) granite ((Late Eocene)) (Stull, 1969).

OKANOGAN MOUNTAINS

5. Cathedral Peak granodiorite and quartz monzonite ((Mid-Cretaceous)) (Daly, 1912, p. 459-464; Goldsmith, 1952; Hawkins, 1968).

6. Various quartz diorite, granodiorite, and quartz monzonite stocks (Goldsmith, 1952; Menzer, 1964).

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* Color Index 0-5, leucocratic; 5-15, mesocratic; 15- , melanocratic.

** Reference for detailed description.

*** Potassium-Argon age according to time scale of Kulp (1961).

\# Rocks particularly useful as indicators of drfit provenance.

Numbers correspond to stocks outlined on Figure 2.

Table 2 cont.

METHOW INSET

\# 4. Golden Horn granite (see above).

\# 7. Pasayten Peak leucocratic biotite granodiorite to melanocratic biotite-hornblende quartz diorite ((Late Cretaceous)) (Tabor, <u>et al</u>., 1968).

\# 8. Rock Creek mesocratic biotite-hornblende quartz diorite to granodiorite ((Late Cretaceous)) (Tabor, <u>et al</u>., 1968).

9. Tamarack Peak melanocratic hornblende diorite.

\# 10. Monument Peak leucocratic biotite-hornblende miarolitic megacrystic (K-spar) granite ((Late Eocene)) (Tabor, <u>et al</u>., 1968).

11. Lost Peak mesocratic biotite-hornblende granodiorite to quartz monzonite (Tabor, <u>et al</u>., 1968).

12. Goat Creek melanocratic diorite (Barksdale, 1968; Pitard, 1958).

13. Button Creek mesocratic granodiorite (Barksdale, 1968; Dixon, 1959).

14. Minor stocks in lower Methow region: mesocratic quartz diorite to granodiorite (Barksdale, 1968).

Table 2 cont.

DIKEROCKS BETWEEN MONUMENT PEAK AND GOLDEN HORN PLUTONS

15. Pink, quartz-K-feldspar porphyry (Tabor, et al., 1968).

16. Gray, plagioclase porphyry (also common in southern Methow Mountains near Cooper Mountain pluton) (Barksdale, 1968; Tabor, et al., 1968).

The most important sources of crystalline erratics in the Methow drift are the Pasayten Peak quartz diorite and the Golden Horn and Monument Peak granites. A dike swarm between the Golden Horn and Monument Peak plutons (Fig. 2; Tabor, et al., 1968, Fig. 4) supply two types of porphyry that, because the dikerocks are very hard and easily quarried from the fissile shale and friable sandstone host rocks, are important drift constituents for more than 50 miles southeast.

TERTIARY VOLCANIC ROCKS

The Island Mountain Volcanics (Staatz, et al., 1971, p. 38), gently folded volcanic rocks of intermediate composition, unconformably overlie Newby Group andesites as well as older crystalline rocks east of the Chewack-Pasayten Fault. Because the volcanic rocks lie athwart the Chewack-Pasayten Fault, they postdate its latest movement; because they were apparently emplaced nonconformably across a low-relief surface, the erosional vacuity beneath them is large.

The Sinlahekin Volcanics (Goldsmith, 1952, p. 337), a field of dacite and andesite that caps part of the Okanogan Highlands has a near-horizontal basal nonconformity indicating emplacement over a low-relief erosion surface. Their intermediate volcanic lithologies, their gentle attitudes, and their nonconformable relation to older rocks suggests their relation to the Island Mountain Volcanics. Sediments containing an "early Tertiary" flora (Waters and Krauskof, 1941, p. 1372) conformably underlie similar rocks in the Okanogan Valley near Tonasket. No outliers of Columbia Basalt have been found in the Methow Drainage Basin despite several basaltic dikes that crop out in the lower half mile of the Methow Valley.

Physiographic and Structural Relations

REVIEW OF TERTIARY RECORD OF THE CASCADES

Owing to greater Plio-Pleistocene uplift in the northern Cascades than farther south, the Tertiary stratigraphic record is mostly missing in the Methow region. Geographic conditions under which the drainage basin evolved, therefore, can be inferred only by reference to the more complete stratigraphic relations further south and from landforms of the Methow region.

Figure 3 is a compilation of stratigraphic and intrusive relations in the Cascade Range in Washington and Oregon and in adjacent lowlands. Unconformity-bounded packages of continental volcanic and clastic sedimentary rocks are underlain by more intensely deformed rocks and overlain by less deformed and less altered rocks. Each episode of widespread volcanism (Eocene, Oligo-Miocene, Late Miocene-Pliocene) probably led to obliteration of

Figure 3. Stratigraphic relations in the Cascade Range and adjoining regions. Line pattern denotes mountain-building episodes when drainage lines might have been tectonically affected; V-pattern denotes episodes of widespread volcanism during which drainage lines might have been deranged. Vertical scale is relative only.

AGE	ROCK TYPES	EXAMPLES	AUTHORITIES
	MODERN HIATUS		Mackin & Cary, 1965; others cited herein
QUATERNARY	Andesite, Glacial deposits	Glacier Peak, Mt. Baker, Mt. Rainier Andesite	Crandell, 1965; Mackin & Cary, 1965; Waters, 1955
	CASCADIAN HIATUS N-S folding, plutons		Mackin & Cary, 1965; Waters, 1939; Wheeler & Mallory, 1970
PLIOCENE	Volcanic-derived sediments	Howson Andesite, Ellensburg Fm.	Foster, 1960; Mackin, 1961; Waters, 1955; Wheeler & Mallory, 1970
? Upper MIOCENE	COLUMBIA RIVER BASALT	Yakima Fm.	
	OCHOAN HIATUS NW-SE folding, plutons	Tatoosh, Snoqualmie, Chilliwack III, Cascade Pass plutons	Mackin & Cary, '65; Misch, '66; Wheeler & Mallory, '70
Lower MIOCENE OLIGOCENE	ANDESITE, DACITE, Basalt, Rhyolite, and volcanic-derived sediments	Keechelus, Hannegan, Skagit, Island Mtn. (?) Sinlahekin (?) Volcanics, John Day Fm.	Foster, '60; Misch, '66; Snavely & Wagner '63; Wheeler & Mallory, '70 H. E. Wheeler, '70, oral comm.
? Upper EOCENE	Conglomerate	rarely exposed	
	HIATUS plutons	Chilliwack I, Monument Peak, Golden Horn plutons	Misch, 1966; Staatz, 1971; Tabor, 1968
Upper EOCENE	Rhyolite to Basalt	rarely preserved	Foster, 1960; Mackin & Cary, 1965; Misch, 1966; Snavely & Wagner, 1963; H.E.Wheeler, 1972, oral comm.
EOCENE	Clastic sedimentary rocks, Coal	Roslyn, Huntingdon, Tyee Fms.	
	BASALT	Teanaway, Naches Fms.	
	LARAMIDE HIATUS		Misch, 1966; H. E. Wheeler, 1972, oral comm.
PALEOCENE	Clastic sedimentary rock	Swauk, Chuckanut, Pipestone Canyon Fms.	Barksdale, 1948, 1960; Foster, 1960; Misch, 1966; Royce, 1959 Ryason, 1954; Willis, 1953
	HIATUS plutons	Black Peak, Pasayten, Rock Creek, Castle Peak plutons	Misch, 1966; Staatz, 1971; Tabor, 1968
Paleocene ? CRETACEOUS	CLASTIC SEDIMENTARY, ANDESITE	Midnight Peak, Winthrop, Virginian Ridge Fms.	Barksdale, '48, '60; Coates, '70 Pierson, '72

SUMMARY OF TERTIARY STRATIGRAPHY IN THE OREGON AND WASHINGTON CASCADES earlier drainage lines and establishment of new consequent streams; each unconformity delineates a mountain building episode that probably modified drainage patterns.

TERTIARY DRAINAGE OF THE METHOW REGION

The Methow River, midway between facing fault-related scarps, is in an ideal position to have originated as a consequent stream along the floor of the graben or along the axis of the attendant southeast-plunging syncline (see Fig. 2). The Methow-Pasayten area must have been a lowland due directly to faulting (a graben) during the accumulation of the six-mile-thick late Mesozoic rock sequence. Movement ceased at least by mid-late Cretaceous time along the western (Twisp-Hozameen) fault, however, when an 84 m.y. pluton was emplaced across the fault (Coates, 1970, p. 154); movement on the eastern (Chewack-Pasayten) fault ceased before Oligo-Miocene time(?), when the Island Mountain volcanics were extruded across a low-relief erosion surface beveled across the fault trace.

Whatever drainage lines were established in response to early Tertiary tectonic settings, the Oligo-Miocene extrusion of volumious lava flows across low-relief surfaces (Fig. 3) bespeaks obliteration of pre-Oligocene drainage and establishment of a new consequent drainage. Therefore, the Methow River probably did not inherit its modern course from a graben-floor consequent stream that was established in late Cretaceous or early Tertiary time.

Because the Skagit River heads in the eastern half of the Cascades and flows west directly across the structural and general topographic crest of the range, it must antedate the Plio-Pleistocene Cascade uplift. The northwest trends of Skagit tributaries and of such eastern Cascade master streams as the Methow are distinctly anomalous to the north-south trend of the Cascade Range; they also must predate the Plio-Pleistocene uplift (Mackin and Cary, 1965, p. 13).

In late Miocene time the Columbia Basalt was spread across the beveled-off folds developed in the Oligocene volcanic rocks. In the northeast Cascade region, where relief on the sub-basalt unconformity is at least 2000 ft (Waters, 1939), and where the Columbia River apparently was forced to the margin of the spreading basalt field (Willis, 1887; Flint, 1935; Waters, 1955; Mackin, 1961; Mackin and Cary, 1965, Fig. 6), the basalt must have lapped northwest against a maturely dissected upland. That the basalt apparently drowned the lower ends of eastern Cascade rivers like the Methow indicates that they were established <u>prior</u> to late Miocene time.

Since most structural axes in Oligo-Miocene and older rocks trend northwest (Wheeler and Mallory, 1970, p. 117); the Miocene Ochoan orogeny must have produced fold mountains trending northwest. The Miocene drainage pattern that developed, whether consequent on the folds or subsequent on weak-rock belts exposed by erosion of the folds, would have been parallel to the strong northwest structural and topographic grain. Thus there is a ready explanation for the apparently anomalous northwest trend of the eastern Cascade master valleys (Mackin and Cary, 1965, p. 13). That such valleys as Entiat, Chelan, tributaries of the Chewack, and even the lower Methow, are excavated entirely in crystalline rocks probably resulted from superposition through the cover of folded sedimentary and volcanic rocks, a cover that has been largely stripped away by by Plio-Pleistocene erosion.

RELATION OF METHOW LOWLAND TO ITS FAULT-RELATED SCARPS

The scarp associated with the Chewack-Pasayten Fault is resequent where erosionally weak Newby shales in the lower Methow region are juxtaposed with the crystalline rocks. But in upper Eightmile Creek where erosionally resistant Newby Group andesite is against the same gneissic rocks, the scarp is obsequent. The Chewack-Pasayten scarp, therefore, is the direct result of erosion--a fault-line scarp, not a fault scarp. Similarly, along the southeastern 20 miles of the Twisp Valley Fault where weak Newby shales are in fault contact with gneiss, the scarp is pronounced and resequent; farther northwest where resistant Midnight Peak andesite is against the same crystalline rocks, the scarp is obscure. The scarp is a fault-line scarp and thus a firm case can be argued for an erosional origin of the

Methow lowland: it is a rift-block valley in the sense of Johnson (1929), not a graben as originally defined by Suess.

STREAM EVOLUTION WITHIN THE METHOW DRAINAGE BASIN

The master Methow stream having been established in its southeast course generally parallel to rock structure and along the axis of the sedimentary block, a necessary consequence is the tendency of its tributaries to head in one of the crystalline-rock blocks and to descend to the Methow River generally across structural trends. Nonetheless, many Methow tributaries pursue courses parallel to rock strike. The Eightmile and upper Twisp Rivers faithfully follow the traces of the two boundary faults; tributaries of the Methow West Fork, of Goat Creek, of the lower Chewack River, and of Libby Creek comprise trellis patterns that are locally striking (Fig. 1). Apparently capture of structure-transverse streams by subsequent streams in belts of weak rock has been important to the evolution of drainage lines <u>within</u> the sedimentary block. The Eightmile and upper Twisp rivers, which probably originated as subsequent streams along erosionally weak fault-gouges, must have developed headward by successive capture of small Methow tributaries that formerly maintained courses across wide belts of resistant andesite. One of the "captors", the lower Twisp River, however, crosses the widest part of the Midnight Peak andesite belt; furthermore, the asymmetry of Falls Creek Valley indicates that if Eightmile

Creek beheaded smaller streams they were Falls Creek tributaries draining eastward from Isabella Ridge, not Methow tributaries draining westward across the Isabella andesite belt. Although capture probably isolated the erosionally resistant plutons and andesite belts as the highest peaks within the Methow rift-block valley, these drainage adjustments took place long ago. Nonetheless, they must have occurred after Oligocene (?) time, by when erosion and volcanism had destroyed the earlier Tertiary graben.

INCISION OF THE METHOW DRAINAGE BASIN

That the Methow valley system incises a widespread low-relief erosion surface sloping southward from 7000 ft to 4000 ft in the Okanogan Highlands indicates at least one major pause during downcutting. The Okanogan surface and others like it farther south, each conspicuous because of its contrast to the generally mountainous topography of the eastern Cascades, is the so-called Entiat erosion surface (Willis, 1903). Waters (1939) inferred that it was a Miocene surface formerly buried by, and recently exhumed from, a carapace of Columbia Basalt. However, because the Okanogan surface extends more than 60 miles north of the Columbia River, which presumably originated as a consequent stream at the periphery of the Columbia Basalt field (Mackin and Cary, 1965, p. 19), its origin as a resurrected surface is unlikely. An alternative is that the Columbia Basalt, having dammed and flooded the lower ends of the Methow and other eastern Cascade rivers, thus raised their local base levels by at least 2000 ft, initiating a latest Miocene interlude of aggradation or reduced downcutting. As the Columbia incised the basalt in Plio-Pleistocene time, the Methow concurrently incised the graded Okanogan surface. The 45 ft/mi southeast gradient of the surface is probably partly the original slope of the late Miocene-early Pliocene pediplain, and partly the result of southeast tilting during the Plio-Pleistocene Cascade uplift.

EFFECTS OF GLACIATION

As summarized in the remainder of this report, most of the Methow region was affected by glaciation. Alpine glaciers excavated cirques and modified former stream valleys into deep U-shaped troughs; the Cordilleran Ice Sheet beveled off ridge crests, excavated anomalous U-shaped troughs across divides, and created a variety of depositional landforms along the trunk valleys; ice-marginal streams associated with both types of glaciers cut thousands of channels trending anomalously across divides and spurs. The southern Methow region, though mostly overriden by the ice sheet, bears little topographic evidence of it: the valleys are V-shaped and the intervening ridges are sharply rounded. The limits of alpine glaciers on Figure 8 approximately delineate the northern limits of stream-erosional topography in the Methow region. Because the Late Wisconsin Cordilleran Ice Sheet entered the region mainly from the northwest, it probably deepened those valleys aligned parallel to its flow. Unlike the alpine glaciers, the ice sheet selectively deepened valleys aligned northwest-southeast, thereby enhancing the topographic grain that was originally etched out by stream erosion.

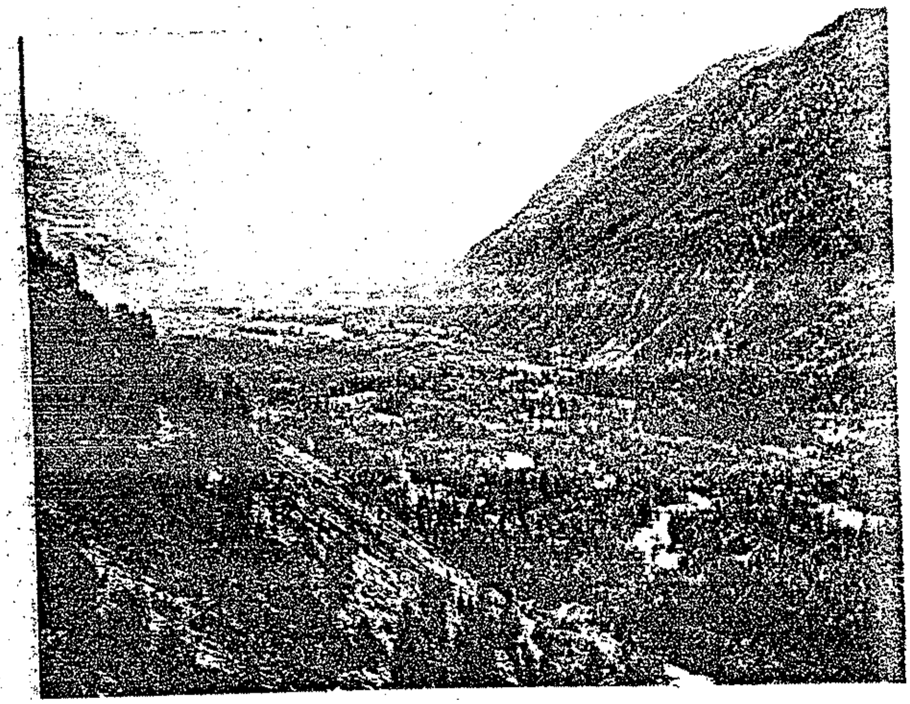
PLEISTOCENE GLACIATION

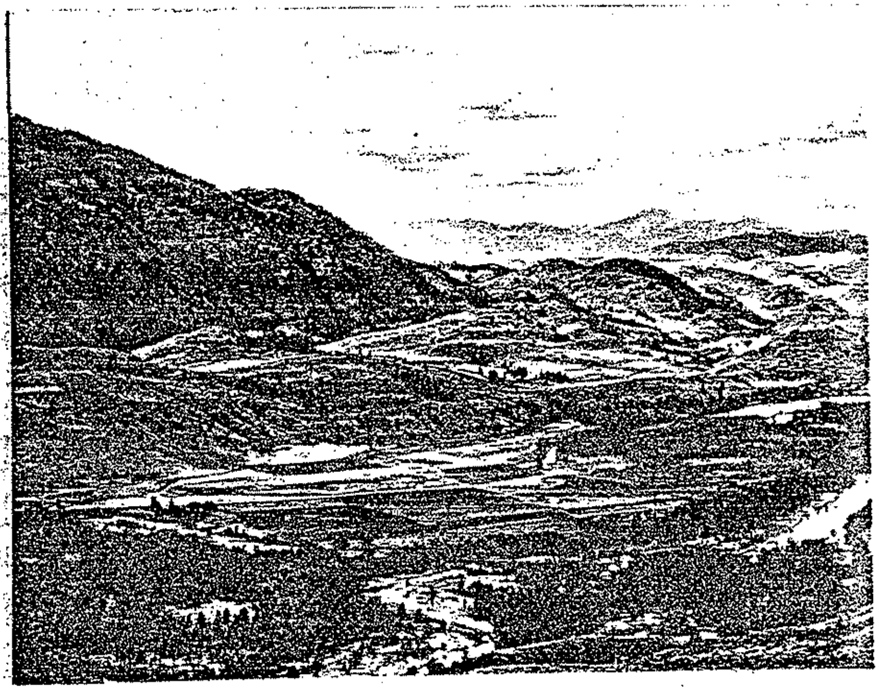
Evidence of Glaciation

EROSIONAL LANDFORMS

Shapes of Valleys. Valleys in the northern parts of the Methow Drainage Basin head in cirques that descend into wider and deeper trunk valleys having graceful, parabolic cross profiles. Depending on the relative sizes of the confluent troughs, the smaller trough typically "hangs" a few hundred to 2000 ft above the floor of the larger trough. The trunk Methow, Chewack, and upper Twisp valleys are spectacular steep-sided U-shaped troughs excavated 1000 to 3000 ft below the glaciated uplands (Fig. 4). Neighboring glaciated valleys generally are separated by arete-like ridges. These and associated landforms, characteristic of alpine landscapes everywhere, constitute the chief evidence of a formerly extensive alpine style of glaciation in the northern Cascades.

Rounded cols. Many cols are not the ragged low aretes of the type normally produced by plucking processes of competing cirque glaciers, but are gently rounded saddles that obviously have been modified by ice that flowed across them from one valley head into the other. These and similar U-shaped troughs extending anomalously across the lower ridgecrests are striking evidence that the most recent glaciation of the northern Cascades was not a typical alpine glaciation confined to valleys, but was an ice





sheet that overtopped ridges and peaks as high as 5000 ft above the trunk valley floors.

Ice-marginal channels. In the southern Methow region most valley sides are interrupted by narrow V-shaped or flat-floored hanging channels that slope downvalley generally parallel to the modern valley floor (Fig. 22). Similar channels that descend generally southeast across many of the rounded passes are anomalous to any strictly stream-erosional landscape but are of similar morphology and position to typical ice-marginal stream channels (Henderson, 1959, p. 40-46; Penttila, 1963, p. 29-39; Rich, 1908; Tarr, 1908). These channels provide minimum upper limits of valley-ice tongues and indicate the direction and, in some places, the ice-surface gradient during deglaciation.

Drainage derangement. Ptarmigan and Johnny Creeks flow north-northeast into Hidden Lake's trough wherein they turn abruptly southeast, forming the master stream, Lost River, that drains south to the upper Methow River (Fig. 1). The barbed relation to Hidden Lakes trough is evidence of the capture of Ptarmigan and Johnny Creeks by the headwaters of Lost River. Insofar as the Lost River has no apparent erosional advantage over the Pasayten River the capture is inexplicable by stream-erosional processes. The northwest orientation of Hidden Lake s trough is, however, ideally located to have been a zone of intensive scour by an ice sheet that moved southeastward, up the former Pasayten tributary. Moreover, a high-altitude area in the Monument Peak-Tatoosh Buttes area would have necessitated bifurcation of such an ice sheet, directing the eastern ice stream up the Pasayten East Fork and Hidden Lakes trough. The favorable northwest orientation of the trough and the funneling of an ice stream through it caused subglacial erosion to the degree that the preglacial divide was shifted five or more miles northwest.

ABRASION OF ROCK SURFACES BY ICE

Smoothly abraded or polished rock surfaces that are striated, grooved, or broadly fluted indicate directions parallel to which ice moved. Where associated with stoss-and-lee topography or with erratics of known provenance, striations and grooves provide a means by which a unique direction ice flow may be determined (Fig. 15). In the Methow region striations are best preserved on andesite, dike rocks, arkose, and other hard, fine-grained rocks. Although along the Cascade crest striations are locally preserved, on plutonic and gneissic rocks and on fissile shales and friable sandstones they are generally sparse. Striations are not only etched into such rocks reluctantly, but they are lost because of weathering, especially in the more arid southeastern areas where postglacial granular disintegration is most advanced.

DEPOSITIONAL LANDFORMS

Although the northern parts of the trunk valleys contain a few small lateral moraines, most depositional landforms in the Methow region are kame terraces, kame-and-kettle topography, and eskers. Small terraces and deltas record ice-marginal streams and lakes as high as 2000 ft above the modern valley floors.

Glacial sediments of the Methow region comprise till, ice-contact stratified drift, glacial lacustrine silts, and erratics. Glacial stones are of several types: a) erratics whose source lies outside the immediate area; b) locally derived erratics that could not have reached their present positions by slope processes; c) stones that, though lying on bedrock of similar lithology, are precuriously perched on ridgecrests or are otherwise positioned so as to preclude an origin by mass-wasting processes (Fig. 19).

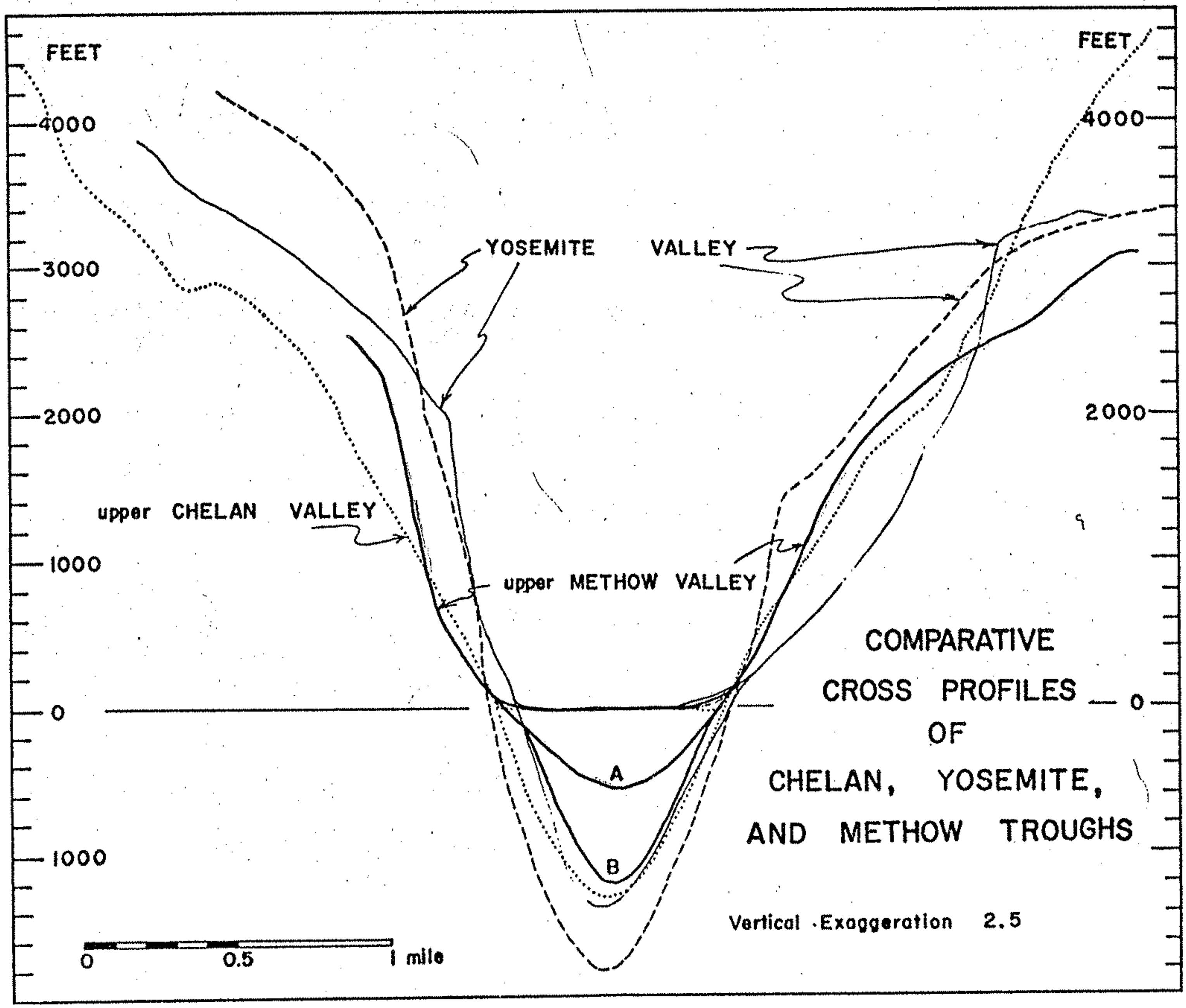
At altitudes above 5000 ft the boulders are generally associated with little or no till. A discontinuous layer of drift (mostly a till of pebbly sandy mud) a few inches to many ft thick generally mantles slopes and ridgecrests. In the Methow, Chewack, Twisp and Eightmile Valleys and in tributaries to the lower Methow, much of the surface drift is stratified sandy gravel, and is generally associated with ice-stagnation landforms. Locally, laminated silt and clay provide evidence of lakes that were temporarily dammed by ice during deglaciation.

Infilling of Trunk Valleys

Unlike the smoothly parabolic profiles of Early Winters Valley and most other tributary troughs, the floor of the Methow trough is a mile wide and is flat (Fig. 4). The floor of Yosemite Valley, whose size and valley-side profiles resemble those of the upper Methow, also is flat-floored because of sediment more than 1200 ft and 1800 ft thick, respectively, in two U-shaped bedrock basins delineated by seismic surveys (Gutenberg, et al., 1956, Fig. 3). According to Whetten's (1967) sparker survey the submerged bedrock floor of upper Chelan trough is parabolic; however, where Stehekin River sediment has filled in the head of the trough to a depth of both 1200 ft the valley floor is flat. Yosemite and Chelan Valleys are flat only because they are deeply filled. Figure 5 compares the valley-side profiles of the Yosemite and Chelan troughs with the upper Methow trough, showing as well the topographic similarity of the flat floors of all three valleys.

Insofar as seismic surveys have not been conducted in the Methow Valley, nor water wells drilled deeper than 90 ft, direct measurement of the thickness of drift is lacking. However, the Methow River during its low-water stage in autumn, commonly runs dry below the Lost River confluence and surfaces 9 mi downvalley, indicating a buried bedrock basin whose deepest part is near the Early Winters confluence. By projection of the Methow valley-side

Figure 5. Comparative profiles of the upper Methow, upper Chelan, and Yosemite glacial troughs. Datum is altitude of flat valley floor. Projection of Methow valley-sides smoothly (A) and approximately parallel (B) to Yosemite and Chelan contours, indicates that fill in Methow Valley is 600-1200 feet deep.


FEET
4000
3000
2000
1000
0
1000
FEET
4000
2000
0
YOSEMITE VALLEY
upper CHELAN VALLEY
upper METHOW VALLEY
A
B
0
0.5
1 mile
COMPARATIVE
CROSS PROFILES
OF
CHELAN, YOSEMITE,
AND METHOW TROUGHS
Vertical Exaggeration 2.5

contours through an approximate parabola similar to the measured curves of Yosemite and Chelan troughs (Fig. 5), the estimated depth to which the upper Methow trough has been infilled is 500 to 1200 ft. Although the flat floors of upper Twisp, Chewack, and Eightmile Valleys also indicate infilling by late-glacial and postglacial sediment, the depth of fill probably is less than in the wider Methow trough.

Correlation of Methow Glacial Record

AGE OF METHOW GLACIATION

Approximately 98 percent of the stones in Methow drift or stranded on bedrock surfaces are unweathered. Weathering rinds on fine-grained rocks are 1 to 2 mm thick; discoloration in coarse-grained crystalline rocks does not exceed 1 cm. When tapped with a hammer, crystalline rocks are almost invariably solid. The depth of weathering on polished bedrock surfaces is no more than a few millimeters. Soils, moreover, are invariably immature: the C-horizon of the soil ranges from 1 to 4 ft deep in compact till, and from 3 to 15 ft in gravel. The freshness of Methow drift indicates that the glaciation correlates broadly with the Fraser Glaciation of the Puget Lowland and British Columbia.

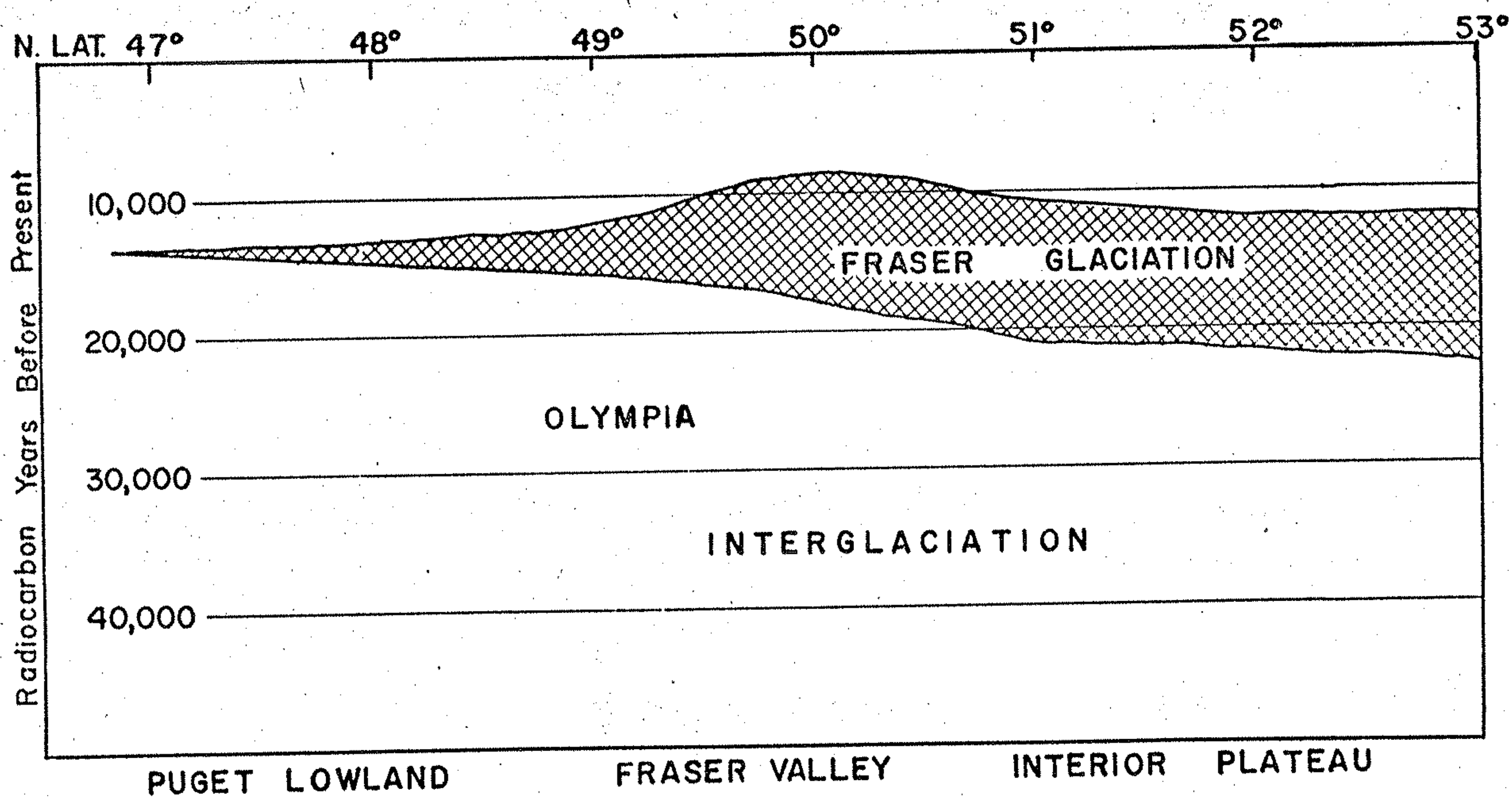
In the Puget Lowland and in the Interior Plateau of British Columbia, the period of the latest glaciation is well bracketed by radiocarbon dates (Armstrong, et al., 1965, Fig. 2, p. 347; Fulton, 1968, 1971; Mathewes, et al., 1972). The ice sheet arrived at latitude 51° in British Columbia about 20,000 years B. P., reached latitude 47°30' (Seattle) about 15,000 and reached its maximum position at latitude 47° about 14,000 years ago. By 13,500 B. P., the ice had retreated far enough north (48°30') to allow reentry of sea water into Puget Sound, and by about 11,450 B. P. it had retreated well into interior British Columbia (Fig. 6).

Because the Okanogan Lake and Puget Lobe issued from the same ice reservoir in interior British Columbia, these radiocarbon dates are probably broadly applicable to the northern Cascade region. Thus, the ice sheet probably arrived in the Methow region by 16,000 years B. P. and had largely diminished by 13,500 B. P. According to an estimate by Armstrong, et al. (1965, Fig. 2), alpine glaciers from the Cascades reached their maximum positions around 18,000-19,000 B. P., well before the culminating advance of the Puget Lobe in western Washington.

NUMBER OF GLACIATIONS IN THE METHOW REGION

In only three places has highly weathered drift been found beneath Late Wisconsin (Fraser) drift: a) in roadcuts along the north side of Beaver Creek road compact till overlies less-indurated, highly weathered sandy gravel; b) in roadcuts along Forest Route 335 on the east wall of Finlay Coulee till with highly weathered quartz diorite clasts underlies fresh stratified drift; and c) a quarter mi

Figure 6. Time-distance diagram of latest advance of Cordilleran Ice Sheet, based on radiocarbon dates cited by Armstrong, et al., 1965; Fulton, 1968, 1971; and Mathewes, et al., 1972. The younger age of the final deglaciation between N. lat. 49° and 51°, the so-called Sumas Stude, is apparently confined to the vicinity of the Coast Ranges. Many radiocarbon dates (Fulton, 1971, Fig. 2; Mathewes, et al., 1972) indicate that the Interior Plateau was deglaciated by 11,400 years B. P.


N. LAT. 47°
48°
49°
50°
51°
52°
53°
Radiocarbon Years Before Present
10,000
20,000
30,000
40,000
FRASER GLACIATION
OLYMPIA
INTERGLACIATION
PUGET LOWLAND
FRASER VALLEY
INTERIOR PLATEAU

northwest of Buck Lake highly weathered gravel, having crystalline clasts that are weathered to the core, apparently underlies late Wisconsin till.

The Beaver Creek gravel very likely is weathered Pipestone Canyon Conglomerate; a small adjustment of Barksdale's (1968) and Ryason's (1954) contacts accommodates this exposure. Late Wisconsin glacio-fluvial drift overlies the weathered Finley Coulee till whose weathered state, therefore, may be due to late-glacial or postglacial groundwater percolation, rather than to age. The strongest candidate for pre-Late Wisconsin drift is the Buck Lake exposure. The Buck Lake gravel probably records a pre-Late Wisconsin diversion through Buck Coulee of water marginal to a glacier tongue in Eightmile Valley. It could relate to a Late Wisconsin alpine phase of glaciation preceding the arrival of the Late Wisconsin ice sheet; its weathered state suggests a relation to some earlier (Early- or Pre-Wisconsin) episode of glaciation. Whatever its age, the older drift indicates that Buck Coulee was close to its present depth before being overridden by the Late Wisconsin ice sheet, and that the shape of the Buck Lake trough was modified but little by the ice sheet. Except for the small exposure in the Buck Lake trough, no sound stratigraphic evidence has been found for a pre-Late Wisconsin glaciation.

MULTIPLE GLACIATION IN EASTERN WASHINGTON

Although Bretz's (1924) evidence for a pre-Late Wisconsin glaciation, the so-called "Spokane Glaciation", was refuted by Flint (1937), Flint himself advanced topographic arguments for multiple glaciation in eastern Washington (1935, p. 171 176). Since then, Roald Fryxell (see Richmond, et al., 1965, p. 238 and Table 1) has defined, on the basis of soil stratigraphy, at least six episodes of loess deposition ("Pinedale"; "Bull Lake" [3]; "pre-Bull Lake" [2]) probably ranging back into Illinoian or earlier time. Therefore, eastern Washington probably was affected by ice sheets, and the eastern Cascades probably by alpine glaciers, on several occasions before the obvious Late Wisconsin ice advances.

North of the Cascades two till sheets separated by interglacial sediments of probable Sangamon age have been described by Fulton (1968, p. 1077). The stratigraphic record of the Puget Lowland indicates that at least three glaciations preceded the Late Wisconsin (Fraser) event (Crandell, 1965, Table 1). Deductively, therefore, the northern Cascades also probably experienced multiple glaciation, even though firm stratigraphic evidence for it is poor. Unless ice-sheet fluctuations west of the Cascades were very different than in eastern Washington, the erosional effects of glaciation in the Methow region are the cumulative results of several glaciations.

NOMENCLATURE

In spite of the lack of radiocarbon dates bearing on the age of ice-sheet glaciation in eastern Washington, Idaho, and Montana, Richmond, et al. (1965) correlated moraines of the Cordilleran Ice Sheet with those of alpine glaciers in the northern Rockies. To complicate matters, he assumes that the maximum stands of alpine glaciers in the northern Rockies correlate with those in the type area of Rocky Mountain glaciation (Wind River Mountains). Hence, Richmond et al. (1965) refer to Early, Middle, and Late Pinedale advances of the ice sheet, the implication being that ice-sheet advances correlate exactly with advances of the Wind River alpine glaciers.

Because alpine glaciers in the western Cascades reached maximum positions prior to the maximum Puget Lobe advance (Cary and Carlston, 1937; Mackin, 1941; Vance, 1957, p. 292-299), the need arises to distinguish between advances of alpine glaciers and of the Cordilleran Ice Sheet. Richmond's use of Rocky Mountain alpine terminology is inappropriate for the Cordilleran Ice Sheet in Washington; his usage is not followed here.

Puget Lowland nomenclature, however, is based on fluctuations of the southern part of the Cordilleran Ice Sheet. Since the same ice mass affected the Methow region as interior British Columbia, Fulton's (1968) precedent of using Puget Lowland terminology is followed here. Because maximum advance of the Puget Lobe of the Cordilleran Ice Sheet and the midwestern sector of the Laurentide Ice Sheet were broadly contemporaneous (Armstrong, et al., 1965; Frye, et al., 1965, Fig. 5), the more widely used term "Late Wisconsin" is also loosely applied to the latest ice-sheet glaciation of the Methow region.

Late Wisconsin alpine glaciation that originated in cirques prior to ice-sheet glaciation, is referred to simply as the "Alpine Phase". Whether this Alpine Phase correlates directly with the Evans Creek Stade of the western Cascades or to alpine advances recorded in other valleys in the eastern Cascades, is not known.

Alpine Phase of Late Wisconsin Glaciation

GENERAL STATEMENT

Glaciated regions of the middle Cascades have looped moraines on valley floors and a distinct upper limit of glacial drift on the valley sides. In the Methow valley, however, valley-floor moraines are flat-topped, irregularly distributed, and composed of stratified drift. Moreover, a single drift sheet extends from valley floors to the summits of all but the highest peaks, indicating that, unlike valleys of the middle and southern Cascades, the most recent glaciation of the Methow region was of the ice-sheet type. That U-shaped valleys heading in steep-walled cirques are characteristic upland landforms indicates, however, that alpine glaciation, not ice-sheet glaciation, was the primary sculptor of the landscape. Because all drift and associated depositional landforms associated with the alpine glaciation are missing, apparently having been eroded away by the younger ice sheet, an analysis of the alpine glaciation is necessarily restricted to erosional landforms.

Erosional landforms in the Methow region fall into four categories: a) those produced or modified by erosion at the base of alpine glaciers, b) those produced by streams marginal to valley glaciers, c) those produced by normal stream erosion and which were but slightly modified during glaciation, and d) those strongly modified by ice-sheet erosion.

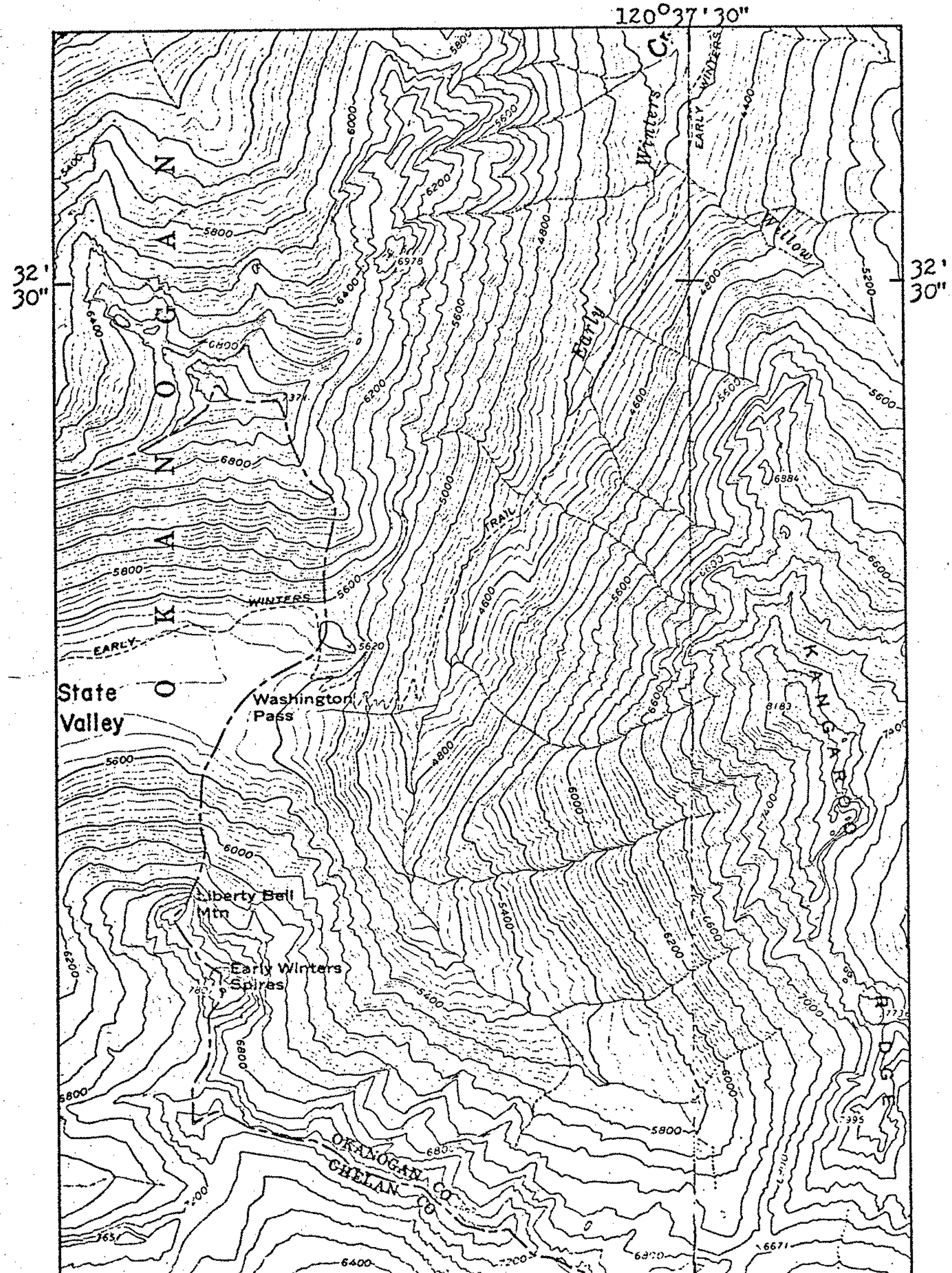
ALPINE GLACIAL-EROSIONAL LANDFORMS

Cirque-floor altitudes and relation to altitude of modern glaciers. Four small glaciers on the ragged north-facing slopes of Tower and Golden Horn mountains have a mean altitude of 7010 ft, as compared to a mean altitude of 6200 ft (range: 5900-6400) for the floors of five small north-facing cirques in the same area. Thus, a lowering of the modern snowline by 850 ft probably would generate glaciers in these cirques. According to the average atmospheric lapse rate of 1°F per 300 ft altitude, a mean temperature of 3°F (1.7°C) less than the present might, with the modern annual snowfall, restore alpine glaciers to the Methow region.

Thirty-five mi east, in the vicinity of Tiffany Mountain, the mean altitude of four north-facing cirques is about 7060 ft (range: 6850-7300), or 860 ft higher than the mean for the Tower Mountain area. Near Raven Ridge, 30 mi south-southwest of Tower Mountain, the mean altitude of six well-formed cirques is 7185 ft. The eastward rise from Tower Mountain by 25 ft/mi and the southeast rise by 34 ft/mi delineate the effects of climate on Late Pleistocene snowline. Porter (1964, Fig. 3) reports an eastward rise across the Olympic Mountains and Cascade Range both of Pleistocene cirques and of modern glaciers, owing presumably to orographic effects on moisture-bearing Pacific air-masses.

Washington Pass area. Early Winters Valley, its graceful U-shaped cross-profile, its hanging tributary troughs, and the arete-like ridges surrounding it, are to the casual observer a typical example of alpine landscape in the northern Methow region. Washington Pass, the "hanging" intersection of the U-shaped State Valley with the west side of the Early Winters trough is, however, patently anomalous. State Valley, a Chelan tributary, descends from Early Winters Valley; it is essentially headless (Fig. 7). Topographic peculiarities are equally abundant in Early Winters Valley proper. Contrasting the generally smooth cross-profiles of the lower valley, vertical rock buttresses near Washington Pass form a constriction through

Figure 7. Topographic map of Washington Pass area, from parts of Washington Pass and Silver Star Mtn. 7.5 topographic sheets.



WASHINGTON PASS AREA C. I. = 40 ft. 120°37'30" which the valley floor ascends steeply to its headward cirque.

The cirque faces northwest, oddly misaligned with north-northeast-trending Early Winters Valley. Across Early Winters Valley and directly aligned with the cirque, however, is the "hanging" intersection, Washington Pass. The curvature of the cirque, no doubt developed during alpine glaciation because of more rapid retreat of north-facing headwalls than those of other azimuths, is also wrong for Early Winters Valley. The trend of Early Winters Valley is concave eastward; the cirque is concave southwestward, identical to the trend of State Valley. The curvature of the cirque walls, projected in a gentle arc westward across Early Winters Valley, in fact, coincide exactly to the arcuate walls of State Valley. The relationship of the cirque walls to Early Winters Valley is, on the other hand, cuspate. Furthermore, the cirque floor is distinctly higher than the floor of Early Winters Valley. Projected westward across Early Winters Valley, however, it roughly coincides with the anomalous end of State Valley. The cirque is clearly the missing head of State Valley; Early Winters Valley has somehow intercepted State Valley, abstracting the cirque.

Insofar as both State Valley and Early Winters Valley are excavated in rocks of equal erosional resistance, and since both valleys traverse similar distances to the

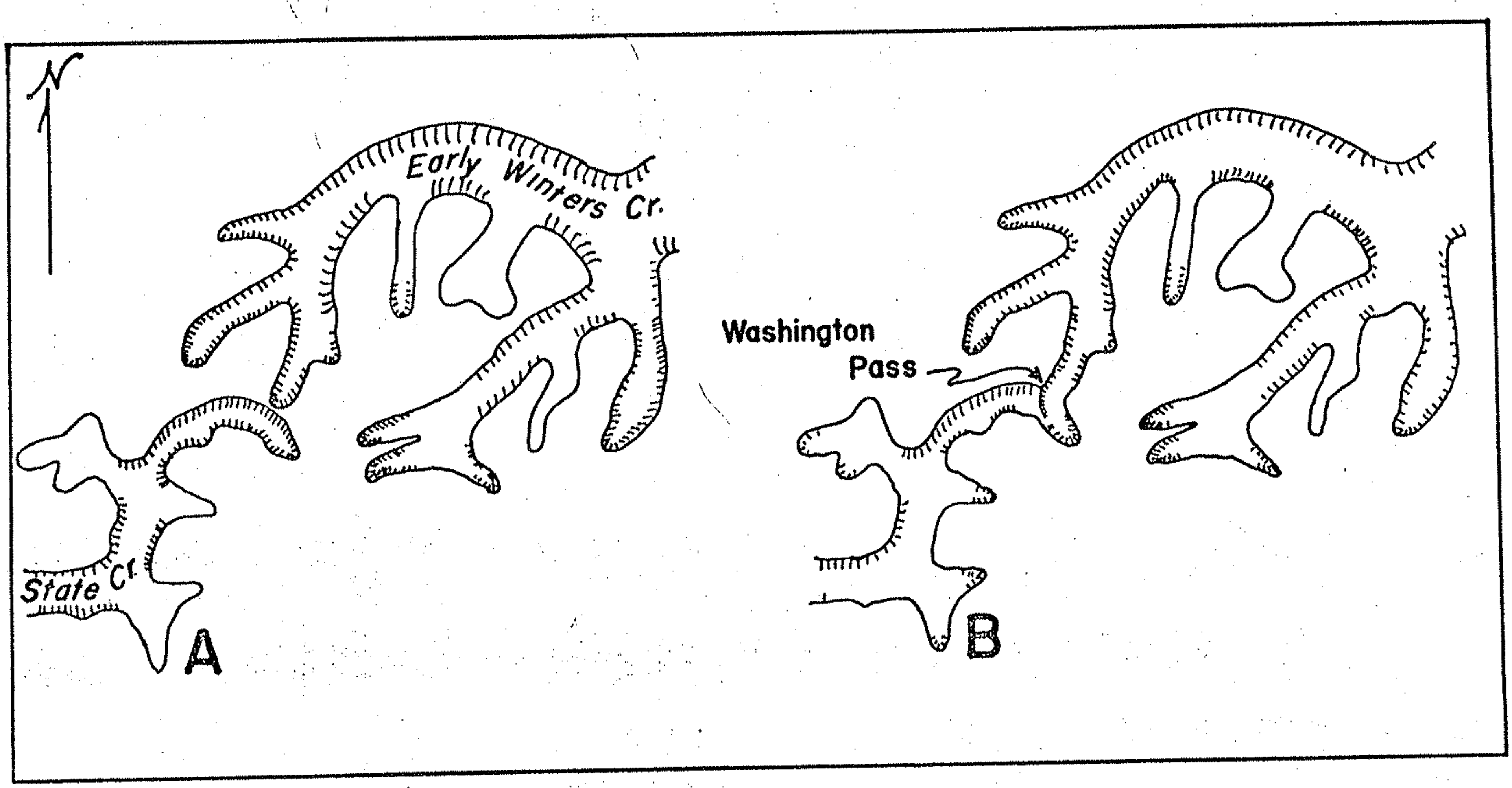
Columbia River baselevel, an explanation of the capture by stream-erosional processes is untenable. The capture must have occurred because of unequal competition during alpine glaciation.

Figure 8 depicts the capture of the head of State Valley by a tributary of the Early Winters Glacier. The Early Winters cirque not only faced north, the orientation most favorable to headwall retreat, but it competed with a <u>side</u> of the State Valley glacier. Thus, during alpine glaciation, Early Winters Valley held a decided erosional advantage over State Valley. Prior to the capture Early Winters Valley (8A), nourished by many large tributary glaciers, had become much deeper than the State glacial trough, nourished by only a few small tributary glaciers. Therefore, headward erosion by the Early Winters cirque resulted in undercutting of State Valley as well as incutting. And thus the Early Winters cirque beheaded its competitor (8B). That the "gorge of capture" is not modified to a graceful U-shaped profile, nor the original knickpoint obscured, nor the cirque walls more nearly aligned with Early Winters Valley, testifies to the recency of the capture: it is clearly an event of the Late-Wisconsin Alpine Phase.

Figure 8. Interpretation of cirque capture at Washington Pass.

A) Pattern prior to capture.

B) Pattern after capture, Washington Pass created.


Early Winters Cr.
Washington
Pass
State Cr.
A
B

<u>Reconstruction of extent of valley glaciers by erosional topography</u>. An estimate of the limits of the valley-glacier system responsible for the alpine topography is possible by delineating a) the downvalley limit of the U-shaped valley cross-profiles, b) the height of hanging glacial troughs and cirques, c) the upper limits of truncated spurs, and d) the lower limits of ragged topography associated with such inter-valley landforms as aretes, cols, and horns. Most alpine features have been somewhat modified by more recent ice-sheet erosion; but since the alpine-glacial sculpture is substantially intact in areas like Glacier Pass where ice-sheet modification was intense (p. 73) there is little problem in reconstructing the alpine landscape elsewhere.

<u>The Methow Mountains</u>. The Methow Mountains, which separate the Chelan and Methow drainage basins, illustrate the combined effects of altitude and longitude on the development of Late Wisconsin alpine-glacial morphology. The mountains, which culminate in several peaks over 7500 ft altitude, extend from Washington Pass 50 mi southeast to Goat Mountain. The northwestern 30 mi is a sharp arete; the southeastern 15 mi is a broadly rounded ridge 500 to 600 ft high.

Cirques along the high northwestern segment are well-formed, deep, and they open into glacial troughs that are U-shaped downvalley to their confluence with the Twisp

Valley trough. Between Star Peak and North Navarre Peak, though cirques facing northeast are well-developed, they feed downvalley into V-shaped stream-erosional valleys. Southeast of North Navarre Peak cirques are poorly developed or absent. A similar progression from well-developed cirques and glacial valleys in the northwest to normal stream-canyon heads in the southeast is apparent on the southwest side of the mountains in the heads of Chelan tributaries, whose lesser degree of cirque development is due no doubt to a greater insolation, and therefore a higher snowline, on those southwest slopes.

South of the great elbow of the Twisp Valley, the lower ends of Twisp and Methow tributaries are V-shaped stream valleys. The heads of Buttermilk Creek (West Fork and East Fork), Libby Creek, and Gold Creek, for example, are cirques; their mouths are V-shaped stream valleys. That these systematic variations survive in each valley despite ice-sheet glaciation even at the higher altitudes (Fig. 16) indicates that the topographic variations in the valleys were established by alpine glaciation. Furthermore, the tributaries and the topographic variations therein trend northeast whereas the linear effects of the ice sheet were southeast, normal to the tributaries (also Fig. 16). Therefore, the downvalley trends under discussion clearly relate to alpine glaciation rather than to ice-sheet glaciation.

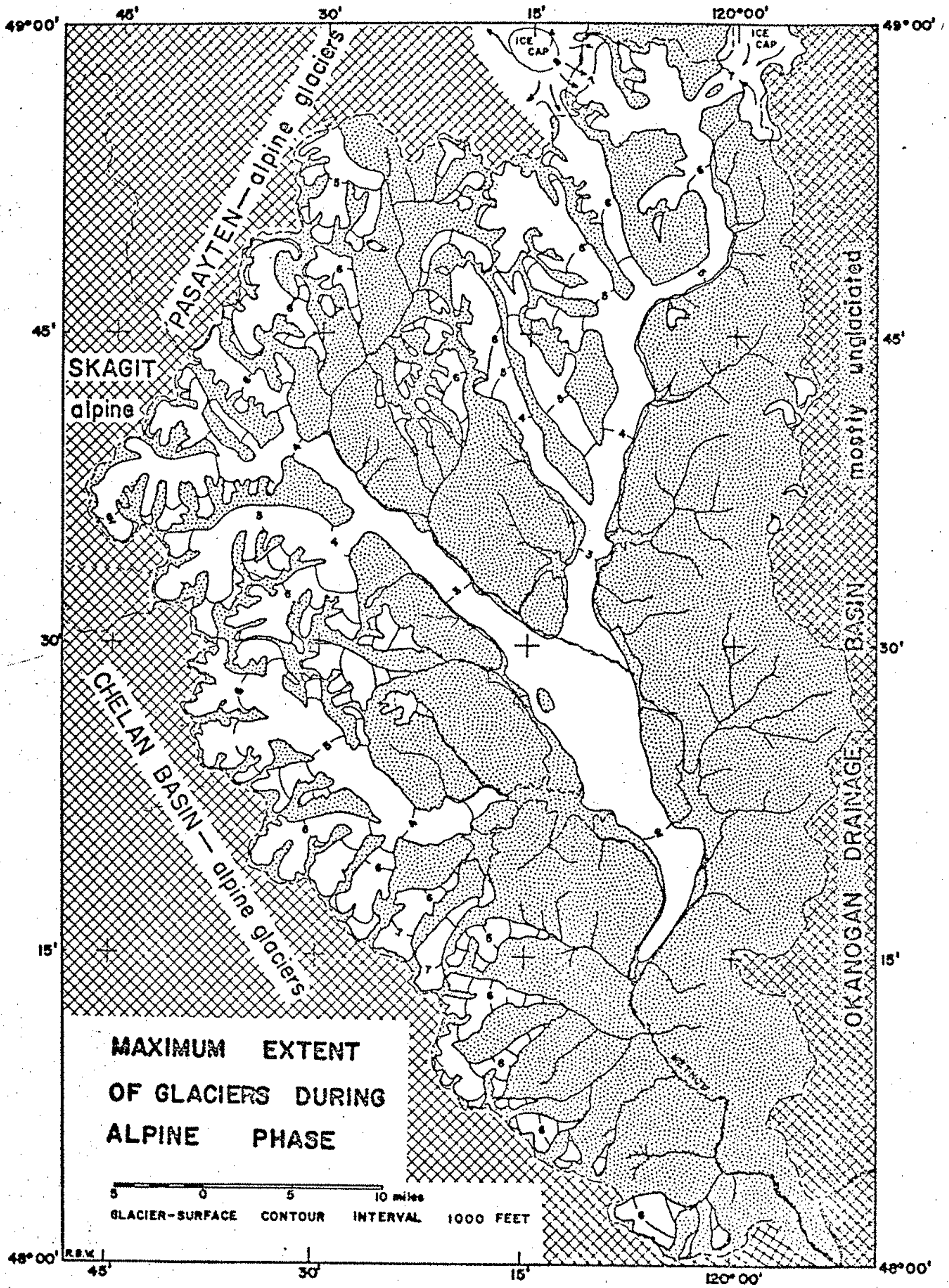
The downvalley limit of the distinctly U-shaped valley cross-profiles is taken as the minimum downvalley extent of Late Wisconsin alpine glaciers. Thus, the alpine glaciers of Buttermilk, Libby and Gold Creeks were 2 to 8 mi long and were not confluent with trunk glaciers in the Twisp and Methow Valleys.

Although the heads of Methow tributaries south of North Navarre Peak are rudely theater-shaped, their steeper downvalley gradients and gentler valley-side slopes contrast sharply with the well-developed fresh cirques farther north. But they contrast equally well with the steep-sided narrow V-shaped, southwest-facing heads of Chelan tributaries on the opposite side of the divide. The heads of the Chelan tributaries are probably of normal stream-erosional origin; the heads of Methow tributaries north of North Navarre Peak are typical cirques; the Methow cirques south of North Navarre Peak are of intermediate development.

Two hypotheses can be advanced for the origin of the poorly developed cirques south of North Navarre Peak. They may be pre-Late Wisconsin cirques that were not glaciated during the Late Wisconsin Alpine phase and therefore have been more strongly modified by slope processes. Alternatively, they may be Late Wisconsin basins that, because the mountain crest is 2000 ft lower than farther north even lower than the snowline inferred in the Raven Ridge area (p. 41) could not have been occupied by large glaciers. Because the only drift exposed in the region is of ice-sheet origin, depositional evidence for resolving the topographic puzzle is lacking. Figure 9 shows the estimated extent of glaciers in the heads of McFarland, Squaw, and Black Canyon Creeks according to the hypothesis that they were glaciated by Late Wisconsin ice.

Twisp Valley Glacier. The downvalley limits of U-shaped cross-valley profiles, truncated spurs, and hanging tributaries approximately delineate the former extent of the trunk valley glaciers. Upvalley of the great elbow of the Twisp Valley all tributary valleys have U-shaped cross-profiles at their junction with the trunk valley. Buttermilk Creek, Little Bridge Creek, and Poorman Creek, however, are V-shaped at the Twisp junction, notwithstanding the glacial U-shapes of their upper ends. The topographic relations indicate that above the Twisp Valley elbow all tributary valley glaciers were confluent with the trunk glacier, whereas below the elbow the tributary glaciers apparently did not reach the trunk valley (Fig. 9). East of Little Bridge Creek even the Twisp Valley narrows to a V-shaped stream-erosional valley. Therefore, the Twisp Valley Glacier probably terminated at or shortly downvalley of Little Bridge Creek.

Figure 9. Maximum extent and inferred surface profiles of local glaciers during the Alpine Phase. Because no recognizable deposits or depositional landforms of the Alpine Phase survive, this interpretation is based wholly on erosional landforms. The downvalley change from U-shaped to V-shaped profiles is taken as the downvalley limit of ice tongues during the Alpine Phase. Cross-hatched areas are ice-dammed lakes inferred from topographic considerations of glacier termini and gradients of the major ice-marginal channels.



49°00'
45'
30'
15'
120°00'
49°00'
ICE CAP
ICE CAP
PASAYTEN—alpine glaciers
SKAGIT alpine
CHELAN BASIN — alpine glaciers
OKANOGAN DRAINAGE BASIN mostly unglaciated
MAXIMUM EXTENT OF GLACIERS DURING ALPINE PHASE
5 0 5 10 miles
GLACIER-SURFACE CONTOUR INTERVAL 1000 FEET
48°00'
45'
30'
15'
120°00'
48°00'

<u>Chewack Valley Glacier</u>. From the Eightmile Creek confluence upvalley, all tributaries that join the Chewack Valley from the west are U-shaped troughs whose bedrock floors hang 400-600 ft above the Chewack Valley floor. However, the first tributary south of the Eightmile confluence, Cub Creek, has a V-shaped cross-profile of a normal stream valley. Thus, during the maximum alpine glaciation the western Chewack tributaries contributed glaciers that were confluent with the trunk Chewack Valley Glacier as far south as Eightmile Creek.

Because most tributaries that join the Chewack Valley from the east head on an upland surface of 6000 to 6500 ft altitude, except in small cirques on the north side of such monadnocks as Tiffany Mountain, they were well below the inferred snowline altitude of 7060 ft (p. 41). Thus, such major eastern Chewack tributary valleys as Thirtymile Creek, Twentymile Creek, and Boulder Creek, as well as smaller tributaries like Ramsey and Pearrygin Creeks, are normal stream-erosional valleys. Only the northernmost tributary valley, Horseshoe Creek (Lat. 48°55' N.), contributed confluent ice to the east side of the Chewack Valley Glacier. The northernmost Chewack tributaries, Horseshoe Creek, Tungsten Creek, Cathedral Creek, and Andrews Creek curiously do not head in cirques. Instead, these U-shaped troughs ascend gradually to a broad upland surface of low relief, suggesting that the source of their valley ice was a small ice cap (Fig. 9). The upland surface is part of the Okanogan Highlands surface (p. 23) that, rising north from an altitude of 4000 ft near the Columbia River, was above the characteristic snowline altitude (7000 ft) north of latitude 48°55' N.

The downvalley limit of the Chewack Valley trunk glacier is more difficult to place than in the Twisp Valley. Although the Chewack Valley is a broad, flat-floored, U-shaped trough having locally truncated spurs nearly to its junction with the Methow Valley at Winthrop, the glaciated aspect is much less distinct below the Eightmile Creek confluence. However, if the Methow Glacier terminated at Carleton as the evidence indicates (p. 54,60), it needed some help from the Chewack ice stream. The Chewack Glacier in Figure 9 is therefore shown to be confluent with the Methow Glacier.

<u>Methow Valley Glacier</u>. Early Winters Valley and the Methow West Fork are deep U-shaped troughs that join the Methow trough accordantly; they were the most important tributaries of the Methow Valley Glacier. Trout, Rattlesnake, and Robinson Creek Valleys, which are U-shaped and have a hanging relation to the Methow trough, apparently also contributed confluent glacier ice. All other Methow tributary valleys have V-shaped cross profiles of their lower ends, and therefore apparently did not contribute confluent ice to the Methow Glacier.

Although all Lost River tributaries head in cirques, their lower ends are V-shaped like the narrow gorge of the trunk canyon. Apparently, therefore, Lost River as far north as Hidden Lakes trough was unglaciated during the Alpine Phase. The largest tributary glacier was in Eureka Creek Valley, whose U-shaped profile terminates 9 mi from its headward cirques, and 2 mi from its Lost River confluence.

Goat Creek, which joins the Methow 6.5 mi south of Lost River, heads in small cirques, but has a V-shaped cross profile near its Methow confluence. Because only a small area in the headwaters is above 6500 ft (extrapolated mean altitude of cirque floors, see p. 41) and because the headwaters face south, Goat Creek apparently was an unfavorable basin for the nourishment of a large valley glacier.

Wolf Creek, which joins the Methow 11 mi south of Early Winters Creek, heads in cirques and has U-shaped upper-valley segments, but its lower 10 mi has the broadly V-shaped cross-profile of a tributary that did not deliver alpine ice to the Methow Valley Glacier. All Methow tributaries south of Wolf Creek have distinctly V-shaped cross-profiles at least along their lower ends. In the trunk Methow Valley, therefore, the downvalley limit of confluent tributary glaciers is Early Winters Valley. As in the Chewack Valley the contribution of confluent ice was decidedly asymmetric, little or none having come from the eastern tributaries (Fig. 8).

The downvalley change from glacial to nonglacial cross-profiles in the Methow Valley is at Carleton (Barksdale, 1941a, Fig. 6). Above Carleton the Methow Valley has a flat alluviated floor 0.75 mi or more wide; downvalley the Methow flows in a bedrock-floored V-shaped canyon. The downvalley end of the Methow Valley Glacier on Figure 8 is therefore drawn at Carleton.

That the limit of confluent tributary glaciers was 6 to 7 mi farther south in the Methow Valley than in the Chewack Valley probably was due to the larger high-altitude catchment basin and of a lower Pleistocene snowline (p.), particularly in the Early Winters and West Fork Valleys, of the Methow system. The comparatively short length (15 mi) of the Twisp Valley Glacier must have been due to a smaller catchment area than in either the Methow or Chewack systems.

ICE-MARGINAL CHANNELS

<u>General Statement</u>. Among the most striking types of land-forms of the Methow region is the dry valley that leads anomalously either across a spur from an upvalley tributary to a downvalley tributary, or along the side of a trunk valley (Fig. 22). These anomalous channels range in size from inconspicuous spur notches a few feet deep and several feet long to such large channels as Alta Coulee, half a mile wide, 6 mi long, and more than 1000 ft deep. Most such channels are so narrow, steep, or crooked that glacier

ice could not have been the agent of erosion; but since one or both ends of most channels have a hanging relation to the main valley floor, they are clearly segments of longer channels that flowed either across or adjacent to glacier ice.

There is as much variation in channel form as in their sizes. Some of the channels are sharp V-shaped gorges; others are steep-walled, flat-floored coulee; others change along their courses from a V-shaped intake profile to a flat-floored coulee profile. Whereas most of the channels have an unmodified stream-erosional morphology and are apparently floored with gravel, others have somewhat subdued U-shaped profiles, and are floored with irregular, hummocky glacial drift. That is, some coulees, generated by streams marginal to expanding ice, were overwhelmed and modified by glacier ice whereas others, generated during deglaciation, were not modified by subglacial erosion. In the latter class it is generally difficult to distinguish a channel that was entirely excavated during deglaciation from a channel cut mainly during glacial advance and occupied only incidentally by deglacial meltwater. Of the channels characterized by glacial morphology there is no direct evidence for distinguishing those formed during alpine glaciation from those formed during expansion of the ice sheet. A further unmeasured effect is that of pre-Late Wisconsin glaciations.

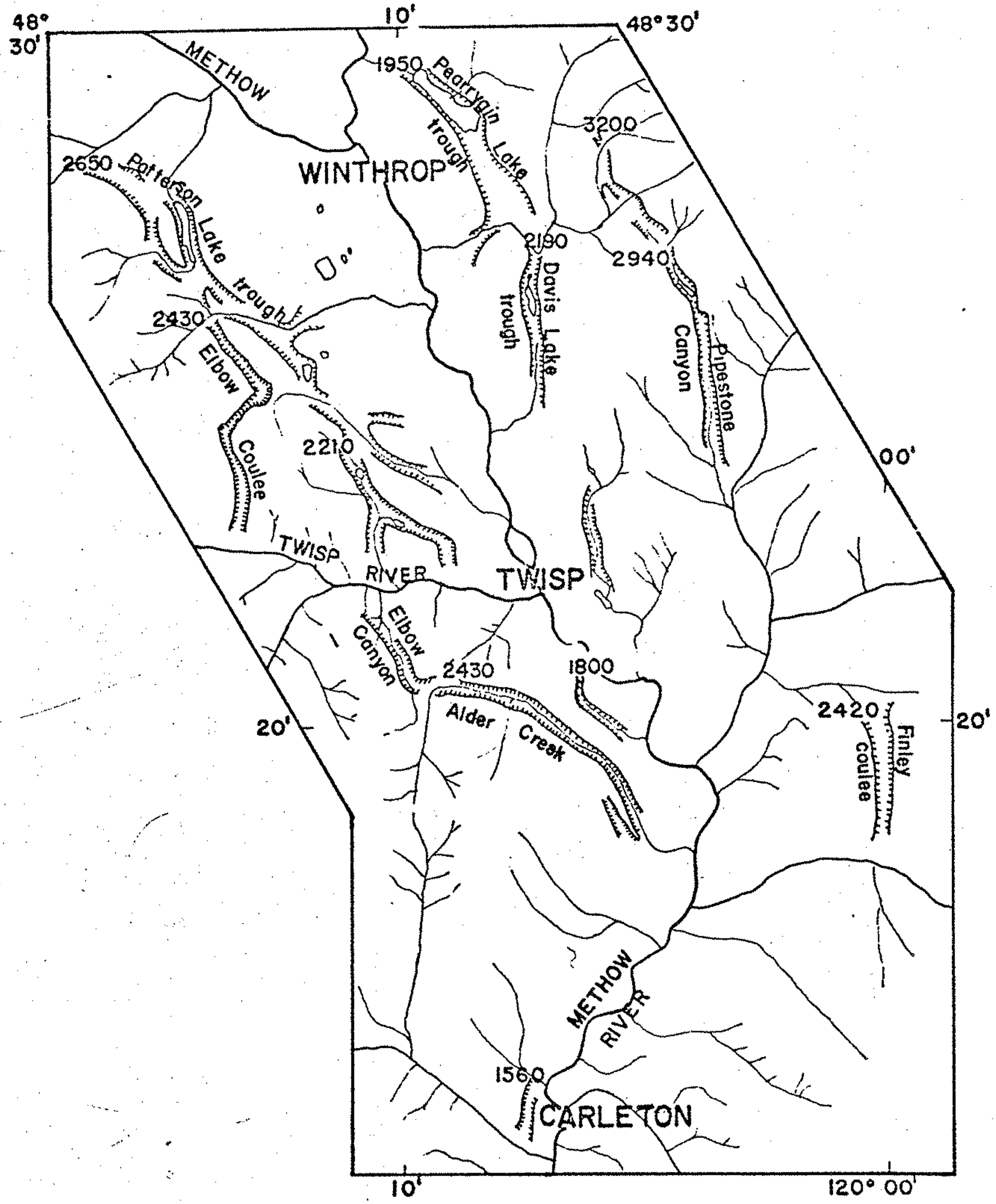
Were the modern valley system invaded by extensive glaciers either of alpine or of ice-sheet origin, many of the old ice-marginal channels would be reoccupied. Except for channels that are clearly related to deglaciation of the ice sheet, it is difficult to assign, a priori, any one channel to a given stage of glaciation, or even to correlate related channels. Nonetheless, by plotting the positions of the more conspicuous channels, some interesting relations are revealed.

By their positions with respect to the Methow Valley, ice-marginal channels are of three types: a) long conspicuous coulees, generally at low altitudes, between Methow tributary valleys, b) short coulees and notches at high to low altitudes between Methow tributaries, and c) channels that enter the Methow drainage basin from another drainage basin.

Methow Valley. The most conspicuous coulee-like ice-marginal channels are concentrated adjacent to segments of the Methow, Twisp, and Chewack Valleys. Figure 10 shows the topographic detail of the complex coulee system adjacent to the Methow Valley.

Two major coulee networks descend along the east side of the Methow Valley. A low-altitude network heads in the Chewack Valley, and descends through Pearrygin Lake trough where it bifurcates, joining the Methow via a direct low-altitude route down lower Bear Creek Valley, and via a

Figure 10. Map showing major coulees adjacent to Methow Valley. Numbers indicate altitude in feet of coulee intake.



PLAN OF COULEE NETWORKS
BETWEEN WINTHROP AND CARLETON

1 0 1 2 3 4 5 6 7 8 9 miles

higher-altitude route through Davis Lake trough. A higher coulee network heads on the eastern side of Chewack Valley and descends south along the mountain front and through Pipestone Canyon into lower Beaver Creek valley, wherethrough it rejoins the Methow south of the Twisp confluence. The southernmost coulee of the network is the anomalous course of lower Finlay Creek valley (Finlay Coulee), which crosses an upland spur between Fraser Creek Valley and Benson Creek Valley. Many minor tributaries and distributaries complicate both coulee networks.

Along the west side of Methow Valley an equally complicated coulee network begins at lower Wolf Creek and 2 mi west of Winthrop, descending through various routes directly to the Methow (lower Thompson Creek, Moccasin Lake trough), and through higher-altitude routes (mainly Elbow Coulee) to lower Twisp Valley. From Twisp Valley the channel network descends along the base of McLure Mountain to the Methow along several subparallel coulees of which the largest and highest is Elbow Canyon-Alder Creek. The eastern spur of McLure Mountain, is truncated not because of ice abrasion but because of ice-marginal stream erosion along its base (Fig. 11). Alder Creek is the post-glacial course of a beheaded tributary that formerly flowed northeast from the saddle between Lookout and McLure Mountains to the Twisp-Methow confluence.

The coulee networks along both sides of the Methow return to the main valley at about the same position 3.5 mi north of Carleton. Though smaller channels occur further south the networks of large coulees do not join tributaries like Texas or Gold Creeks. The southernmost coulee that is 200 ft or deeper is a half-mi-long bypass west of Carleton that heads at the Methow and descends into lowermost Libby Creek.

Because the quantity of ice-marginal drainage generally increases toward the glacier terminus, had these coulee networks developed in association with the ice sheet, their downvalley limit at Carleton would be truly anomalous. Although the ice sheet was deep enough to bury, or almost bury, the divides between the Methow and adjacent drainage basins, and extended southeast of the Methow Valley 25 mi onto the Waterville Plateau, the most conspicuous ice-marginal channels are restricted to low altitudes and to a 20-mi segment of the Methow north of Carleton. Indeed, smaller ice-marginal channels extend to altitudes 4000 ft higher and are found as far southeast as the mouth of the Methow, but the restriction of the deep coulee networks to the Methow Valley between Winthrop and Carleton indicates their development in association with a valley glacier, not with an ice sheet.

The downvalley limit of the coulees at or north of Carleton delineates the terminus of a valley glacier that was stationary at that latitude or advanced to that position during several glaciations. Carleton is also the estimated terminal position of the Methow Valley Glacier based on the downvalley change from a wide, flat-floored valley to a narrow V-shaped canyon (p. 54). The terminus of the Methow Valley Glacier, therefore, is identically placed on two wholly independent lines of topographic evidence.

The altitude of the Wolf Creek intake for the western coulee network is 2650 ft while the intake altitude for the eastern coulee network east of Pearrygin Lake is 3200 ft. Both networks descend 8 mi south to the floor of Methow Valley at 1400 ft. The average downvalley gradients are about 70 ft/mi for the western system and 100 ft/mi for the eastern system.

The coulee gradients probably approximate the average gradient of adjacent segments of the Methow Valley Glacier, which therefore must have been about 1500 ft thick near Winthrop, thinning downvalley to the terminus at Carleton. By such an analysis, the inferred contours across the lower end of the Methow glacier are drawn on Figure 9. Surface-ice contours in the accumulation zone are drawn a few hundred ft above the floors of the highest cirques; the contours in between are interpolated and are based on the altitudes of hanging tributaries and truncated spurs, and on general topographic considerations.

Except Pipestone Canyon and Finlay Coulee,all of the large coulees described above are floored with till and have a locally irregular, hummocky coulee-floor drift morphology. Clearly,they have been overridden by glacier ice. The overriding glacier must have been the Cordilleran Ice Sheet, whose drift extends more than 2500 ft higher on the slopes above the highest members of the coulee systems. Here lies one line of evidence that ice-sheet glaciation was preceded by a full-scale alpine glaciation. Because ice-sheet drift floors the coulees, it is impossible to tell whether the various members of the coulee networks were cut in ascending order during expansion of the Methow Valley Glacier, or in descending order during retreat of the Methow Glacier.

The Pipestone Canyon and Finlay Canyon branches of the eastern network, free of hummocky drift and apparently floored with fluvial deposits, were occupied by drainage marginal to the dwindling Cordilleran Ice Sheet. Finlay Coulee has tributary valleys that are graded to its present valley floor, indicating that whatever modifications it underwent in Late Wisconsin time, it was established in its anomalous course and excavated to its present depth during some earlier glaciation.

Winthrop Area. Two of the coulees, one from the eastern network and one from the western network, are deeper, have lower altitudes, and have gentler side slopes than other sections of the coulee networks. One is currently occupied by Pearrygin Lake, the other by Patterson Lake. They form a symmetrical arrangement with respect to the Methow Valley at Winthrop (Fig. 29); both return via low-altitude routes to the Methow Valley 2.5 mi south of Winthrop. Their distribution and altitude are evidence of diversion of the Chewack River through the Pearrygin coulee, and of Wolf Creek and other west-side Methow drainage through the Patterson coulee owing to the presence of a Methow Valley glacier terminating at Winthrop. Whether the Winthrop stand occurred during a late stade of the Alpine Phase, or was due to a pre-Late Wisconsin alpine advance is uncertain, but the relatively gentle coulee sides and the fact that tributaries are generally graded to the coulee floors, suggests that they are very old.

Twisp Valley. In the Twisp Valley 4 mi east of its great elbow, a channel passes anomalously between the Twisp Valley and the lower end of Little Bridge Creek, analogous to the Methow-Libby Creek channel at Carleton. From the Little Bridge Creek confluence downvalley the Twisp is a V-shaped canyon developed by fluvial and mass-wasting processes. As in the Methow Valley, the downvalley limit of conspicuous ice-marginal channels corresponds to the downvalley limit of a broad, flat valley floor. Thus, the terminus of the Twisp Valley Glacier can also be fixed on two lines of topographic evidence. Although on the south side of Twisp Valley there are higher and smaller channels leading from Buttermilk Creek into Newby Creek (thence into Poorman Creek), from Buttermilk Creek directly into the head of Poorman Creek, and from Buttermilk Creek into the head of Libby Creek, these channels must be the result of successive drainage diversions during the more recent expansion of the ice sheet that eventually filled the Twisp Valley and overtopped the highest nearby peak, Lookout Mountain (p. 80).

Chewack Valley. Meltwater coulees adjacent to the Chewack valleys are conspicuous south of Twentymile Creek on the east side and south of Eightmile Creek on the west side. They lie downvalley from where tributary valleys contributed confluent glacier ice to the Chewack Valley Glacier during the Alpine Phase maximum (p. 51). The most conspicuous ice-marginal channels coincide with segments of the Chewack Valley that, based on the pattern of confluent tributary glaciers, should have had voluminous marginal drainage, indicating development during the Alpine Phase. The distribution of the conspicuous meltwater channels probably approximately outlines the ablation zone of the Chewack Valley Glacier, just as ice-marginal coulees delineate such zones of the Methow and Twisp Valley Glaciers (Fig. 9).

Concluding statement. Stratigraphic evidence of the Alpine Phase, such as alpine drift underlying ice-sheet drift, has not been found in the Methow region, nor has any bona fide alpine drift been found at all. Exposures of drift in tributaries invariably contain far-traveled stones having provenances outside the immediate drainage basin, though from within the Methow sedimentary block. Either the ice sheet managed to erode away all traces of an older alpine drift, or alpine drift is lithologically too similar to ice-sheet drift to be readily distinguished. Therefore, arguments for the Alpine Phase rest wholly on geomorphic evidence. Neither the alpine sculpture of the highlands nor the restricted distribution of prominent ice-marginal channels in the trunk valleys could have been generated by a large ice-sheet of northern origin.

The Late Wisconsin Cordilleran Ice Sheet

INTRODUCTION

It has been obvious for many years that the northern Cascades have been affected by lobes of the Cordilleran Ice Sheet that spilled southward from the accumulation are in interior British Columbia (Daly, 1912). The Puget Lobe of the ice sheet pushed up into Cascade valleys that were previously shaped by alpine glaciers (Cary and Carlston, 1937; Mackin, 1941; Vance, 1937, page 292 ff); the Okanogan Lobe invaded lower Chelan Valley, constructing the 400-ft morainal embankment that impounds Lake Chelan (Water, 1933). In the Harts Pass area of the northern Cascades, Barksdale (1941a,b) demonstrated that tributaries of the Skagit and Methow drainage basins received substantial contributions from the Cordilleran Ice Sheet, which overtopped divides of 7000 ft altitude.

The present chapter includes a restudy of the Harts Pass area and of summits within the Methow Drainage Basin as well as an extension of observations eastward to the Methow-Okanogan divide and southwest along the Cascade Crest to the Methow-Chelan divide.

EVIDENCE OF THE ICE SHEET

Table 3 is a compilation of evidence from summits, ridges, and spurs that were affected by the ice sheet within and peripheral to the Methow Drainage Basin. Among the types of evidence considered useful to a reconstruction of the ice sheet are a) topographic relations that are anomalous to a strictly alpine-glacial terrane; b) ice-flow directional indicators like striations, grooves, stoss-and-lee topography, whalebacks, and distribution of erratics; and c) upper limits of erratics, polish, and striations. Figure 12 is a map of all glacial flow indicators measured on rock surfaces.

Table 3. Summary of Ice-Sheet Indicators

No.	Place	Altitude (ft)	N. lat.; W. long	Upper limit (ft) of glacial evidence: type of evidence	Flow-direction indicators (azimuth): type of evidence	Divide crossed
				t=topography, s=striations, g=grooves, pb=perched boulder, mc= meltwater channels, e= erratic, sl=stoss-and-lee topography, m= moranine, sc= striated cobble, wb = whole-back		Ca=Cascade Ch=Chelan Co=Columbia M=Methow O=Okanogan P=Pasayten S=Skagit
1	Azurite Pass	6700	48°39';120°46'	--	SE : t	S -> M
2	Billy Goat Pass	6600	48°49';120°19'	7800: e, pb	155-162:s,wb	M -> M
3	Black Canyon Channel	5050	48°04';120°08'	5050: t	--	M -> Ch ?
4	Buckhorn Mtns	3850	48°05';119°55'	3850⁺: e,pb,mc	130-135: s,sl	M -> Co
5	Buffalo Pass	6540	48°46';120°42'	--	225: s,g,e,mc	P -> S
7	Buttermilk Butte	5474	48°19';120°18'	5474⁺: s,e	135: s	M -> M
8	Cascade Pass	5400	48°28';121°03'	7500:t,s,pb	SE: s, mc	Ca -> Ch
9	Copper Pass	6900	48°29';120°38'	7300: s	170-200:s,sl,g	Ch -> M
10	Cutthroat Pass	6820	48°33';120°42'	7400: 6	115-135: s,pb	S -> M
11	Dolancy Ridge	8000	48°37';120°35'	6991⁺:s,e,sc	120-207: s	M -> M

Table 3. Summary of Ice-Sheet Indicators (cont'd)

12	Doe Mtn	7154	48°44';120°11'	--	180: s,g	M -> M
	Ike Mtn	7186	" "	7186+:g,pb	--	
13	Driveway Butte	5982	48°38';120°32'	5982+: s,e	130-142:s,sl	M -> M
14	Early Winter Pass	6671	48°30';120°37'	7250: t	100-135:s,sl	M -> M
15	Eightmile Pass	5400	48°48';120°21'	--	100-130:s,g,sl	M -> M
12	Farewell Pk	7439	48°44';120°14'	7300?: e	--	M -> M
18	Glacier Pass	5540	48°40';120°44'	--	140-200:s,t	S -> M
19	Goat Peak	7001	48°38';120°25'	7001+:s,pb,e	145: s,sl	M -> M
20	Granite Pass	6300	48°34';120°41'	--	048-115:s,pb	S -> M
	Tower Mtn	8444	" ; "	7400: wb,t	--	"
18	Grasshopper Pass	6740	48°41';120°43'	--	142: s	S -> M
21	Hancock Ridge	7125	48°39';120°39'	6600?:sc	087-118:s,wb	M -> M
22	Harts Pass	6198	48°43';120°40'	--	100-150:s,e	S -> M
23	Isabella Ridge	8200	48°44';120°20'	7960+: S	090-180:s,g	M -> M
24	Last Chance Pt	7046	48°41';120°34'	6700?:pb,e	100-122: s	M -> M
25	Leecher Mtn	5020	48°15;120°00'	5020+:s,e	000-110: s	M -> M
26	Lookout Mtn	5692	48°14';120°11'	5692:s,e,pb	160-168:s,sl	M -> M

Table 3. Summary of Ice-Sheet Indicators (cont'd)

27	Mebee Pass	6500	48°38';120°47'	--	SE: t	S --> M
28	Methow Pass	6580	48°35'	7000^+:pb,t,t	SSE: t	M --> S
29	N. Navarre Pk	--	--	--	--	-->
30	Ninetynine Pk	7405	48°42';120°41'	7380^+: s	155-188: s	S --> M
31	N. Twentymile Pk	7437	48°46';120°05'	7437^+?:pb,e	255: s	M --> M
32	Old Goat Mtn	5228	48°01';119°59'	4310: e	144:g,sl	M --> Co
33	Rainy Pass	4850	48°31';120°44'	--	SE: t	S --> Ch
34	Rattlesnake-Trout Ridge	6300	48°40';120°37'	6300^+: s	090-160:s	M --> M
35	Raven Ridge	8600	48°14';120°20'	6200^+:wb	SE:wb	M --> M
36	Robinson Pass	6220	48°45';120°38'	6970^+: s,e	155-174:s,g	P --> M
36	"Robinson Pass W."	6180	48°44';120°39'	--	145:s,g,wb,sl	P --> M
37	Sandy Butte	6076	48°42';120°27'	--	007-030: s	M --> M
	Storey Peak	7821	"	6900^+:t	--	
38	Scramble Pt	6340	48°42';120°32'	6400^+:e,s	130-220: s	M --> M
39	Setting Sun Mtn	7253	48°42';120°27'	7125: s,e	137-190: s,wb	M --> M
22	Slate Peak	7440	48°44';120°41'	7400^+: e	155-203: s,e	P --> M
	Slate Pass	6940	" "	--	192: s	P --> M

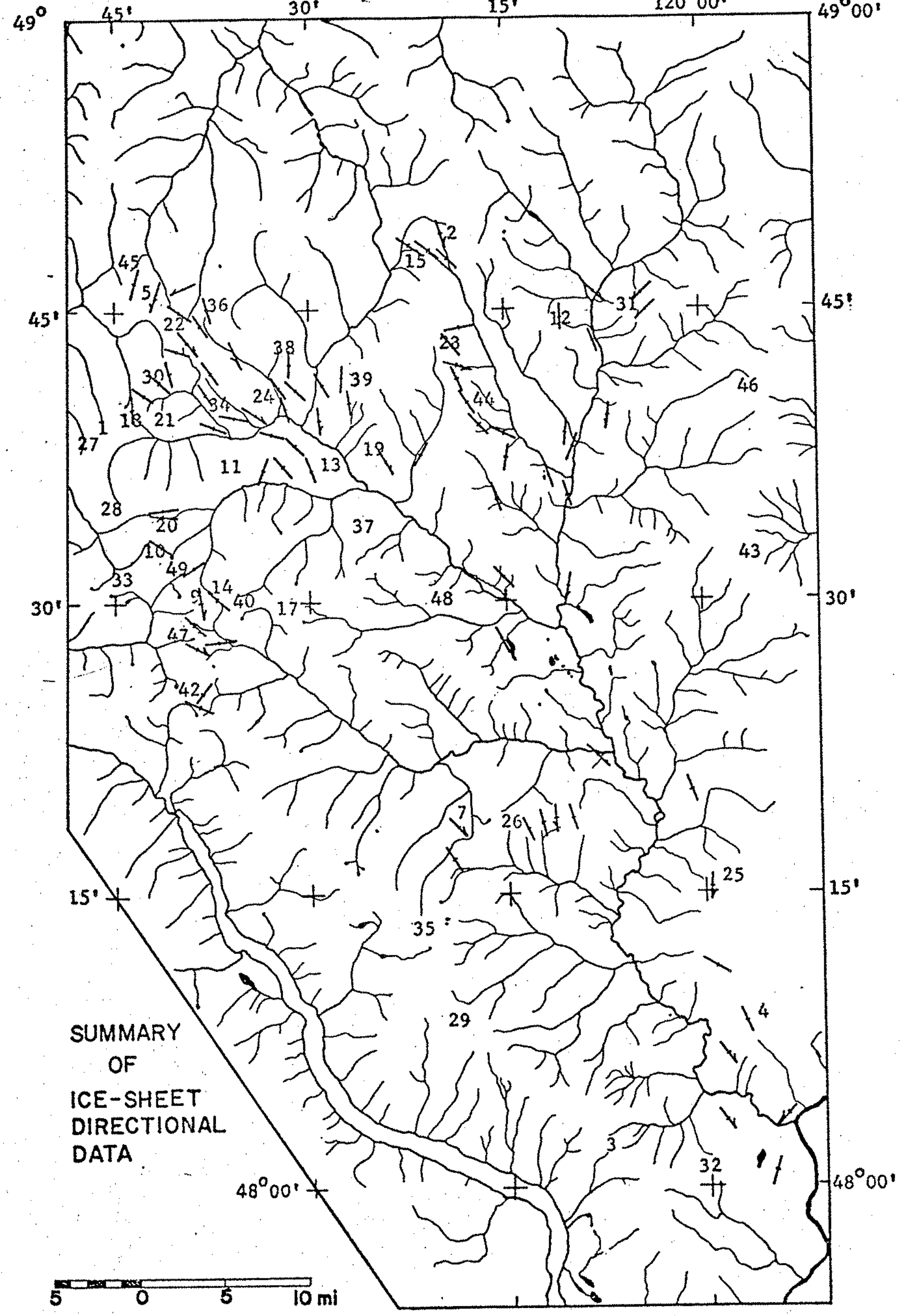
Table 3. Summary of Ice-Sheet Indicators (cont'd)

28	Snowy Lakes Pass	6850	48°36';120°43'	7000^{+}:pb	S: t	M → S
42	South Pass	6300	48°25';120°38'	7500:pb,s,sc	010-135:s	Ch → M
43	Starvation Mtn	6765	48°32';119°57'	6250?:sc,m	--	?
44	Sweetgrass Butte	6104	48°40';120°18'	6104^{+}:s	145-170:s,sl	M → M
45	Tamarack Pk	7290	48°46';120°43'	7290^{+}:e,sc	--	P → S?
30	Tatie Peak	7386	48°42';120°42'	7386^{+}:s	130-150:s,sl	S → M
46	Tiffany Mtn	8278	48°40';119°56'	7600:pb,t	SE:t	M → O
47	Twisp Pass	6064	48°28';120°38'	7065^{+}:s,pb	110-170:s, sl,wb	Ch → M
47	"Twisp Pass SE"	6260	48°27';120°39'	--	115-130:s, sl	Ch → M
48	Virginian Ridge	6470	48°31';120°22'	6470^{+}:t	SE:wb	M → M
49	Washington Pass	5460	48°32';120°39'	--	040-080:s	C → M?
45	Windy Pass	6257	48°46';120°43'	--	193:e,mc	P → S

Figure 12. Summary of ice-flow directional data, and index map for Table 3. Numbers, approximately located, correspond to those of Table 3.

striations, or whaleback

unique indicator (*e.g.* striated stoss-and-lee surface)


49°
45'
30'
15'
120°00'
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2
15
45
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36
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12
45'
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4
SUMMARY
OF
ICE-SHEET
DIRECTIONAL
DATA
3
32
48°00'
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10 mi

THE METHOW ICE STREAM

Harts Pass, Robinson Pass and Northern Cascade Crest. Because of glacial striations on Harts Pass and Psayten Pass and because of the southward distribution of granodiorite erratics of Pasayten-stock provenance, Barksdale (1941a, Fig. 2) inferred that ice, having flowed up the Pasayten Valley entered Methow tributaries over Harts and Robinson Passes. He reasoned that the source of a least some Methow Valley ice was, therefore, the Cordilleran Ice Sheet.

Dozens of measurements on glacial striations whale-backs, and stoss-and-lee topography, and occurrences of Pasayten-stock granodiorite erratics confirm Barksdale's inference that a high ice-sheet flowed generally southeast through the Harts Pass-Robinson Pass area (Fig. 17). Further evidence of ice-sheet glaciation springs from the deductive consideration that although each pass area is positioned between two cirques that must have been actively competing during the Alpine Phase, the passes are not the type of ragged cols characteristic of alpine terranes. Instead, they are flat or gently rounded passes hundreds of feet broad. The formerly ragged alpine cols have been lowered several hundred feet--apparently by abrasion beneath the high ice sheet responsible for the striations and erratics on the passes.

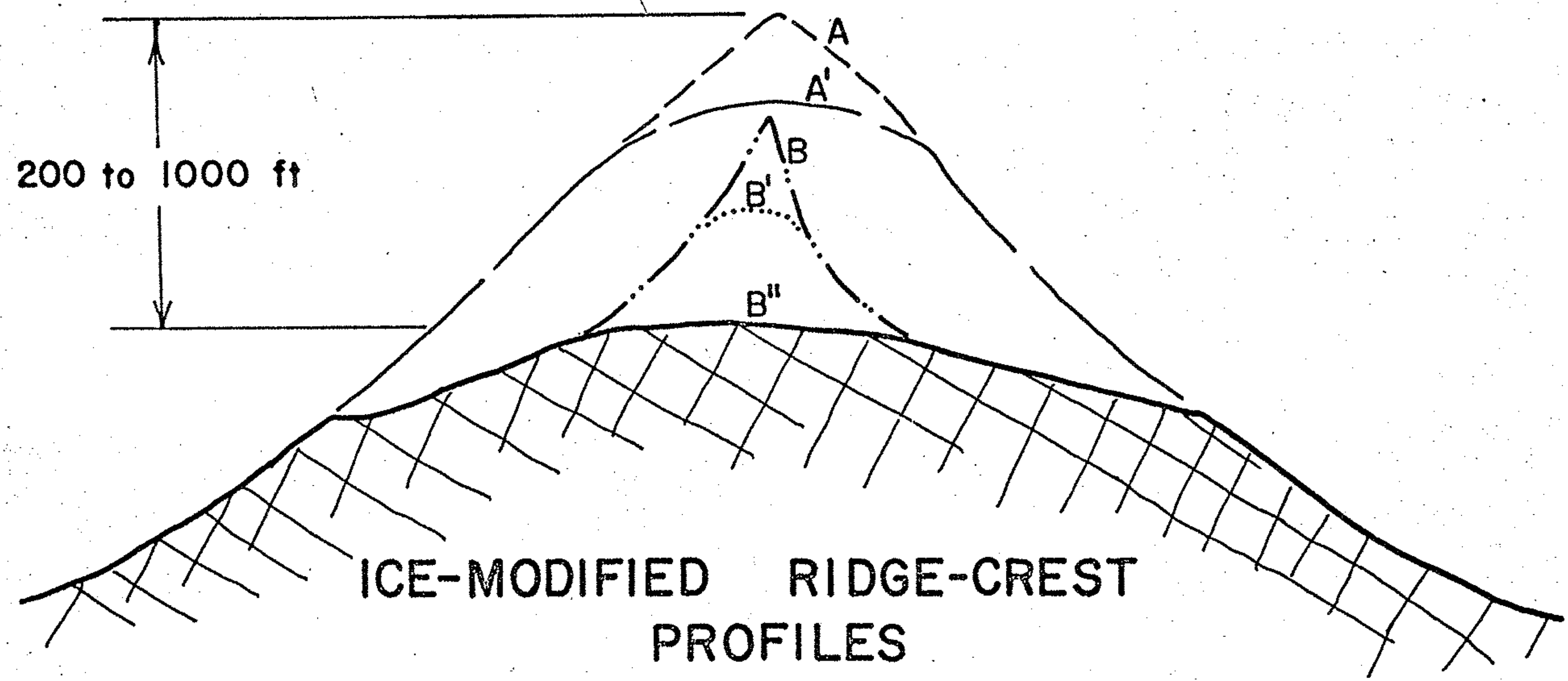
On Buffalo and Windy Passes, anomalously broad saddles across the Cascade Crest north of Slate Peak, perched granodiorite erratics of Pasayten-stock provenance record the former passage of glacier ice southwest across the ridge. Glacial striations and erratics near the summit of Slate Peak and on the summit of Tamarack Peak indicate that none of the northern Cascade Crest protruded the surface of the ice sheet. Not only was ice delivered across the divide via the passes, but the entire ridge was buried; the passes are merely where the thickest ice streams crossed the divide.

Subglacial erosion was much more effective where the width of the ridge was narrowed by retreat of cirque head-walls during the Alpine Phase (Fig. 13) and where the ice that overtopped the ridge was thickest. Broad saddles like Buffalo Pass and Windy Pass were developed only where the ridge crest had previously been narrowed by competing alpine cirques.

Tatie Peak, Glacier Pass, Azurite Pass, Mebee Pass. Striations near the summits of Tatie Peak and Ninetynine Peak indicate that the entire Cascade Crest between Harts Pass and Glacier Pass was overridden by ice that flowed southeast from Skagit tributaries into Methow tributaries (Fig. 17). Grasshopper Pass is an especially broad segment of the ridge that was selectively more deeply eroded because it was the headward col of Trout Creek during the Alpine Phase.

Figure 13. Transverse profiles of ice-modified ridges.

A	Typical ridge between trunk valleys occupied during the Alpine Phase.
A'	Same ridge beveled off by ice-sheet erosion.
B	Typical ridge at site of competing cirque heads.
B'	Same ridge somewhat beveled off by ice sheet erosion.
B''	Same ridge greatly beveled off by ice-sheet erosion.



ICE-MODIFIED RIDGE-CREST PROFILES

During the Alpine Phase, Glacier Pass was probably a ragged col between competing cirque glaciers in the heads of the Slate Creek South Fork (Skagit drainage) and Brush Creek (Methow drainage). The southward-expanding ice sheet was channeled up Slate Creek between Mt. Ballard and Azurite Peak on the west and the Cascade Crest on the east. The only barrier to flow along the otherwise efficient conduit of the Slate South Fork was the ragged col, which, therefore, the ice sheet beveled off to form the low-altitude (5560 ft) Glacier Pass 1400 ft broad. If the mean slope of the original cirque headwalls was 45°, the altitude of the former col was reduced some 700 ft.

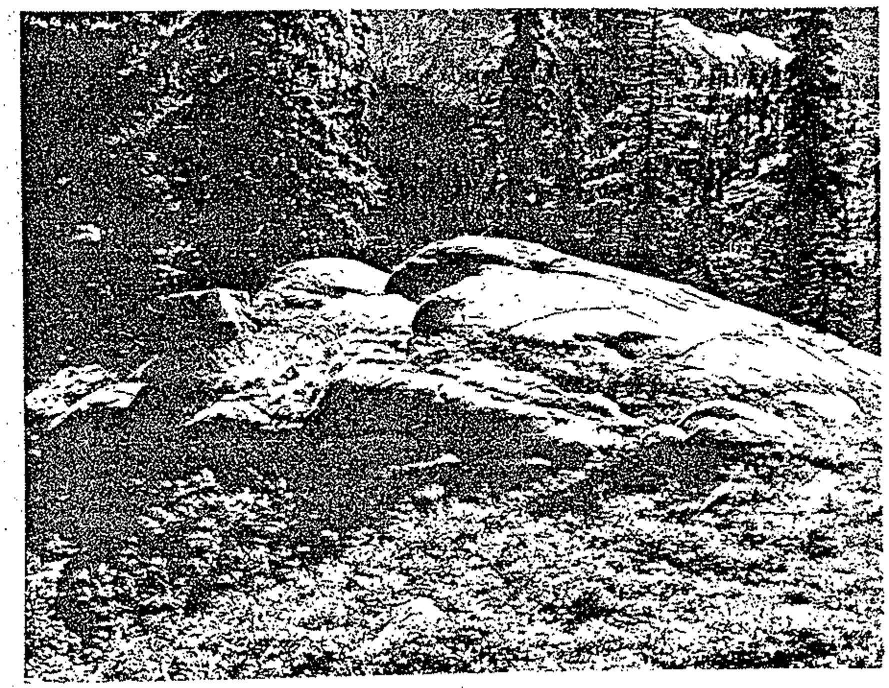
Southwest of Azurite Peak topographic relations similar to those at Glacier Pass exist at Azurite Pass and Mebee Pass, where Cordilleran ice flowing up Mill Creek and up East Creek (Skagit tributaries) overtopped former alpine cols, thereby invading the Methow West Fork. The parallel, northwest-trending alpine-glacial troughs apparently were such effective conduits for the ice sheet that, except for lowering former cols that were positioned directly across the troughs, the ice sheet modified the topography but little.

The Cascade Crest between Methow Pass and Rainy Pass. The Cascade Crest south of Mebee Pass is characterized by high ragged peaks, by broadly rounded ridgecrests, and by low trough-shaped passes. Methow Pass and Snowy Lakes Pass at the head of the Methow West Fork are strewn with perched boulders of Golden Horn Granite and apparently were lowered many hundreds of ft as the Cordilleran Ice Sheet spilled southward from the Methow West Fork into Swamp Creek (Skagit) drainage.

Two mi southeast of Snowy Lakes Pass the Cascade Crest is crossed by two U-shaped troughs (Granite Pass), each striated parallel to their east-west trends, apparently as Cordilleran ice invaded the head of Pine Creek. The lower part of the south spur of Tower Mountain is gently rounded and conspicuously marked with subhorizontal bedrock whalebacks (Fig. 14); the upper parts of the spur and the mountain itself have a normal ragged alpine aspect. The altitude (7400 ft) of the topographic change approximates the former surface of the Cordilleran Ice Sheet.

Cutthroat Pass, a broad trough across the Cascade Crest 1.5 mi southwest of Granite Pass, has perched boulders up to 15 ft diameter and striations trending east-southeast, indicating passage of the ice sheet across the divide from Porcupine Creek (Skagit) into Cutthroat Creek (Methow). On the north spur of Cutthroat Peak, the upward change from smoothly abraded to ragged morphology occurs at about 7300 ft, approximating the altitude of the ice-sheet surface.

The above examples show that the delivery of Cordilleran ice into Methow tributaries was not confined to the Harts Pass-Robinson Pass area but that the ice-



sheet invaded the Methow drainage basin from the northwest, having flowed freely across the divide as far as 25 mi southeast of Harts Pass.

Washington Pass. Hornfelsed mudstone and sandstone erratics on Washington Pass are foreign both to State Valley and to the cirque at the head of Early Winters Valley. The nearest possible source is the contact aureole of the Golden Horn Batholith exposed 6 mi north on Delancy Ridge, implying movement of ice up Early Winters Valley and westward across Washington Pass into the Chelan drainage basin, a flow pattern that is compatible with the east-west striations on Washington Pass. However, cummingtonite in the metamorphic-mineral assemblage of the erratics renders the Delancy source unlikely (Peter Misch, personal communication, 1972). The probable provenance, according to Misch, is the North Creek volcanic-sedimentary suite exposed west of Granite Creek (see Misch, 1966, map), which implies that ice flowed southeast across the Cascade Crest, almost normal to the trend of State Valley (Fig. 7). Although the second hypothesis accords with the trends of striations on Cutthroat and Granite Passes (p. 74), the lack of North Creek erratics at those places and the east-west trend of striations at Washington Pass, do not. Until the provenance of the erratics is precisely located, they indicate only that the most recent glacier affecting the area was the ice sheet flowing generally southeast, normal to either ice flow pattern implied by Figure 8.

Early Winters Pass, Copper Pass, Twisp Pass, SE Twisp Pass, South Creek Pass. Most of the cirques clustered at the head of Twisp Canyon show evidence not only of alpine-glacial landform development, but of more recent erosion by the ice-sheet. Although the col between cirques at the heads of north-draining Early Winters Creek and of southeast-draining Twisp Creek is but a few tens of feet broad, its overall rounded aspect as well as an abundance of southeast-heading striations up to 7250 ft altitude on its southwest shoulder indicate ice flow southeast across the pass.

Southwest along the divide whalebacks and striations trending southeast and stoss-and-lee topography on Copper Pass indicate unequivocally that ice flow was to the southeast into the head of Twisp canyon, not out of it. Striations were found up to an altitude of 7300 ft on the broad northeast shoulder of the col.

The Twisp Pass area has abundant striations trending south-southeast and is almost 1000 ft broad, indicating lowering of the former col by many hundred ft during ice-sheet glaciation. Conspicuous well-striated stoss-and-lee surfaces in the pass area (Fig. 15) indicate beyond doubt that ice flow was southeast into the head of Twisp Canyon, not out of it. Striations on the summit of nearby Lincoln

Butte indicate that the ice stream across Twisp Pass was at least 1000 ft thick. Another col ("SE Twisp Pass") one mi southeast of Twisp pass is anomalously broad and striated parallel to its trend. A divergence of striations southeast of Twisp Pass (Fig. 12) is probably a dynamic effect of the ice sheet that, having flowed 1000 ft deep through Twisp Pass, distended toward the Copper Pass ice stream, which was only 200 ft thick.

That the ice invaded Twisp Valley over cols at the heads of Alpine-Phase cirques that were tributary to Chelan trough indicates, beyond doubt, that Chelan trough has been glaciated by the Cordilleran Ice Sheet.

South Creek Pass and the ridgecrest rising north have numerous striations and perched boulders up to an altitude of 7300 ft. Striations trending east-southeast to southwest, diverge into the head of South Creek. Although unique flow-directional indicators were not found, ice flow probably paralleled the pattern of Twisp Pass, flowing east <u>into</u> the Twisp tributary. The diverging flow pattern, like that near Twisp Pass, probably reflects distention of the confined flow of ice across the pass. Another low, broad col between South Creek Pass and "SE Twisp Pass" also probably transferred a small stream of Cordilleran ice into Twisp Valley.

<u>Sawtooth Ridge, Raven Ridge, Buttermilk Butte, Lookout Mountain</u>. For 12 mi southeast of South Creek Pass the divide between Chelan and Methow drainages has the generally ragged aspect of an unmodified alpine-glacial landscape. Although a few such low sags in the divide as at the head of War Creek probably transmitted small streams of Cordilleran ice to Twisp canyon, the divide was almost entirely a nunatak above the Cordilleran Ice Sheet.

The Chelan-Methow divide southeast of Twisp Valley is particularly well suited for a comparison of the geomorphic effects of alpine and of ice-sheet glaciations. The headwaters of Libby and Gold Creeks are alpine cirques that lead within a few mi downvalley to V-shaped stream valleys apparently unmodified by alpine ice (see Fig. 9 and p. 47). The local glaciers, small and independent tongues that they were, could not have overtopped major ridges. Most upland ridges south of Twisp Valley, however, bear the unmistakable effects of glaciation. The summit of Buttermilk Butte has southeast trending striations and is strewn with quartz-diorite erratics of Black Peak Stock provenance. Quartz diorite erratics copiously strewn in the headward tributaries of Libby and Gold Creeks further indicate that the ice sheet flowed southeastward and oblique to the Twisp Valley Fault, apparently because of the necessity to flow east of the Raven Ridge nunatak. Whalebacks trending southeast, normal to the trends of alpine valleys, are further evidence that the topography excavated by the small northeast-flowing Alpine Phase glaciers was later overprinted by an enormous southeast-flowing ice sheet (Fig. 16).

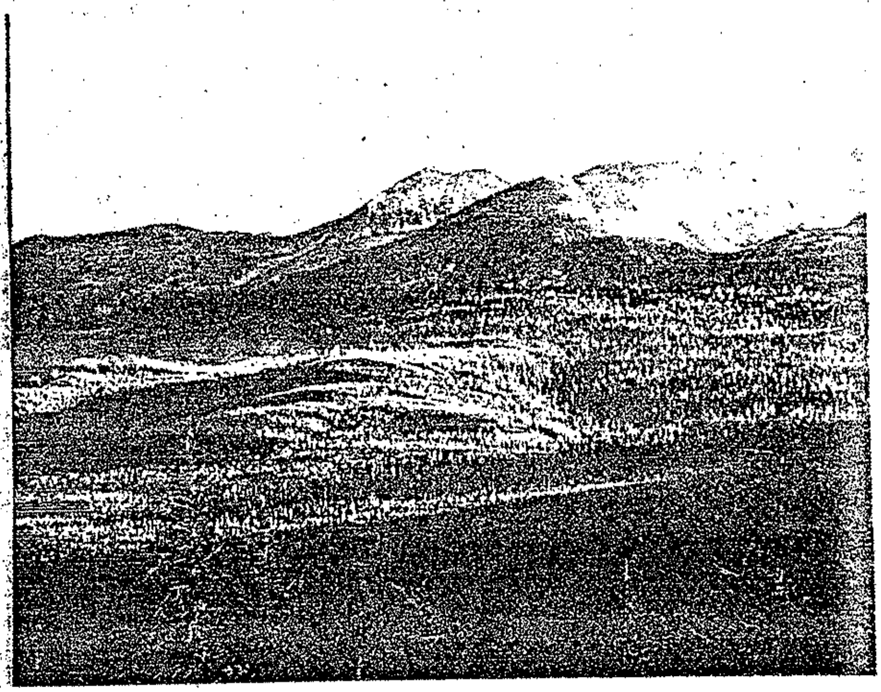
Lookout Mountain and its lower ridges, strewn with arkosic erratics and well striated generally south-southeast, indicate that the ice-sheet flowed south from Twisp Valley into Libby Creek, more or less parallel to the ice-flow indicators cited above along the mountain front.

Methow Mountains. Although exposures are generally poor, a careful search along the summit fire lane freshly bull-dozed in 1970 along the summits of the Methow Mountains, south of North Navarre Peak failed to reveal a single erratic or any boulders of local lithology that could not be explained by normal weathering and mass-wasting processes. I agree with Waters (1933, p. 787) and Barksdale (1941a, p. 723) that these summits were not overridden by the ice sheet. Fresh erratics from the Virginian Ridge, Ventura and Winthrop Formations indicate, however, that a Late Wisconsin ice sheet bearing diagnostic Methow lithologies flowed freely along the Methow Mountain Front over divides at least as high as 4700 ft.

By comparison with the sharply rounded unglaciated ridge crests on the Chelan side of the Methow Mountains, the lower ridge crests on the Methow side of the mountains are a) much more broadly rounded, b) characterized in

Figure 16. View south from Buttermilk Butte showing cirques excavated along Raven Ridge during Alpine Phase. Whalebacks trending southeast across beveled-off spurs (middle ground) indicate that the Cordilleran Ice Sheet later overwhelmed the region from the north-west, flowing approximately perpendicular to the trend of the former Alpine Phase valley glaciers.

Figure 19. Boulder of locally derived mesocratic gneiss perched on north ridge of Doe Mountain. Boulder is on crest of ridge; it could not have been deposited there in its precarious position by mass-wasting processes.



in detail by bedrock whalebacks generally aligned south-southeast, and c) crossed by anomalous short channels that generally descend southeast across spurs. The upper limit of these topographic peculiarities is about 6200 ft in upper Libby Creek, 5600 ft in the Gold Creek South Fork, and about 4500 ft in the Black Canyon Left Fork, probably reflecting the general gradient of the Late Wisconsin Ice Sheet.

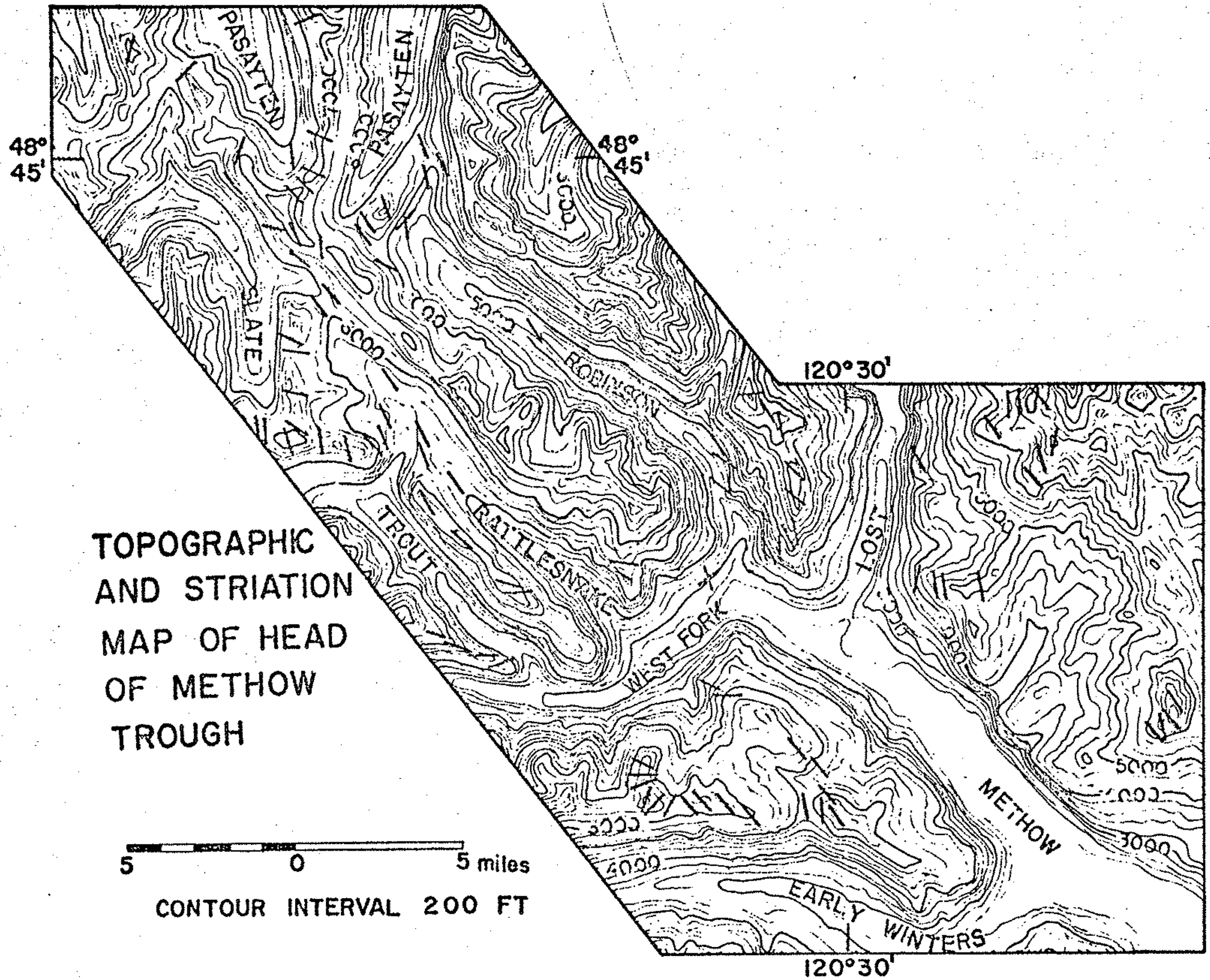
The most striking topographic anomaly of the Methow Mountains is a flat-floored coulee 700 ft wide and about 2500 ft long that appears to be an overdeepened and widened segment of competing tributaries at the heads of Black Canyon Creek (Methow drainage) and Gold Creek (Chelan drainage). The channel-floor altitude of 5040 ft is the lowest pass between Black Canyon tributaries and Chelan trough. The only logical origin of the channel is an ice-dammed lake outlet across the mountains into Chelan basin.

Erratics of gray arkose (Winthrop Formation ?), black mudstone, quartz-K-feldspar porphyry, and Golden Horn Granite found on a northeast spur of Old Goat Mountain, the easternmost summit of the Methow Mountains, occur up to an altitude of 4310 ft. Barksdale (1941a, p. 723) reported erratics up to 4415 ft; he quoted R. F. Flint as having found erratics up to 4530 ft and reported that A. C. Waters found erratics up to 4675 ft. Possible ice-marginal stream notches occur up to 4910 ft altitude 1.5 mi north of Goat Peak.

Columbia Valley. Insofar as the Methow Mountains were apparently a nunatak, erratics of Ventura conglomerate and Virginian Ridge conglomerate in morainal drift at altitudes of 3480 ft in Antione Creek, a Columbia tributary south of the Methow Mountains, is remarkable. At the Methow-Columbia confluence, however, the Methow sector of the ice sheet joined the Okanogan Lobe proper that, as reported by Waters (1933) and indicated by striations on Figure 12, flowed along the Cascade mountain front as far south as Chelan trough. Having joined the Okanogan Lobe, Methow Valley ice maintained its integrity as an ice stream along the mountain front at least as far south as Antoine Creek.

Trout Creek, Rattlesnake Creek, Methow West Fork. Striations in upper Rattlesnake, Trout, and Robinson Creek Valleys indicate that the ice sheet, having entered the Methow drainage basin over the Cascade Crest, was channeled southeast, parallel to the topographic corrugations (Fig. 17). Striations along the Trout-Rattlesnake ridge crest, however, swing gradually from southeast-trending at the north end of the ridge to east-west trending two mi farther southeast. These transverse striations, which also characterize the southeast ends of the Hancock Ridge,

Figure 17. Topographic map of the upper Methow trough, showing striations in tributaries and on ridgecrests. Note general convergence of striations toward the head of Methow trough.



TOPOGRAPHIC
AND STRIATION
MAP OF HEAD
OF METHOW
TROUGH

Last Chance, and Scramble spurs, indicate that within 3 mi of the Methow West Fork the ice flowed parallel to the trunk valley. Although West Fork ice clearly overflowed Delancy Ridge southward into Early Winters Valley (below), the ridge was apparently a barrier that caused most of the ice to follow the West Fork Valley to the head of the Methow trough (Fig. 17). Complementary evidence that ice flowed east along the West Fork consists of Golden Horn Granite erratics on Driveway Butte, where southeast-trending summit striations indicate that the flow of ice was transverse to the end of Delancy Ridge and parallel to the Methow trough.

Delancy Ridge, Sandy Butte. On Delancy Ridge erratics occur at least as high as 6991 ft, the summit of a horn between cirques at the heads of Hangry and Driveway Creeks. On the summit of the horn striations trend southeast; on the ridge to the north striations trend generally east-west; on the ridge to the east striations trend generally south. Although ice clearly flowed southeast over Delancy Ridge into Early Winters Valley, flow was generally parallel to the topography, being clockwise around the horn and transverse to the secondary ridges. The modified col at the head of Cataract Creek no doubt transmitted a thin ice stream across the ridge; only the Needles could have protruded the maximum ice-sheet surface.

A cluster of peaks (Silver Star Mountain, Mt. Gardner), all high enough to have been nunataks, caused a bifurcation of ice-sheet flow south of Early Winters Valley. Golden Horn erratics on Sandy Butte (6076 ft) and the anomalous broadness of the north and east spurs of Storey peak up to all altitude of 6900 ft indicate that the eastern arm flowed southeast over Sandy Butte and Virginian Ridge, parallel to the Methow trough.

<u>Setting Sun Mountain, lower Lost River, upper Methow Trough, Goat Peak</u>. A ridge generally higher than 6000 ft extends east of the upper Methow trough and culminates in several summits 8000 ft or higher. Striations and dikerock porphyry erratics of northern provenance up to 7125 ft on arkose of Setting Sun Mountain indicate that southeast-flowing ice overtopped most of the ridge. Striations on the north slope of the mountain, however, trend north-northeast, parallel to the Lost River canyon to the north.

The striations of Figure 17 delineate a flow pattern within the ice sheet that was convergent toward the head of the Methow trough. Thus, despite the overflow of ice across divides, the greatest flux of ice was parallel to the northern trunk valleys and into the Methow trough. As well as serving as the trunk valley that gathered tributary glaciers during the Alpine Phase, the upper Methow trough was the main conduit through which the Cordilleran Ice Sheet transected this part of the northern Cascades.

The polished summit of Goat Peak (7000 ft) is striated southeast like other striations closer to Winthrop (Fig. 12), indicating that ice across the uplands flowed generally parallel to the upper Methow trough.

Hancock Ridge, Last Chance Point. Between Last Chance Point and a point 5 mi northwest along the same ridge no evidence of glaciation was found above altitude 6700 ft. The upper limit of striated cobbles on Hancock Ridge, 5 mi to the west-southwest, is 6600 ft. Southeast of these areas, however, striations occur at least as high as 6991 ft on Delancy Ridge, 7125 ft on setting Sun Mountain, and 7000 ft on Goat Peak (above). Because the ice-sheet surface sloped southeast, the apparent southeast rise of surface-ice indicators, and in particular the low limits on Last Chance and Hancock Ridges, must be in error. Extrapolated from more reliable evidence to the north and south, the ice-sheet surface must have been at least 7100 ft in this area; it is contoured as such on Figure 20.

Isabella Ridge, Sweetgrass Butte. Although east of Setting Sun Mountain an imposing ridge borders Lost River Valley, the high striations on Setting Sun Mountain, and yet higher striations on Isabella Ridge, indicate that the ice sheet overtopped part of the ridge and spilled southward across the maturely eroded upland that descends generally southeast towards Winthrop.

On Isabella Ridge east to southeast-trending striations as high as 7200 ft on Burgett Peak, and up to at least 7960 ft on the southeast flank of Sherman Peak record the flow of ice obliquely across the sharply rounded ridges and into ragged cirques that face upper Eightmile Valley. Up to 7600 ft, striations are of the normal closely spaced type 1 to 3 mm deep and having rounded cross-profile contours. Striations between altitudes 7600 and 7960 ft, however, are shallower (0.5 to 1 mm), are more widely spaced (0.5 to 2 cm), have angular cross-profile contours, and occur on flat barely abraded joint-block surfaces. Although the trend and the lengths (5 to 20 cm) of these high striations are consistent with those of "normal" striations below 7600 ft, their unusual morphology indicates something different about their origin. Perhaps they were etched by joint blocks that,while being quarried by an overriding ice sheet, scratched the basal joint surface. Not only the unusual joint-block tool but a fairly light overburden of ice might have promoted the unusual morphology of these striations. That striations occur up to 7960 ft implies that the pronounced pass immediately west of Isabella Ridge carried an ice stream 1000 ft thick from Lost River Valley into the mature upland between the Methow and Eightmile Valleys. Striations on Sweetgrass Butte, along Sweetgrass Ridge, on Goat Peak and at other localities within this upland reflect the general southeast flow of the ice-sheet, generally parallel to the bordering master troughs of Eightmile Creek and the upper Methow (Fig. 12).

THE CHEWACK ICE STREAM

Evidence of Glaciation. Mesocratic and weakly foliated gneissic and plutonic rocks east of the Chewack-Pasayten Fault extend as far north as the Canadian border (Fig. 2). Most glacially deposited boulders in this area are of the same lithology as the underlying bedrock; true erratics are rare. To be useful as evidence of glaciation, boulders must be precariously perched or otherwise in positions impossible to a mass-wasting origin. Unlike the muddy till of the sedimentary-rock belt, till east of the Chewack-Pasayten Fault is sandy, not unlike grus or coarse fluvial sand. Moreover, coarse-grained crystalline rocks are not striated or fluted by glacier ice nearly as readily as the fine-grained Cretaceous sedimentary and volcanic rocks. Therefore, the determination of upper-ice limits is more difficult east of the Chewack-Pasayten Fault than within the sedimentary-rock belt.

Billy Goat Pass, Eightmile Pass. The extensive rugged upland developed on resistant plutonic rocks between Tatoosh Buttes and Monument Peak (Fig. 2) forced a bifurcation of the south-flowing Cordilleran Ice Sheet. One stream, funneled around the western side through Pasayten lowland, entered the Methow Drainage Basin over the divide between Robinson Mountain and Azurite Peak, as discussed above (p. 71-74). A second stream was channeled up the Pasayten East Fork and entered the head of Lost River mainly through the Hidden Lakes Trough (see p. 29). There it found two outlets, a) southwest along lower Lost River Valley, partly overflowing the Setting Sun-Isabella Ridge divide as explained above, and b) southeast over a complicated upland into Drake Creek (a Lost River tributary), thence into Eightmile Valley via Eightmile Pass and Billy Goat Pass.

Eightmile Pass (5300 ft), the lower of the two passes, is a U-shaped trough having abundant striations trending southeast, parallel to the trend of the valley. Stoss-and-lee surfaces similar to those at Twisp Pass (Fig. 15) indicate, unequivocally, that Cordilleran Ice flowed southeast into the head of Eightmile Valley.

Though much higher than Eightmile Pass, Billy Goat Pass (6700 ft) is anomalously broad and is marked with striations and whalebacks that trend south-southeast, parallel to the valley walls. Like Eightmile Pass, the only tenable explanation of the topography is that Cordilleran ice overwhelmed the headward col of competing Alpine Phase cirques aligned southeast across the divide, and reduced the former col to a broad, saddle-shaped ridge.

Perched boulders on nearby Burch Mountain are conspicuous up to 7600 ft; probably evidence of glaciation extends up to the summit (7800 ft), more nearly approaching the altitude of striations on nearby Isabella Ridge. Thus, Cordilleran ice flowed into the head of Eightmile Creek at least 900 ft thick across Billy Goat Pass, and at least 2500 ft thick across Eightmile Pass.

Falls Creek. Falls Creek trends nearly parallel to the much deeper adjacent Eightmile Valley to which it connects through two short, U-shaped lateral passes that hang 1300 and 1800 ft, respectively, above the floor of Eightmile Valley (Fig. 18). A beveled-off pass (7000 ft) at the head of Falls Creek indicates that Cordilleran ice spilled southeast into Falls Creek just as it did through Eightmile and Billy Goat Passes into the head of Eightmile Valley.

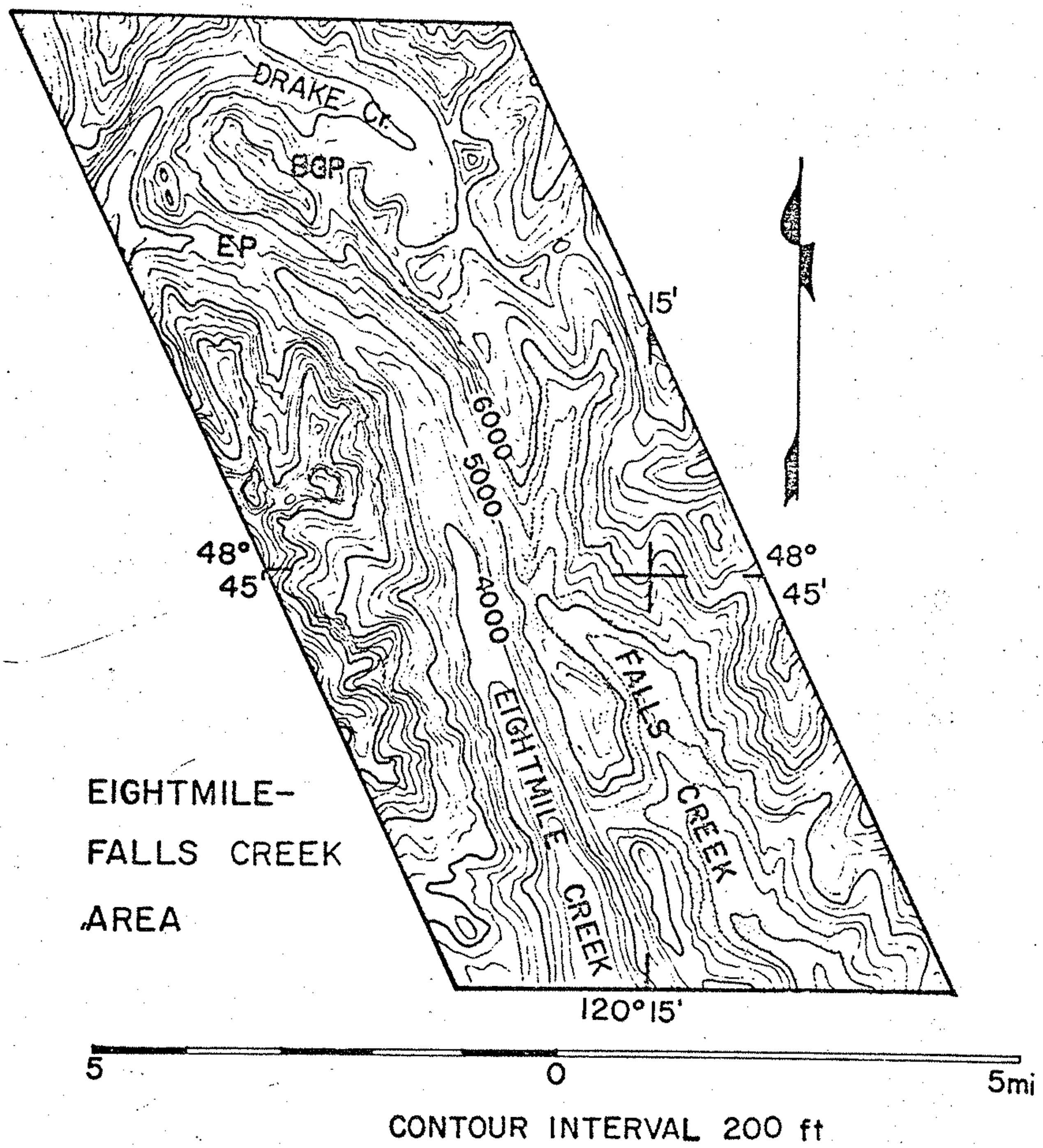
Falls Creek is east of the Chewack-Pasayten Fault; both it and its northern neighbors are underlain entirely by quartz-diorite gneiss (Fig. 2). Nevertheless, drift on the floor and walls of upper Falls Creek contains 1 percent of erratics of porphyritic gray andesite. Andesite does not crop out in Falls Creek Valley, nor in any northern neighbor; the only possible source is the belt of Newby Group andesite forming Isabella Ridge and underlying Eightmile and Billy Goat Passes.

Andesite clasts in the drift along upper Falls Creek indicates that the most recent ice in the valley

Figure 18. Topographic map of the Eightmile Creek-Falls Creek area. Note lateral passes between the two subparallel valleys. Stipple pattern marks known extent of Newby andesite clusts in Falls Creek drift.

BGP = Billy Coat Pass

EP = Eightmile Pass


DRAKE Cr.
BGP
EP
15'
6000
5000
4000
48°
45'
48°
45'
FALLS
CREEK
EIGHTMILE
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120° 15'
EIGHTMILE-
FALLS CREEK
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5
0
5mi
CONTOUR INTERVAL 200 ft

did not originate in the cirque heads, nor did it enter via the 7000-ft pass at the head of the valley. Rather, it must have entered from Eightmile Valley through the lateral passes (Fig. 18). Topographic relations, not unlike those near Twisp and Copper Passes, permitted the ice sheet to issue with greater flux through the lower Eightmile Pass (5300 ft) and Billy Goat Pass (6700 ft), outstripping the flow over the Falls Creek Pass (7000 ft). Eightmile ice naturally would have distended into Falls Creek Valley, mainly through the two lateral passes. Andesite pebbles in the lower end of Falls Creek Valley indicate that Eightmile ice, once through the lateral passes into the upper valley, followed Falls Creek Valley southeast to the Chewack confluence.

An alternative hypothesis for the entry of Eightmile ice into Falls Creek is that the ice flow, having been cut off over the 7000-ft Falls Creek pass because of downwasting, continued to invade Eightmile Valley through Eightmile Pass. From Eightmile Valley it would naturally have flowed over the northern lateral pass into Falls Creek. In upper Falls Creek andesite clasts confined below 6200 ft on the ridge crest and below 6000 ft on the valley floor (Fig. 18), suggest that andesite-bearing drift was deposited during deglaciation when the ice surface probably was about 6500 ft high over Eightmile Pass. At that time Billy Goat conduit would have been deglaciated as well as Falls Creek Pass, and ice entering Falls Creek would necessarily have flowed entirely via the Eightmile Pass conduit.

Doe Mountain, Farewell Peak, Upper Chewack Valley, No. Twentymile Peak. The ridge northeast of Falls Creek culminates in several summits between altitudes of 7150 and 7450 ft. The summits and lower spurs of Doe Mountain (7154 ft) and Ike Mountain (7186 ft) are strewn with locally derived gneiss boulders, some of which are precariously perched (Fig. 19), indicating ice-sheet glaciation. On the other hand, nearby Farewell Peak (7439 ft) and its flanks above 7300 ft show no evidence of glaciation.

Flutes on the summit of Ike Mountain trend north-south, parallelling Chewack Valley, the pronounced glacial trough to the east, in which striations also trend south. The ice sheet, though channeled southeast along Eightmile and Falls Creek Valleys, flowed south along Chewack Valley, thereby converging south-southeastward, as shown on Figure 12.

North Twentymile Peak (7437 ft), the highest summit immediately east of the Chewack Valley bears perched boulders up to an altitude of 7000 ft; equivocal evidence of glaciation extends to the summit. Striations up to 7000 ft on the broad south flank of the mountain trend southwest, probably recording a clockwise diversion of the ice sheet around North Twentymile Peak. The low-relief upland at 6000 to 6500 ft altitude extending east and southeast to the Tiffany Mountain monadnock area was clearly beneath the surface of the ice sheet.

Tiffany Mountain-Starvation Mountain area. Most of the Okanogan highland is a low-relief surface that descends from 6800 ft near North Twentymile Peak to 6000 ft near Starvation Mountain. A line of monadnocks between Tiffany Mountain and Starvation Mountains, inclusive, rises 800 to 1400 ft above the surface.

The summits and upper slopes of Tiffany Mountain (8242 ft) "Little Tiffany" Mountain (7988 ft), Middle Tiffany Mountain (7972 ft), and Rock Mountain (7980 ft), are characterized by block fields of locally derived angular rock debris showing no evidence of having been glaciated. Like other highlands in the Methow region, these peaks have well-developed cirques that open to the north and east. Between Tiffany and Rock Mountains, however, a col that formerly separated opposing cirques has been reduced to no more than an abrupt rise along the northwest-southeast glacial trough that now transects the mountains. Unless one of the two competing Alpine Phase cirque glaciers performed the unlikely feat of reversing its flow, thereby overwhelming its own headwall, the topography dictates that the area has been affected by the Cordilleran Ice Sheet.

Freezeout Ridge, which leads to Tiffany Mountain from the southeast, is flat, 400 ft broad, and strewn as high as 7600 ft with subrounded boulders. Several anomalous tors, apparently remnants of a formerly more rounded ridge crest, rise 30 ft above the flat ridge crest. Between the tors Freezeout Ridge apparently has been beveled off by ice-sheet erosion. One mile north of Freezeout Ridge a flat-crested spur that leads to "Little Tiffany" Mountain from the west is, like Freezeout Ridge, strewn with unwe unweathered subrounded boulders up to 7600 ft; the mountain slope above is covered with blockfields of angular, locally derived weathered rock. The west slope of Rock Mountain up to 7600 ft is interrupted by long, narrow nearly horizontal meadows whose occurrence in the apparently homogeneous local plutonic rock is inexplicable unless they were carved by glacier ice or by ice-marginal drainage. Three ice-limit indicators, therefore, place the maximum ice-sheet surface near Tiffany at 7600 ft. A tongue of the ice sheet, channeled southeastward along the Boulder North Fork (Methow) and Salmon (Okanogan) valleys, must have flowed 400 ft thick over the modified col.

Except for one well-rounded boulder on the summit of Starvation Mountain (6735 ft), which may have been placed there during construction of a radar installation, the upper limit of glacial evidence is a moraine of boulder till 1.5 mi farther south at altitudes of 5600 to 5800 ft.

Vaguely striated, subrounded boulders of local lithology are found up to 6250 ft, but a mass-wasting origin of them is not disproved. The upper-ice limit near Starvation Mountain is, therefore, somewhat uncertain except that its minimum altitude was 5800 ft. High summits like Granite Mountain (7366 ft) and Starvation Mountain must have projected above the ice sheet, as Tiffany and Rock Mountains did 12 mi farther north. In between, summits like Clark Peak (7900 ft), Mt. McCay (7500 ft), and Old Baldy (7844 ft) complete a discontinuous line of nunataks around which the ice sheet must have mostly bifurcated.

The Chewack Valley is the only glacial trough in the Methow region aligned north-south: all others trend northwest. Glacial striations parallel to the valley sides indicate that the last ice to occupy the valley, the Cordilleran Ice Sheet, flowed south despite the general push of ice from the northwest (p. 95). However, because the higher peaks of the Okanogan Mountains constituted a nunatak, the ice sheet necessarily would have been deflected southward, parallel to the Chewack Valley. The ice-sheet, though flowing generally from the northwest, thus aided the excavation of the north-south Chewack Valley trough.

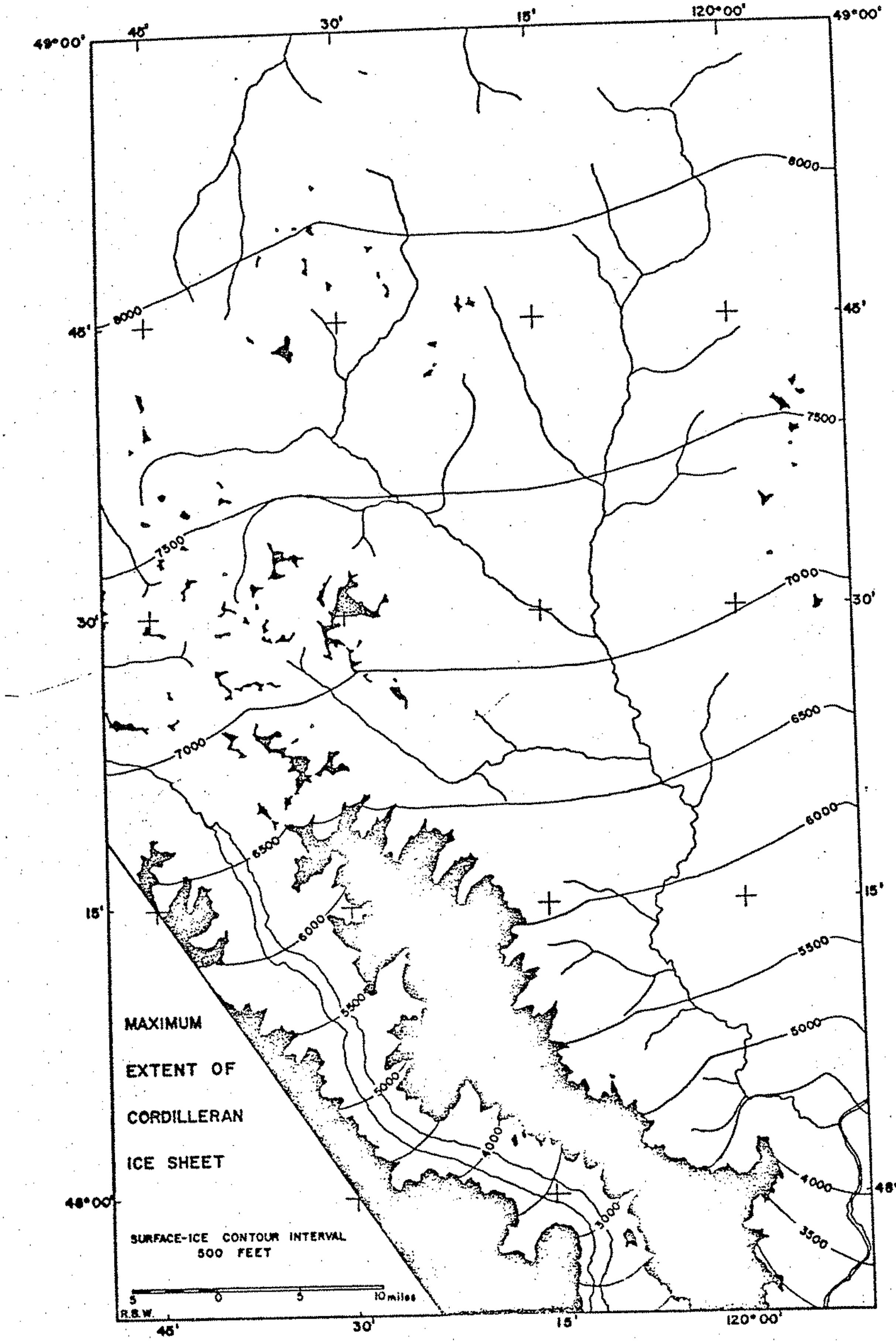
Southern Okanogan Mountains. South of Starvation Mountain the Okanogan Highlands are a low-relief erosion surface that, unlike the Tiffany-Starvation region, is uninterrupted by monadnocks. Because the southern highlands are mostly mantled with stratified sandy drift, only a few summits exhibit erosional evidence of the ice sheet.

Leecher Mountain (5012 ft), the highest summit in the southern Okanogan Highlands, has a small polished surface at the summit striated south-southeast; black mudstone and gray arkose erratics, both of northwestern provenance, indicate that ice-flow was southeast and that the summit as well as the surrounding low parts of the upland, lay beneath the ice surface. The Buckhorn Mountains, having sedimentary-rock erratics, southeast-trending striations, local stoss-and-lee asymmetry and a southeast distribution of dikerocks from local sources (also reported by Barksdale, 1941b, p. 4-5), are evidence that the ice sheet flowed southeast across the southernmost part of the Koanogan highlands.

LIMITS AND FLOW DIRECTIONS OF ICE SHEET

Figure 20 depicts the distribution and flow pattern of the Cordilleran Ice Sheet during its maximum stand in the Methow drainage basin. Striations indicate that in most parts of the area ice generally flowed southeast to south-southeast. The ice was thickest and the flux therefore greatest along the broad troughs of the upper Twisp, upper Methow, and the Chewack tributaries. Striations and drift in the northeastern part of the area indicate that the flow was south, parallel to the Chewack trough.

Figure 20. Inferred maximum extent and surface contours of the Cordilleran Ice Sheet. The Chelan sector implies an extension by at least 400 mi^2 over the compilation by Crandell (1965, Fig. 2), who placed the terminus in Granite Creek at about lat. 48°40' N., just off left margin of map area.


MAXIMUM
EXTENT OF
CORDILLERAN
ICE SHEET
SURFACE-ICE CONTOUR INTERVAL
500 FEET
49°00'
48°00'
45'
30'
15'
120°00'
8000
7500
7000
6500
6000
5500
5000
4000
3500
3000
5 0 5 10 miles
R.S.W.

In all parts of the Methow Valley ice-sheet flow was influenced to some degree by the underlying topography, as indicated by the trends of low-altitude striations subparallel to the trunk valleys. Striations southeast of the Methow Mountains and the Buckhorn Mountains trend northeast, indicating southwest flow of the Okanogan Lobe and the confluent Methow ice, parallel to the Columbia Valley.

In contouring the maximum ice-sheet surface on Figure 20, concessions were necessarily made in the few places where the observed upper limits of drift are incompatible with the regional pattern. If, for example, the evidence for ice limits at 7600 ft and more than 8000 ft are accepted for Tiffany Mountain and Isabella Ridge, respectively, the inferred limit of 7300 ft on Farewell Peak (p. 95) is impossibly low. Contours on Figure 20 are drawn consistent with the regionally most reliable evidence. Summits like Farewell Peak, N. Twenty-mile Peak, and Last Chance Point (p. 88) apparently were beneath the maximum ice-sheet surface, even though evidence of glaciation at those places is, at best, vague.

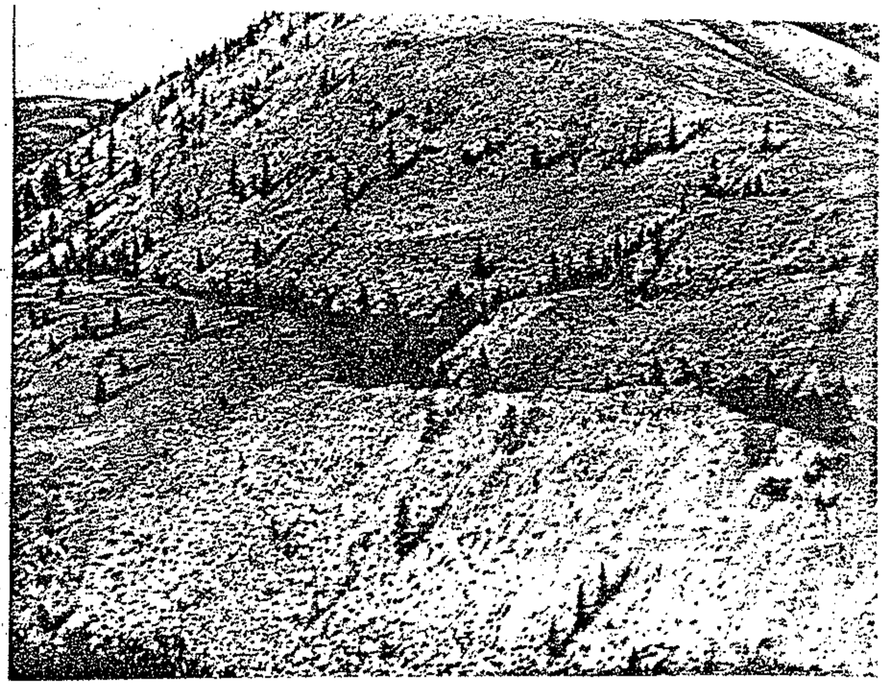
From a height of more than 8000 ft at Isabella Ridge, the ice surface descended at an average gradient of 60 ft/mi southeast to 4500 ft on Goat Mountain. Although the entire southern Okanogan Highlands were below the ice sheet, the summits of the Methow Mountains apparently were not (p. 81). Of particular importance is the origin of the coulee across the Methow Mountains between the Methow and Chelan basins (p. 83). Methow ice descending along the Methow Mountain front as shown in Figure 20, renders a simple ice-marginal lake drainage over the divide impossible, because the drainage would follow the ice margin southeast and discharge into Columbia Valley. However, such a lake would have been dammed in the lower Methow if the arrival of Columbia ice at the mouth of the Methow Valley preceded the Cordilleran ice descending Methow Valley. Because passes into the upper Okanogan valley from the ice reservoir in interior British Columbia are much lower and broader than the comparatively high, restrictive passes into Methow tributaries, the earlier arrival of the Okanogan Lobe is deductively logical. Whatever the origin of the Methow-Chelan channel, it, like the entire crest of the Methow Mountains, is mantled with Glacier Peak Tephra and therefore was abandoned at least by 12,000 to 13,000 years B. P. (Fryxell, 1965; Wilcox, 1969).

COMPOSITION OF TILL AND RELATION TO BEDROCK

Till within the Methow sedimentary-rock belt consists of a poorly sorted gravelly sandy mud (Fig. 21); pebbles and larger stones constitute about 30 percent of the deposit, sand about 25 percent. East of the Chewack-Pasayten Fault, on the other hand, till is a gravelley

Figure 21. Exposure near Pearrygin Lake of poorly sorted, rudely stratified, gravelly sandy mud--typical of till within the Methow sedimentary rock belt. Lighter clasts are plutonic, gneissic, and dike rocks; darker clasts are andesite, arkose, and mudstone. Abundance of shale and friable muddy sandstone of local bedrock is responsible for muddy matrix of till.

Figure 22. View northeast of hanging ice-marginal channel along valley side north of, and descending toward, lower Finlay Canyon. It is a good example of a channel not modified by overriding glacier ice; it was therefore cut during deglaciation.



sand; pebbles and larger stones compose 50 percent of the deposit, sand and granules about 50 percent. These variations in grain size closely reflect the underlying bedrock. Within the sedimentary-rock belt, where shale and muddy sandstone are common bedrock lithologies, the till matrix is muddy; in the crystalline-rock tracts where there is very little fine-grained rock that is friable or fissile, the till is sandy. The underlying bedrock likewise determines the quantity and size of gravel clasts in till. In crystalline-rock tracts, the percent of gravel is higher and the mean size larger (cobble-boulders), owing to the propensity for widely spaced joints in plutonic and gneissic rocks. Although andesite and chert-conglomerate erratics up to 12 ft in diameter occur in the lowland till, large boulders are much less common than in the uplands.

Several rock types from the Methow sedimentary block (see Table 1, diagnostic lithologies) are sufficiently resistant to glacial abrasion to have survived transport tens of miles downvalley of the nearest bedrock source. Because most of these resistant lithologies crop out only north of the latitude of Winthrop, their occurrence in drift in the lower Methow Region is undeniable evidence that the most recent glaciation was by a southeast-flowing ice sheet.

Variations in composition of drift stones are distinguishable in the Winthrop area along east-west transect, normal to the general trend of the Methow Valley. The cobble-sized fraction of drift composing the Winthrop kame-moraine (p.134) is about 45 percent fine-grained leucocratic plutonic rocks and dikerocks, 25 percent gray arkose, 15 percent porphyritic gray andesite, 9 percent coarse-grained leucocratic granite and granodiorite, and 3 percent each of Virginian Ridge and Ventura chert conglomerate. The percentage of Ventura and Virginian Ridge lithologies in drift decreases gradually eastward toward Pipestone Canyon; dike rocks and quartz-dioritic plutonic and gneissic stones increase markedly, and andesite and arkosic graywacke remain constant. These changes in drift-stone lithology parallel eastward changes in bedrock lithology 15 to 20 mi farther north (Fig. 2). The eastward increase of crystalline rocks in Methow till probably records the influence of a Chewack ice stream that maintained its identity below its confluence with the Methow ice stream. East of Pipestone Canyon, till stones are entirely of crystalline rocks. The flow of the ice sheet, therefore, was parallel to the boundary of the Methow sedimentary block.

Till stones on the western side of the Methow Valley are a mixed assemblage of the northern Methow lithologies. Golden Horn Granite, Ventura and Virginian Ridge conglomerates, and Midnight Peak andesite, all of which crop out west of the upper Methow Valley, are common constituents of till exposed up to an altitude of 4700 ft along the Chelan-Methow mountains. The abrupt westward increase in size and quantity of quartz dioritic erratics in upper Libby and Gold Creek drainages probably reflects a divergence of ice southeast from the Methow-Chelan crystalline block into the Methow sedimentary block, as outlined on page 80.

The widespread distribution of sedimentary, volcanic, and dikerock lithologies of upper Methow provenance dictates that the last glaciation of the lower Methow region, where these rocks do not crop out, was by the southsoutheast-flowing Cordilleran Ice Sheet.

REVISION OF LIMITS OF CORDILLERAN ICE SHEET IN NORTH CASCADES

Tributaries to upper Chelan trough display a basically alpine-glacial topography; but well-rounded ridgecrests and spurs, and broadly U-shaped passes aligned northwest-southeast bespeak the same type of topographic modifications that overprint alpine landforms in the Methow region. Moreover, the Chelan trough, like the Methow region, lacks the alpine moraines typical of alpine valleys farther south.

That glacier ice crossed divides from Chelan tributaries into the head of Twisp Valley (p.78-79) proves ice-sheet glaciation of the Chelan trough. A major extension into Chelan trough therefore is required of the ice-sheet margin estimated by Crandell (1965, Fig. 2), who inferred that the terminus lay in the Granite (Skagit) drainage basin.

At the head of the Stehekin River, Cascade Pass and its northern shoulder, Sahale Arm, are anomalously broad and rounded up to altitude of 7400 ft and both are striated and have perched boulders. Several V-shaped channels, apparently relic meltwater channels, descend southeast across Sahale Arm from Cascade Creek (Skagit) to Stehekin River (Chelan). Apparently Chelan basin received Cordilleran ice that flowed up Cascade Valley and southeastward over Cascade Pass. Several anomalously broad passes along the Cascade Crest between Cascade Pass and Rainy Pass suggest that the flow of Cordilleran ice across the Cascade Crest into Chelan trough was nearly as common as into the Methow drainage basin. Although Cordilleran ice evidently passed into Chelan trough along several conduits, the largest must have been Rainy Pass, a deep U-shaped trough between the heads of Granite Creek (Skagit) and Chelan drainage. The inferred 7400 to 7500-ft ice-limits on Tower Mountain and Sahale Arm (Table 3) indicate that the ice stream across Rainy Pass was about 2500 ft deep.

Cascade Valley and Granite Valleys, like Chelan trough, completely lack looped valley moraines. Although alpine-glacial erosional sculpture dominates the landscape of these valleys, the lack of complementary alpine drift renders these valleys more nearly like the Methow and Pasayten Valleys than the strictly alpine-glacial valleys of the middle Cascades. The lack of alpine moraines in a North Cascade valley apparently can be taken as evidence that it was occupied by the Cordilleran Ice Sheet.

The dearth of sediment in Lucerne Basin of Chelan trough indicates that the ice-sheet probably terminated south of the Narrows in Wapato Basin, where sediment is several hundred ft thick (Whetten, 1967, p. 255). The most likely position for the terminus is at Wapato Point, an arcuate peninsula of Chelan valley-derived drift that, despite Waters (1933) contention of its antiquity, is sufficiently unweathered to be of Late Wisconsin Age (Barker, 1968, p. 43). Although haystack erratics are not unequivocal evidence of ice-sheet glaciation from the east (p. 120-121), the occurrence at Lakeside of striated cobbles of basalt, a lithology foreign to the Chelan trough, is undeniable evidence that the Cordilleran ice ascended lowermost Chelan Valley. The chief revision of Waters' (1933) and Barker's (1968) interpretations is that the contemporaneous ice tongue descending Chelan trough was not locally derived alpine ice, but was a bona fide tongue of the Cordilleran Ice Sheet.

Within Chelan trough the upper limit of anomalously rounded spurs descends from 6600 ft at 48°25' N. Lat. to

6000 ft 15 mi southeast, suggesting a gradient of 40 ft/mi for the surface of the Cordilleran Ice Sheet in upper Chelan trough. In order that the inferred ice-sheet surface in upper Chelan trough meets the upper limits of drift in lower Chelan trough (Barker, 1968), the Chelan sector of the ice sheet is contoured steeper on Figure 20 than the inferred surface along the eastern surface of the Methow Mountains.

Deglaciation of the Cordilleran Ice Sheet

STRATIGRAPHY

Stratigraphic relations in the Methow region, mostly from exposures south of Winthrop, are fairly straightforward. Stratified drift, whether sandy gravel or finely laminated silt, invariably overlies nonstratified drift that, because of its content of well-striated stones, is demonstrably till. Laminated lacustrine silt commonly is capped with 5-20 ft of fluvial gravel, nowhere by till. Where stratigraphic exposures are absent the hummocky or kame-and-kettle topography of the stratified glaciofluvial drift rises above, and is apparently built upon, the less-hummocky till surface. Stratigraphic evidence indicates, therefore, that ice-sheet glaciation, which deposited a discontinuous but ubiquitous till blanket, was followed directly by deglaciation that partly eroded the till and deposited glaciofluvial, fluvial, and lacustrine sediments.

Although most exposures of stratified drift in the lower Methow region are undisturbed, a few such places as the riverside cliff at the mouth of McFarland Creek, the Methow channel bottom half a mi upstream of the Gold Creek confluence, and exposures along the Pateros-French Creek road 2.5 mi northwest of Pateros show laminated silt or sand that is distinctly contorted. The origin of the distortion of these deposits cannot definitely be proved, but insofar as no evidence was found indicating that glacier ice deformed these deposits, the contortions probably denote collapse of stratified drift upon removal of glacier ice.

ICE-MARGINAL LAKES

Finely laminated silt and fine sand in parts of the lower Methow Valley and its tributaries record the former existence of lakes temporarily dammed by residual valley ice that blocked the lower ends of the tributaries. Stratified lake silt in the Black Canyon Left Fork up to 1800 ft above the floor of the Methow Valley, and near the mouth of Black Canyon up to 650 ft, record lakes dammed successively at altitudes of 2700 ft and 1535 ft. Stratified silt at various altitudes up to 2000 ft in Gold Creek, Libby Creek, and Texas Creek, indicates that ice-dammed lakes in tributaries were a characteristic of ice-sheet deglaciation of the lower Methow region. Stratified lake silts to an altitude of 3840 ft on both sides of the Loup-loup summit pass record ice-dammed lakes early during deglaciation that were nearly at the level of the divide.

Although the lakes undoubtedly discharged through short coulees that cross spurs into downvalley tributaries, the coulees are so numerous that correlation of former lake surfaces with possible outlets is difficult. Because no lake silts are overlain by till or were deformed by overriding glacier ice, the lakes clearly relate to final deglaciation of the valley. The whole array of lake sediments and outlet coulees were products of ice-marginal streamflow that was graded to progressively lower, stagnating ice masses. Silts of lower altitude and closer to the tributary mouths are therefore somewhat younger than higher silt deposits, even though there are no distinguishable variations in their degrees of weathering.

ICE-MARGINAL CHANNELS

General Statement. In addition to the wide, deep, and locally concentrated coulee systems described on pages 56-62, several thousand smaller coulees segments are distributed from the Methow headwaters to the Methow-Columbia confluence, and from valley floors to the crests of all but the highest ridges. Because of their widespread distribution they clearly relate to ice-sheet glaciation, rather than to the Alpine Phase. Their great numbers and their occurrence across almost every upland pass below 6000 ft altitude as well as on virtually every spur downvalley from Winthrop renders mapping of each coulee impractical.

The coulees are of two types: ice-modified coulees cut during expansion of the ice sheet, and unmodified coulees cut during deglaciation (Figs. 22 and 23). The size and density of both types of coulees increase with decreasing altitude and with distance from the headwaters. The ice-modified coulees provide minimum upper limits of ice sheet glaciation (Table 3); the unmodified coulees record successively lower ice-sheet surfaces during deglaciation. The especially impressive array of unmodified coulees indicates the large volume of ice-marginal drainage associated with deglaciation of the Methow region.

Channels across major Methow divides. Some of the unmodified coulees do not lead between adjacent tributaries, but enter the Methow drainage basin across major divides. One such divide between the Pasayten Middle Fork and the head of Robinson Creek is notched by two passes, Robinson Pass and "West Robinson Pass" (Table 3; Fig. 17). A flat-floored channel 3 to 8 ft deep descends across the western pass, entering Robinson Valley through an anomalous, narrow, V-shaped gorge that could only have been excavated by late-glacial meltwater and probably was the outlet of an ice-dammed lake in the head of the Pasayten Middle Fork. Smaller channels with intakes up to 800 ft above the passes, and descending generally south into Robinson (Methow) valley, occur only on the spurs that lead to the passes from the west. Because the channels must have formed in the natural gutter between the glacier margin and the valley side, the confinement of these channels to the western spurs indicates that the ice sheet surface sloped west or southwest, parallel to its slope during the glacial maximum.

Across Buffalo Pass and Windy Pass, dry channels that have multiple intakes descending and uniting westward indicate that during deglaciation the ice sheet was higher to the east--just as it was along this part of the Cascade Crest during the ice-sheet maximum.

A channel system leading southeast across Eightmile Pass descends to an impressive dry cataract, plunge-pool, and gorge 200 ft wide and 0.25 mi long, that forms the head of Eightmile Valley. The huge cataract must have been the outlet of a lake dammed in upper Drake Creek by downwasting valley ice. The geometry requires that the ice surface sloped southeast: during downwasting of at least the upper 2000 ft of the ice sheet, the glacier surface retreated roughly parallel to the slope during the ice-sheet maximum and during earlier phases of deglaciation (Figs. 20 and 18).

The highest of the conspicuous dry coulees descending along the west side of the Okanogan Valley intersects the head of Finlay Canyon, forming an ice-marginal conduit leading across the divide from Okanogan into Methow drainage. During deglaciation at the 6000 ft level, therefore, the surface of the ice sheet was higher over the Okanogan Valley than over the Methow Valley. Southeast of the Buckhorn Mountains a myriad of smaller dry coulees that descend across the divide from Watson Draw (Columbia) basin toward the lowermost Methow likewise indicate higher ice to the east during deglaciation.

The occurrence of meltwater channels leading into Methow tributaries across divides from the northwest, north, northeast, and east indicates that even during the earliest stages of deglaciation voluminous meltwater drained <u>into</u> the Methow drainage basin. Which is to say: during deglaciation the surface of the ice sheet was lower in the Methow region than in adjacent drainage basins. That ice remained higher to the northwest and north, whence most of the Cordilleran ice entered the Methow drainage basin, is not surprising. During deglaciation ice might have remained higher over the Okanogan Valley because downwasting rendered ice flow over the high Methow passes less efficient than through the comparatively low and open upper Okanogan Valley. For the same reason, as suggested on p.   , ice seems to have arrived at the Methow mouth via the Okanogan Valley earlier than via the upper Methow passes. Thus, ice-sheet glaciation of the Okanogan Valley appears to have commenced earlier, and persisted somewhat later, than in the adjacent Methow Valley.

Deglacial meltwater flowing along the Okanogan Valley and along the east side of the Methow Valley must have crossed the Methow glacier tongue at the Methow Valley mouth; because at least during the later stages of deglaciation the chief ice-marginal channels along the Columbia Valley were Alta Coulee (Fig. 24) and Antoine Coulee (Waters, 1933).

Channels between adjacent tributaries within Methow Drainage Basin. Sets of coulees occur between almost every pair of tributaries in the lower Chewack-Methow region; some coulees (Pipestone Canyon, lower Finlay Canyon) that were part of the Alpine-Phase networks (Fig. 10) were reused during ice-sheet deglaciation (p. 61). Enumeration of all coulees, the implied successions of drainage changes, and the systematic distinction of coulees cut during ice advance from those cut during deglaciation is beyond the scope of this report. Suffice it to describe a small and probably briefly occupied coulee of the type not modified by later ice-sheet abrasion.

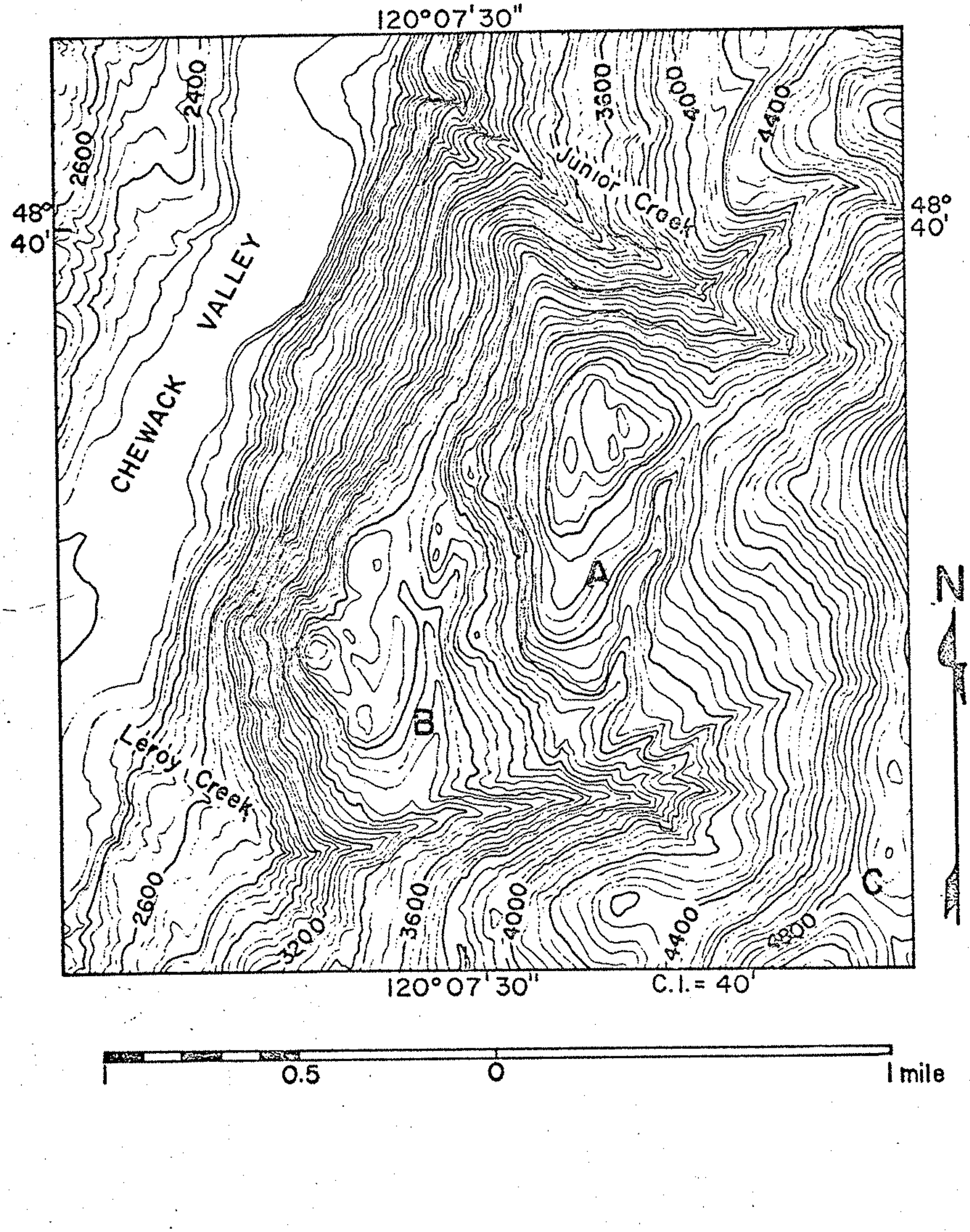
An especially well-formed coulee (Fig. 23) whose intake is 2100 ft above the Chewack trunk valley, descends along a narrow canyon 350 ft deep and 0.75 mi long, cut directly across an interfluve. The coulee, which like all Methow coulees descends from an upvalley tributary to a downvalley tributary, falls 460 ft from its shallow

Figure 23. Topographic map of coulees between Junior and Leroy Creeks.

A. Coulee described in text.

B. Lower less emphatic coulee.

C. Higher coulee leading to Bromas Creek, also a Chewack tributary.


120°07'30"
2400
2600
3600
4000
4400
Junior Creek
48°
40'
48°
40'
CHEWACK VALLEY
A
B
C
N
Leroy Creek
2600
3200
3600
4000
4400
4800
120°07'30"
C.I.= 40'
0.5
0
1 mile

intake on Junior Creek to a small, flat-topped delta in Leroy Creek. The delta and a companion surface 200 ft higher, consist of coarse sandy gravel; they indicate that the high-velocity cataract emptied into a small ice-dammed lake in Leroy Creek. Insofar as the volume of the two deltas approximately equals the volume of rock removed along the coulee, and because the gravel clasts are entirely local mesocratic gneiss, the coulee intake was probably the outlet of an ice-marginal lake in Junior Creek. The upper delta in Leroy Creek represents an earlier surface of the Leroy Lake which dropped suddenly to the altitude of the lower surface when a part of the ice dam collapsed.

The fresh coulees clearly related to deglaciation are a great variety of length, width, depth, and straightness. Multifurious topographic effects were superimposed on those coulees later modified by ice-sheet glaciation. Though many low-altitude coulees are associated with kame terraces (Fig. 27), most higher coulees lack deltas at their lower ends. Some coulee intakes are pronounced and deep; others are inconspicuous flat segments of the interfluve; a few are spectacular, complex, former waterfalls. Thus, the Junior-Leroy coulee above is intended only as an example, from which there is much deviation in form among the thousands of similar channels elsewhere in the Methow region.

CATASTROPHIC FAILURE OF ICE-MARGINAL LAKES

Evidence from Beaver Creek. There is little evidence to indicate the process by which the many ice-marginal lakes were emptied. One supposes that they were mostly unspectacular lowerings as new outlets were uncovered by the downwasting ice dam. On the surface of the alluvial fan at the mouth of Beaver Creek, however, an anomalous concentration of 10-ft boulders represents an unusually high discharge down Beaver Creek. The boulder field, though far enough from steep slopes to preclude an origin by mass-wasting, is near the apex of the fan where the comparatively confined valley widens abruptly, exactly where a flood down Beaver Valley would have experienced an abrupt decrease of competence. Excepting failure of a landslide-dammed lake, for which there is no evidence, the only reasonable source of flood water could have been an ice-dammed lake higher in Beaver Valley or an ice-dammed lake that drained into Beaver Valley via Pipestone Canyon. Of the conspicuous ice-marginal coulee system along the east side of the Methow Valley (Fig. 10), Pipestone Canyon displays the best evidence of having been intensely modified, or excavated entirely, by late-glacial meltwater. Ryason (1959, p. 6) suggested that Pipestone Canyon was the outlet of a sizeable lake dammed by ice east of Winthrop; I would concur, adding that because the outlet was positioned in weakly consolidated Paleocene sand and gravel, the stream was capable of downcutting and headward erosion so rapid as to promote, finally, catastrophic failure and discharge of flood proportions. The large deltas or alluvial fans that were built against stagnant ice at the mouths of Beaver Creek and Benson Creek, are downvalley from the only two large Methow coulees that have been obviously affected by marginal streams during deglaciation.

A smaller field of boulders that are 3 to 9 ft in diameter 2 mi above the town of Methow records a similar high discharge down the lower Methow Valley, possibly coincident with the high discharge from lower Beaver Creek.

Lower Methow and Columbia Valleys. Erratic "haystack" boulders of Columbia River Basalt at altitudes of 2200 to 2600 ft, 3 mi northwest of Pateros led Barksdale (1941a, p. 734) to the conclusion that Cordilleran ice of eastern origin blocked the mouth of the Methow Valley at a very late stage of glaciation. The basalt erratics, however, are not associated with till of eastern provenance; they rest directly on till containing striated pebbles from the Virginian Ridge, Winthrop, and Ventura formations, and other lithologies indicating ice-sheet glaciation from the northwest. The boulders occur, peculiarly, only on low-relief surfaces 1200 to 1600 ft above the floor of the Methow Valley. Moreover, the regional evidence of deglaciation, whether from the Columbia Plateau (Flint, 1937; Hanson, 1970), the southern Okanogan Valley (Flint, 1935), or the lower Methow region indicates downwasting and stagnation of the ice sheet, especially at the lower altitudes. No known evidence in these regions indicates significant renewal of actively flowing Cordilleran ice after deglaciation had commenced.

An alternate explanation for the boulders is that they were rafted into the lowermost Methow Valley by icebergs carried on the crest of the Missoula Flood that, contrary to popular belief, discharged down the northwest segment of the Columbia as well as through the Grand Coulee and the Channeled Scablands (Waitt, 1972). Evidence for the flood in the Pateros segment of the Columbia includes the field of enormous boulders up to 500 ft above the Columbia 0.5-1.5 mi south of the Methow confluence and an associated pebble gravel in which giant ripple crests trend perpendicular to the Columbia Valley. A late-glacial readvance of the ice-sheet from the east is a hypothesis clearly in conflict with the general conditions of deglaciation as outlined above; the flood hypothesis for the origin of the boulders not only reconciles this deductive consideration, but accounts for the narrow range of altitudes of the boulders and for the lack of any associated till of eastern provenance. Because the projected profile of the Missoula Flood crest is above the highest "haystack" boulders near

Chelan, the use of these erratics as evidence of glaciation of lower Chelan trough by the Okanogan Lobe, is incorrect.

LOCAL SUPERPOSITION OF STREAM SEGMENT

Winthrop Area. In contrast to the typically wide alluvial floors of the Chewack and upper Methow valleys, their meander belts through the Winthrop kame-moraine (p. 134 and Fig. 29) are narrow and the channels floored with bedrock.

The Winthrop kame-moraine is protected from river erosion by bedrock along its northern and eastern sides, but a distinct reentrant at its northwest end,directly facing the upper Methow Valley, reveals a buried valley that can be followed southeast through a string of kettles along the base of Patterson Mountain (Fig. 29). The floor of the largest kettle, Twin Lakes, is nearly the altitude of the floor of the modern Methow Valley. A less distinct reentrant marks the intersection of the lower end of the buried valley with the modern valley, which at that point widens abruptly to 1 mi, the width of the upper Methow Valley. The reentrants resulted as Methow meanders found the loose drift of the kame-moraine more erodable than the bedrock along the rest of the valley. Both ends of the drift-filled former course of the Chewack are similarly disclosed by reentrants: the buried valley trends south-southwest from the Chewack Valley 1 mi north of Winthrop, to the Methow Valley 0.5 mi northwest of Winthrop.

There is no ready explanation for the drainage changes near Winthrop except that since they coincide with the largest morainal feature in the region, they are related to deglaciation. Possibly, stagnant valley ice lingered last in the shade of Patterson Mountain, thereby relegating the voluminous ice-marginal drainage to the eastern side of the valley. As the ice disappeared, the streams became superposed through glaciofluvial drift into their anomalous courses that, because the bedrock is Newby Group shale previously reduced to low spurs by stream and glacial erosion, are less spectacular gorges than they might otherwise be.

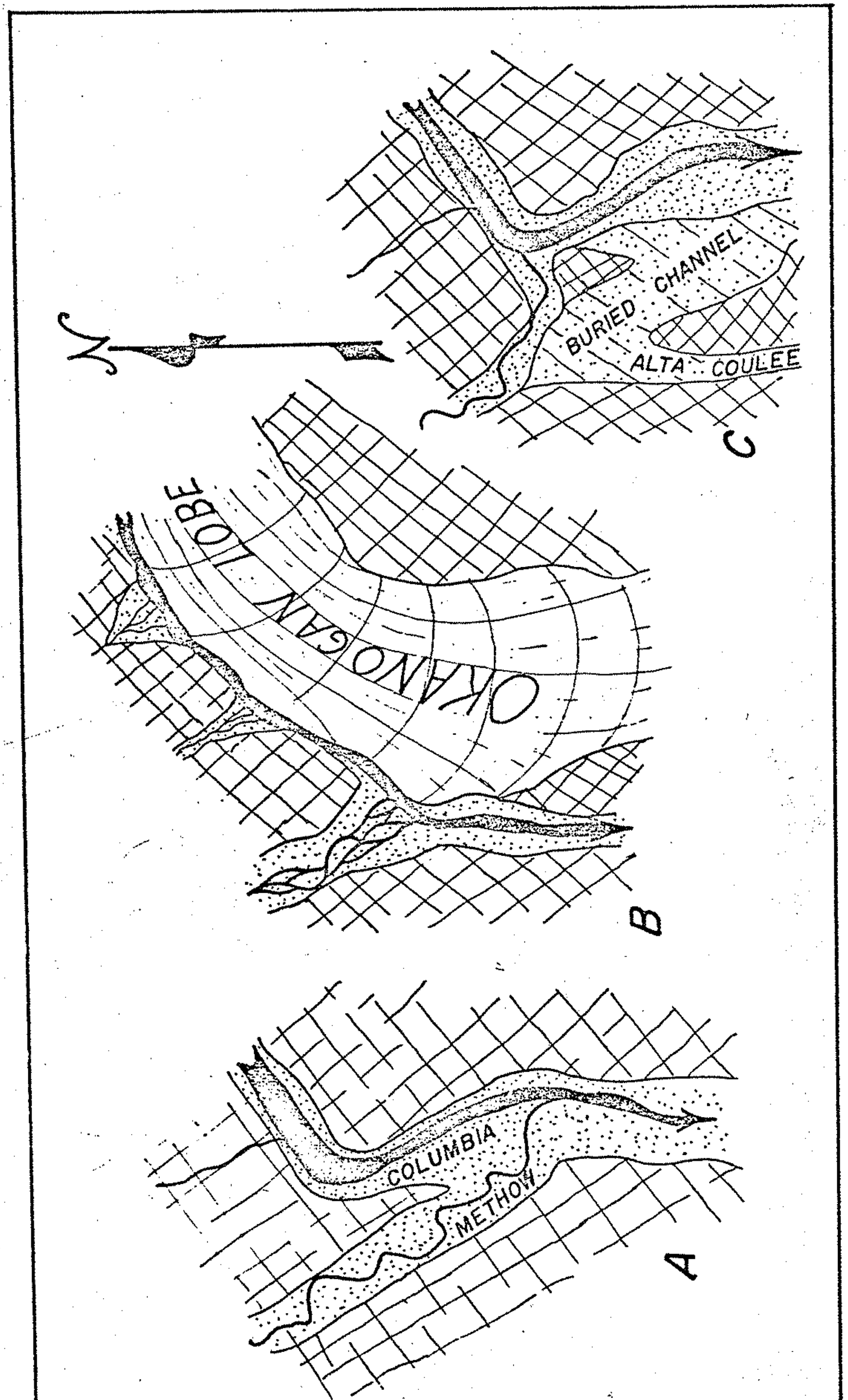
<u>Mouth of Methow Valley</u>. The Methow Valley at the Columbia confluence is a comparatively narrow gorge through which the river flows northeast, toward the <u>upstream</u> segment of the Columbia. Meander-cut reentrants into unconsolidated drift of the Great Terrace, however, mark the upper and lower ends of a buried valley segment traceable through a line of shallow kettles to the Columbia Valley 3 mi south of the present Methow confluence. Apparently the lowermost mile of the Methow Valley was superposed into its anomalous course through a former blanket of drift (Fig. 24).

Figure 24. Drainage derangement at mouth of Methow Valley.

A Preglacial pattern.

B Ice-marginal drainage pattern during deglaciation.

C Postglacial pattern, showing superposed segment of lowermost Methow River.



N
A
COLUMBIA
METHOW
B
OKANOGAN LOBE
C
BURIED CHANNEL
ALTA COULEE

<u>Chewack River Eightmile Creek, Libby Creek</u>. Opposite the mouth of Boulder Creek a 0.2-mi segment of the Chewack River flows across bedrock; for 15 mi upvalley and 7 mi downvalley the Chewack is a broad, flat valley floored with alluvium. The bedrock-floored segment is clearly anomalous, as is a small bedrock knoll stranded in mid-valley. The Chewack at this point is crowded against the western valley wall, apparently because of a late-glacial alluvial fan built across the valley by Boulder Creek. During incision of the fan instead of returning to its original valley, which remained obstructed with remnants of the fan, the Chewack became superposed in its bedrock canyon across a buried bedrock spur.

Several other examples of local superposition of streams through late-glacial drift produce conspicuously anomalous bedrock segments of streams that otherwise flow along broader, aluviated valley floors. Among them is a segment of Eightmile Creek 1.5 mi upvalley of its Chewack confluence, and Libby Creek near the Smith Canyon confluence. In each case the natural course of the valley is choked with a partly dissected remnant either of glacio-fluvial drift or of a tributary fan built in response to a base level held temporarily high because of residual ice in the trunk valley.

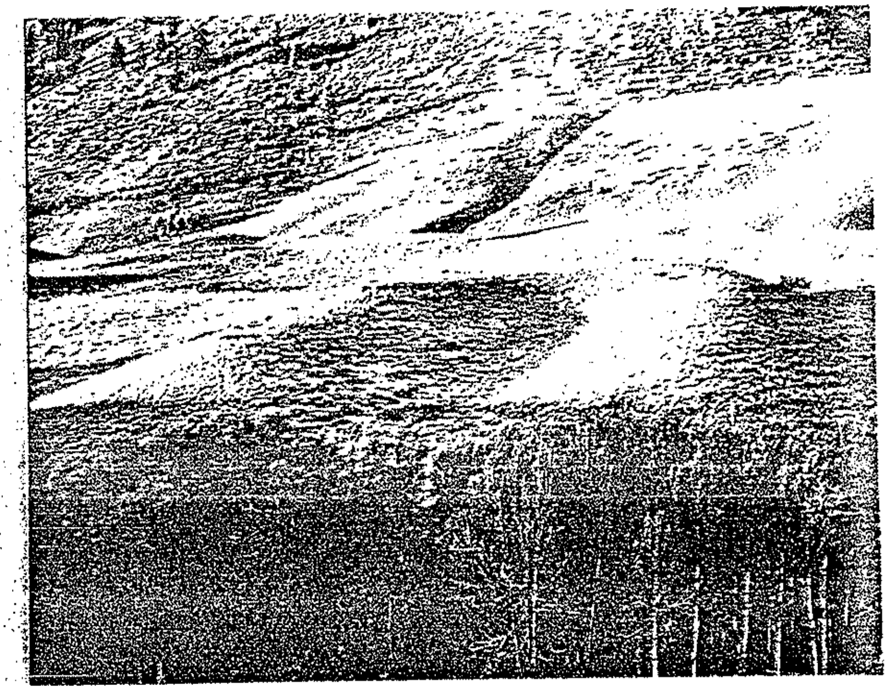
UPLAND ICE-DISINTEGRATION LANDFORMS

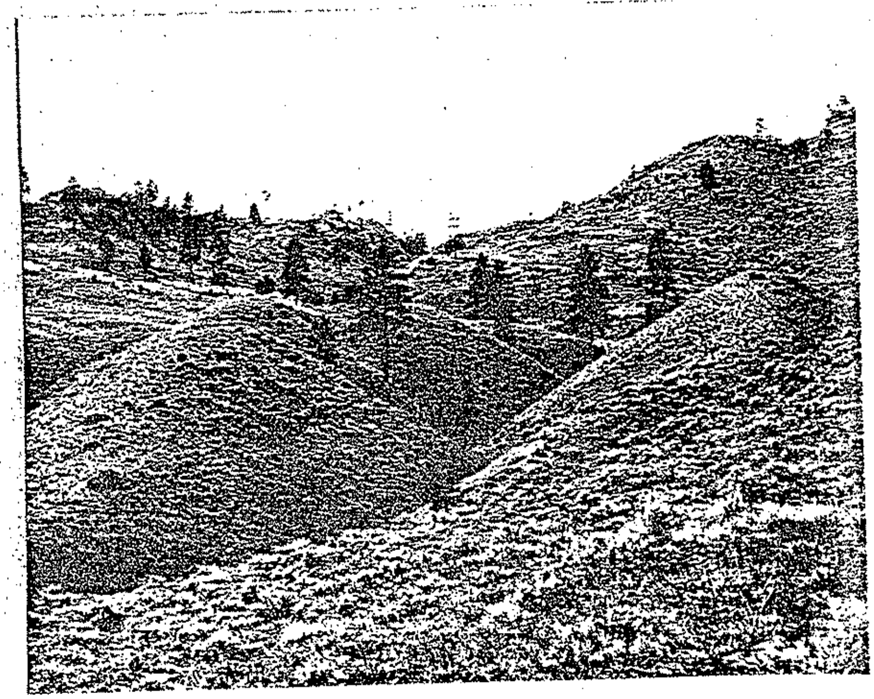
Although the discontinuous mantle of upland drift mainly reflects the topography of the buried bedrock surface, small areas of drift having an irregular hummocky surface are found as high as 2000 ft above the floor of the lower Methow Valley. Pearrygin Lake, for example, although mainly an ice-scoured depression cut in erodible Newby Group shales, is partly dammed on its southern end by hummocky drift at least 50 ft thick. This deposit of loose gravelley sand, which displays kettles, small kames, and a southeast-trending esker, clearly overlies postdates the regional till sheet that is distinguished by its striated cobbles and large boulders. Similar deposits partly obstruct the floors of Elbow Coulee and other large ice-marginal channels (see p. 61).

Three miles southeast of Pearrygin Lake a flat-topped, kettled deposit with a local relief of 60 ft partly obstructs a high-level ice-marginal channel at its merger with Pipestone Canyon. Within Pipestone Canyon and 80 ft above Campbell Lake a horseshoe-shaped ridge of muddy gravelly sand (Fig. 25), a maverick in the Methow region, resembles doughnut-shaped ridges characteristic of stagnant-ice topography in the northern Great Plains (Gravenor and Kupsch, 1959; Parizek, 1969). The sandy drift that anomalously plugs the tributary valley shown in Figure 26 could only have been deposited

Figure 25. View north of horseshoe-shaped drift ridge and higher-level terrace above Campbell Lake in Pipestone Canyon. The drift ridge, similar to doughnut-shaped ice-disintegration ridges in the High Plains, denotes stagnation of the Cordilleran Ice Sheet during deglaciation.

Figure 26. Mound of sandy drift in floor of V-shaped tributary of the lower Methow River. The anomalous position of the mound and its lateral channels defy an explanation by normal stream or active ice processes. The drift was probably washed against and over a margin of stagnant at left of picture; when the last residual ice along the axis of the valley melted, the small tributary abandoned the lateral channels and returned to its normal valley-fllor position.





against stagnant ice. These and many similar ice-contact deposits and landforms south of the latitude of Winthrop indicate that at least the residual lower 2000 ft of the ice sheet disappeared largely by downwasting and stagnation.

MORAINES

Chewack, upper Twisp, upper Eightmile, and upper Methow Valleys. The crest of a conspicuous bouldery lateral moraine 300 ft above the floor of upper Eightmile Valley descends to the valley floor 1.5 mi to the southeast. An ice-marginal channel slopes downvalley between the moraine and the valley wall. On the south flank of Burch Mountain and about 2300 ft above the floor of Eightmile Valley three inconspicuous lateral moraines 5 to 10 ft high and spaced 40 to 60 ft above one another, descend 0.25 mi downvalley at a similar gradient.

Barksdale (1941a, p. 735) tentative correlated a conspicuous "moraine" against the western side of the Chewack Valley 2 to 3 mi south of the Twentymile Creek concluence with the Winthrop "moraine". However, the Chewack "moraine" is mainly a flat, beveled-off bedrock spur that slopes gently south from a maximum height of 350 ft above the valley floor. The spur is mostly mantled with fluvial gravel, but along its valleyward edge a ridge of till rises as much as 100 ft higher. The crest of the ridge, which contains boulders up to 15 ft in diameter (in contrast to 4-ft maximum diameters among the fluvial boulders) is indeed a small lateral moraine. Rising 120 ft above the valley floor 1.5 mi upstream and across from Twentymile Creek, a similar ridge consisting entirely of sandy fluvial gravel rises 30 ft above its bedrock foundation.

Three lateral moraines descend fro 2500 ft along the eastern side of Lake Creek near the Chewack confluence. The ridgecrests, copiously strewn with boulders up to 12 ft in diameter, are spaced about 100 ft altitude apart and have less than 50 ft of relief with respect to conspicuous ice-marginal channels between them and the valley wall. The lowest moraine is probably an extension of an inconspicuous bouldery ridge along the valleyward edge of a broad terrace along Lake Creek, 2 mi farther north.

Near the great elbow of Twisp Valley the outer edge of a conspicuous bench along the southwest valley wall displays a ridge of sandy bouldery till that, like its Chewack analogue, masquerades as a formidable moraine; it is but an incidental bump atop a bedrock platform 240 ft above the valley floor. At the North Creek confluence a small irregular valley-floor moraine with a lateral meltwater channel may record a brief stillstand of a remnant of active glacier ice in the head of Twisp Valley.

Ridges and closed depressions having a maximum relief of 40 ft obstruct Cutthroat Creek Valley for half a mile below Cutthroat Lake. The rude upvalley concavity of the ridges suggests an origin as looped recessional moraines; their concavity toward, and their positions close to, the northern valley wall suggests otherwise. The anomalous topography, incomparable to any other depositional landform in the Methow drainage basin, probably resulted from mass-wasting, perhaps from avalanche.

The small lateral moraines of bouldery till at low altitudes in the upper Chewack, upper Eightmile, and upper Twisp valleys suggest the possibility of a few very brief and locally significant recessional pauses when tongues of the ice sheet had retreated to the upper parts of the trunk valleys. These moraines are restricted to the localities cited; I have seen no comparable moraines elsewhere in these valleys nor in the upper Methow or its larger tributaries.

Small ridges of sandy till containing boulders 8 to 10 ft in diameter occur at several such places in the Okanogan Mountains as Roger Lake and Starvation Mountain (p. 97-98). However, these and dozens of less conspicuous ridges associated with meltwater channels or glacoifluvial terraces adjacent to the lower Methow Valley are rare, widely spaced, and distributed through at least 4000 ft of altitude. As indicators of pauses during the wasting of the Cordilleran Ice Sheet, these landforms are insignificant. I have not attempted to relate them to one another or to moraines in other areas.

Lower Methow Valley. Although Barksdale (1941a, p. 732-734 and Fig. 6) cited a morainal origin of several drift bodies in the lower Methow Valley as far south as Carleton, those landforms are flat-topped, are composed of stratified gravelly sand, and are located at tributary junctions: they are terraces. Furthermore, there are many more such features than those singled out by Barksdale. They occur, moreover, downvalley to the Columbia confluence and beyond.

The terrace sediment, up to 500 ft thick, is sandy gravel and gravelly sand deposited on bedrock surfaces. The terraces are clearly fill terraces or, in the case of some of the lower benches, erosional terraces cut in alluvium. The abundance of associated kettles, occasional associated eskers, and the intimate association of ice-marginal channels across spurs (Fig. 27) indicate that most of the terraces were built against stagnant valley ice: they are kame terraces. Discontinuous remants of kame terraces occur in the Methow Valley from the Columbia confluence to Winthrop. Virtually every tributary to the lower Methow contains an anomalous sedimentary fill whose surface grades downvalley to a kame-terrace surface in the main valley. Although the most prominent and most abundant terraces are within 500 ft of the floor of the Methow Valley, some occur more than 1000 ft above the floors of the Methow, Twisp, and Chewack Valleys.

Figure 27. View southeast along spur transected by numerous ice-marginal channels that are parallel to the trend of lower Methow Valley (middle ground). Note intimate association of ice-marginal channels with the great fill terrace of the Methow.

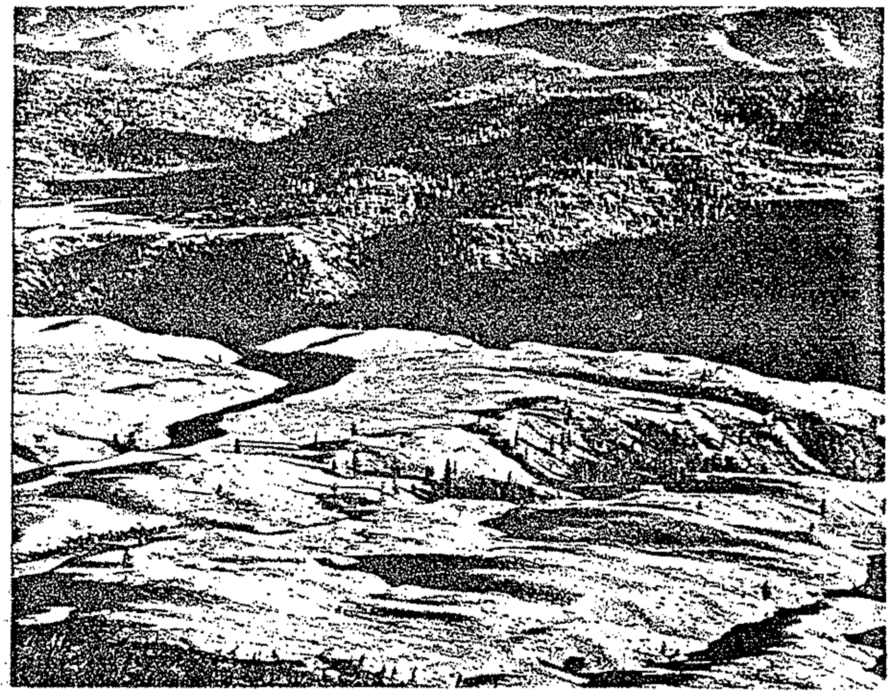
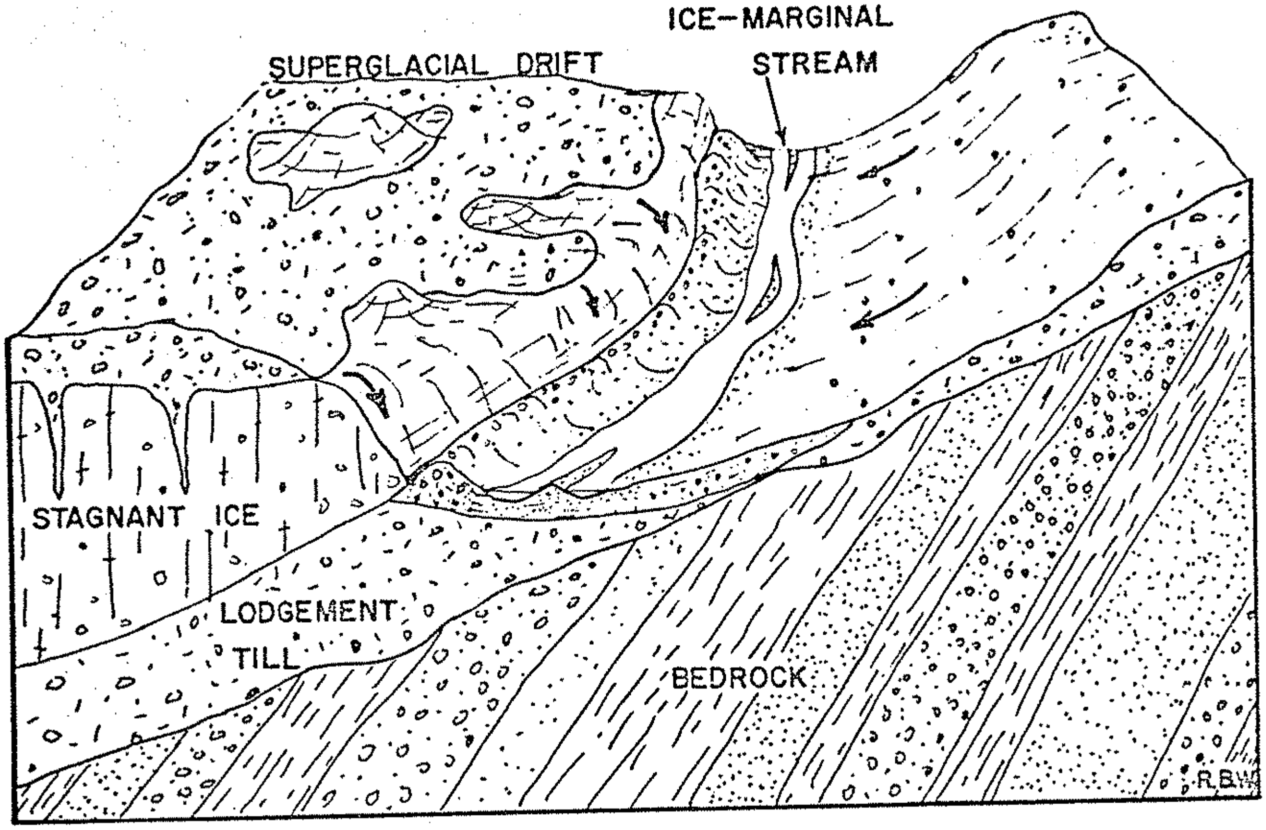


Figure 28. Origin of terrace-front drift ridges in lower Methow region.

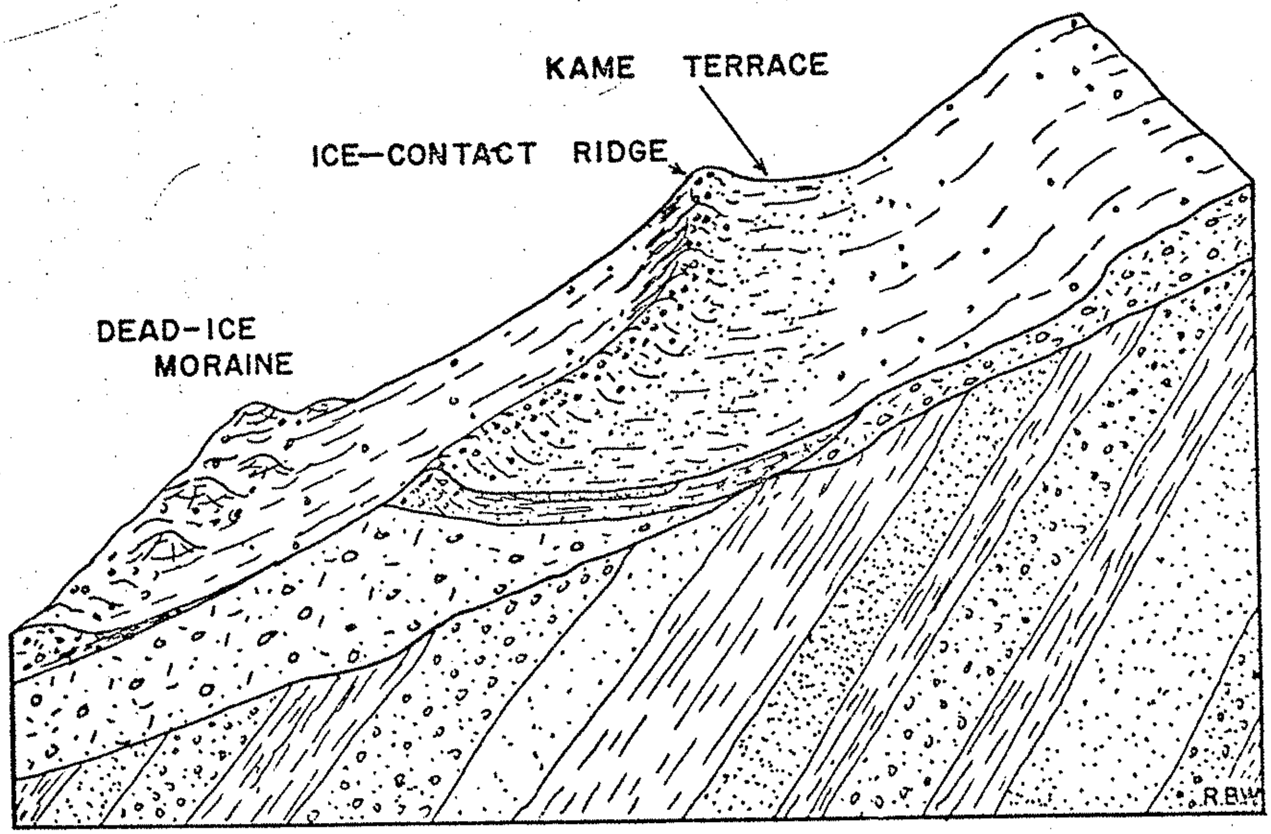
A. Superglacial drift is shed from steep, unstable margin of stagnant ice into marginal gutter.

B. Ice-contact-drift ridge is positioned at valleyward edge of former ice-marginal stream surface.

A ridge of bouldery drift 10 to 50 ft high resembles a lateral moraine along the edge of some terraces. But a prominent ridge along the south side of Benson Creek Valley is not unique as implied by Barksdale; others occur elsewhere in Benson Creek, McFarland Creek, Squaw Creek, and at several places along the Methow Valley south of Twisp. Some, like the Benson Creek ridge and a ridge on the floor of the Methow Valley near the Benson Creek confluence, are concave towards, and enclose depressions against, the valley wall. The ridges, almost invariably associated with kame terraces, are similar in form and stratigraphic association to landforms described by Clayton (1967) and figured in detail by Parizek (1969) in areas of stagnant-ice topography on the Dakota and Saskachewan high plains. Figure 28 is a diagramatic interpretation of their origin: poorly sorted bouldery ablation debris sliding off stagnant ice comes to rest in the moat between the ice margin and the valley wall, where marginal streams had previously deposited moderately well-sorted terrace-forming sand and gravel. In the lower Methow Valley numerous ridges that Barksdale interpreted as moraines are now believed to have originated in this manner and provide evidence of widespread stagnant ice rather than active tongues of valley glaciers.


ICE-MARGINAL
STREAM
SUPERGLACIAL DRIFT
STAGNANT ICE
LODGEMENT
TILL
BEDROCK
R.B.W.

A

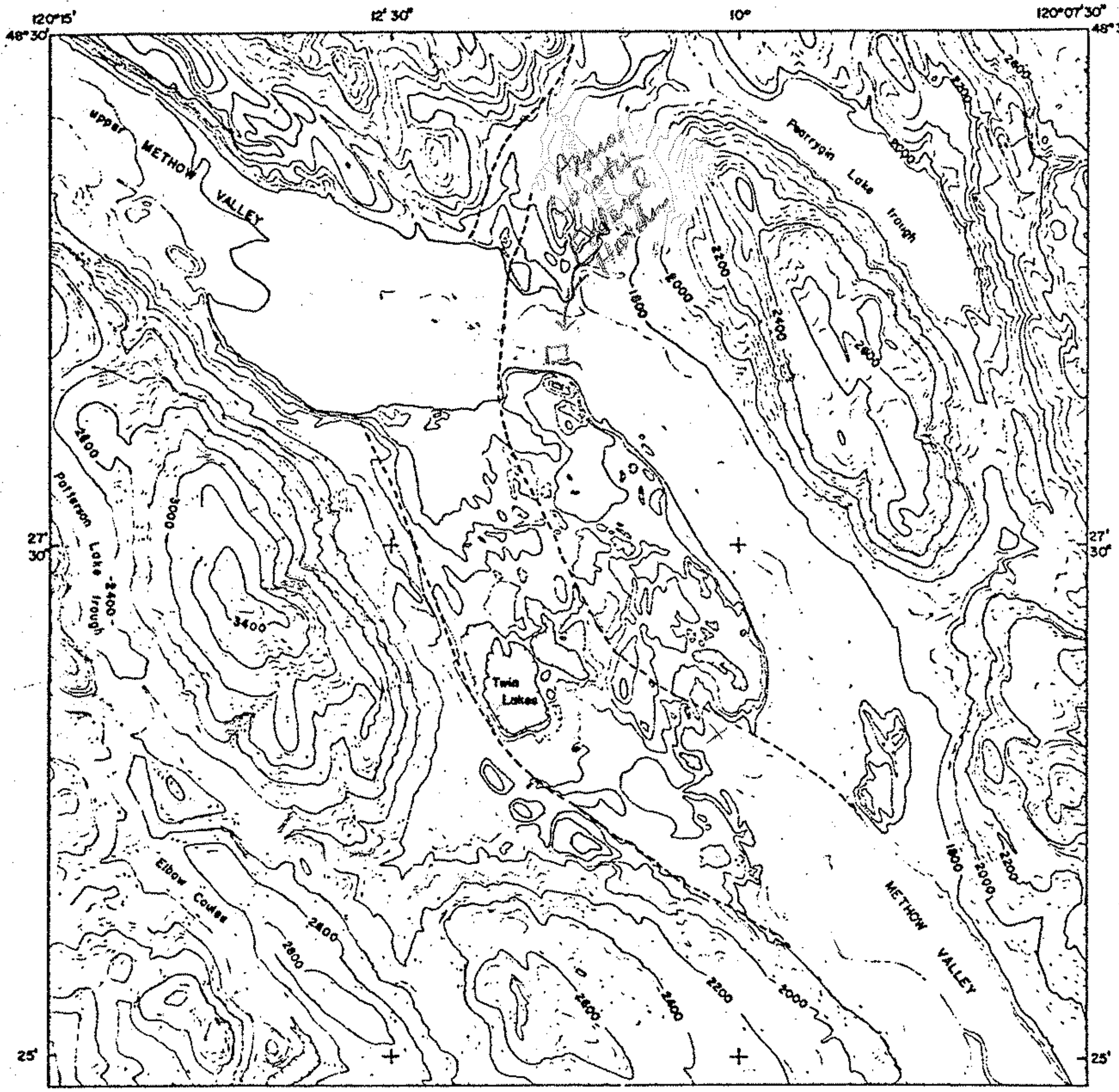

KAME TERRACE
ICE-CONTACT RIDGE
DEAD-ICE
MORAINE
R.B.W.

B

Winthrop area. South of the Methow-Chewack confluence a $4\ mi^2$ deposit of sandy gravel, whose flat-topped surface lies mostly at altitudes of 1840 to 1910 ft, rises abruptly 100 to 200 ft above the floor of the Methow Valley (Fig. 29). Bodies of drift to the northwest, north, and east indicate a formerly greater extent of the feature. Although inferred by Russell (1899, Plate 18) to be the terminal moraine of the Methow Glacier and by Barksdale (1941a, p. 732-734 and Fig. 6) as the most important of a series of recessional moraines, the Winthrop surface is interrupted by kettles 100 to 2000 ft wide and is crossed by dozens of eskers 10 to 50 ft high trending south-southeast, parallel to the modern valley. Except for a knoll rising 120 ft above the north corner, which is possibly cored with bedrock and which exhibits striated erratics as much as 15 ft in diameter, the Winthrop feature is a kame-moraine of glaciofluvial drift having stagnant-ice morphology. As explained on pages 121-122, the local superposition of the Methow and Chewack rivers into bedrock-floored channels through the Winthrop has protected the kame-moraine from postglacial erosion.

GREAT TERRACE OF THE METHOW VALLEY

In the Winthrop area of the Methow Valley the kame-moraine surface stands out far more boldly than any other terrace; along most of the Methow Valley south of Winthrop



TOPOGRAPHIC DETAIL OF THE WINTHROP AREA

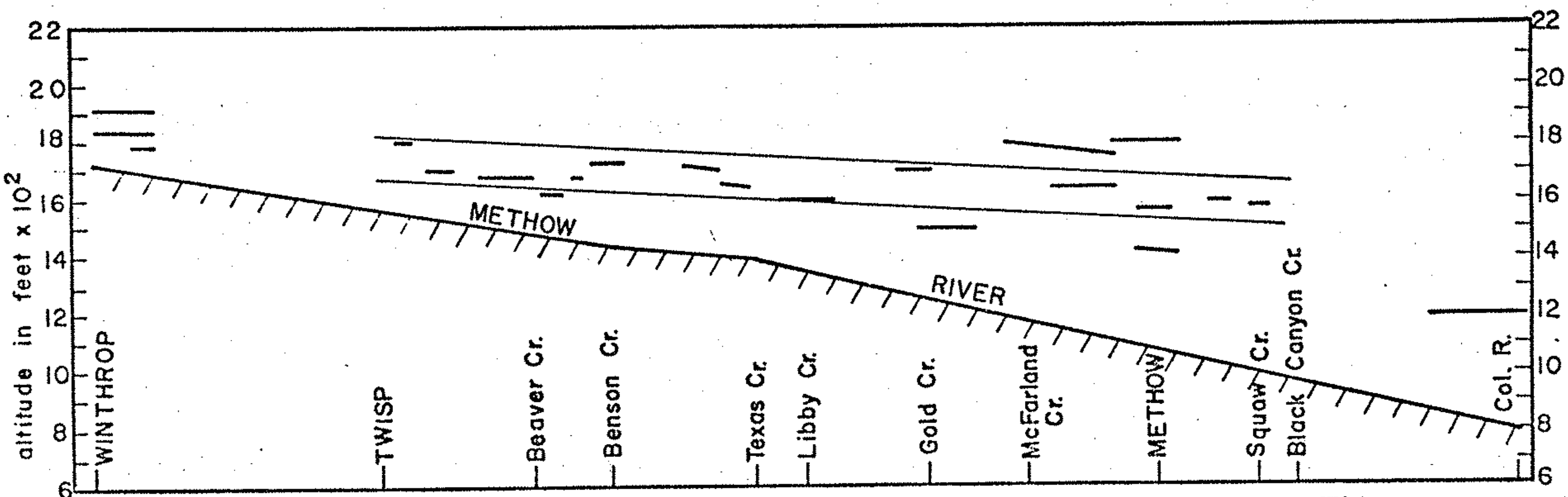
1 0.5 0 1 mile

CONTOUR INTERVAL 40 feet

one terrace is distinctly more prominent than all others. However, the terrace has neither a smooth slope nor a constant altitude above the valley floor (Fig. 30). Rather, it ranges from 100 ft on parts of the Winthrop kame-moraine to over 700 ft near Methow. Clearly, the terrace is not a single feature, but consists of several broad surfaces that formed at somewhat different times. But because most of the conspicuous terrace remnants occur in a narrow range of altitudes some 150 to 500 ft above the Methow River, they indicate a unique period of base-level stability and stream alluviation. Between Winthrop and the Squaw Creek confluence the mean gradient of the 200-ft interval enveloping most of the terrace remnants is 8 ft/mi; the modern Methow River is three times as steep through the same segment.

The great terrace in the lowermost 5 mi of the Methow Valley is a continuation of the Great Terrace of the Columbia, whose surface (1210 ft) is more than 400 ft above the valley floor. Although the Columbia Great Terrace consists mainly of finely laminated silt (Flint, 1935, p. 85; Russell, 1898, p. 365; Waters, 1933, p. 790), the Methow counterpart comprises sand and gravel. Like the Columbia Great Terrace, the Methow great terrace is locally pitted with kettles and is intimately associated with ice-marginal channels, indicating its origin as a kame terrace.

Figure 30. Distribution and altitudes of the Methow great terrace complex, whose variable altitude such as near Methow indicates development during some interval of time. The mean slope of the envelope around most of the terrace remnants is 8 ft/mi.



MAJOR KAME TERRACES ALONG METHOW VALLEY

Through a particularly narrow, steep segment of the Methow Valley between the upvalley and of the Great Terrace branch and the mouth of Squaw Creek, prominent terrace remnants are lacking. From Squaw Creek upvalley to the Twisp confluence, however, a single prominent terrace is conspicuous alongthe Methow Valley and includes such enormous surfaces as the delta-tops at the mouths of Benson and Beaver Creeks.

A single conspicuous fill terrace in the Methow Valley above Winthrop is much less continuous than below Winthrop. Remnants of this terrace occur along a 3-mi segment south of Goat Creek 400 ft above the valley floor, in lower Early Winters Creek 400 ft above the valley floor; and in the Methow West Fork upvalley from Rattle-snake Creek 250 ft above the valley floor. Conspicuous fill terraces rise 100 to 200 ft above the floors of the lower Twisp and lower Chewack Valleys. Although none have distinctive ice-contact features, they are clearly fill terraces or terraces cut in alluvium; that they are graded to the great kame terrace of the Methow suggests that they are related to the same general interval of valley-side aggradation.

Most tributaries to the Methow, Chewack, and Twisp Valleys, have gently sloping alluvial surfaces approximately graded to the great terrace levels in the trunk valleys. Incision of the lower several miles of these surfaces by modern streams reveals that the tributary surfaces, like the mainstream terraces, are underlain by fill of sandy gravel many tens of ft thick.

Thus, terraces in the Methow, Twisp, and Chewack valleys and in most smaller tributaries denote a single episode of alluviation of regional significance that, as the ice-contact origin of the great terrace complex indicates, was caused by a high regional base level was maintained by residual Cordilleran ice in the Methow Valley. Although the Methow great terrace complex developed over a time interval of unknown duration, the surface of the whole aggradational system may be regarded as nearly contemporaneous. This late-glacial surface can thus be followed from the Great Terrace along the Columbia Valley northwest well into the northeastern Cascade Range.

DISCUSSION OF ICE-SHEET DEGLACIATION

Except for small lateral moraines in parts of the upper Twisp, Eightmile, and Chewack Valleys, the Methow drainage basin is devoid of the looped end moraines built by actively retreating ice-tongues in Cascade farther south (see, e.g., Hopkins, 1966; Long, 1951; Porter, 1969). North of the Methow-Columbia confluence the only valley moraine having trends of a traditional arcuate pattern lies athwart the valley of the Psayten West Fork 9 mi north of Harts Pass, where morainal loops, concave downvalley and associated with kame-and-kettle topography, were therefore also built by ice that receded northward by stagnation. The striking absence of such moraines, the fact that every Methow moraine south of Winthrop is characterized by dead-ice topography, indicates that final deglaciation of the region was accomplished not by backwasting of active ice but predominently by downwasting of stagnant ice.

Although Barksdale (1941a, p. 735) indicated that "moraines" in the Methow Valley have progressively less mature soils in the upvalley direction, restudy of these features disclosed no such systematic trend. The depth of oxidation of the drift is variable, ranging from 2 to 15 ft, but the mean depth is close to 4 ft and is fairly consistent throughout the drainage basin, indicating that surface sediments are approximately of uniform age. Differences in depth of oxidation probably reflect variations in precipitation and runoff, in permeability, and in parent material. Although these bodies of drift relate to a single deglacial interval, they occur at various altitudes between 30 and 500 ft above the valley floors. Therefore, they must be of slightly different ages.

Deductively, there are two reasons for supposing that drift bodies of comparable altitude are slightly younger upvalley. First, because the ice sheet was wedge-shaped and thinner southward, a uniform rate of downwasting would have caused deglaciation of the downvalley regions earlier. Second, if the Late Wisconsin climate downvalley was, like the modern climate, both warmer and more arid, the rate of downwasting would have been more rapid to the southeast, thereby imparting an additional component of backwasting to the dominant process of downwasting.

Probably no ice tongue wastes away entirely either by downwasting or by backwasting. Backwasting dominated during deglaciation of alpine ice in the Cascades south of Chelan trough; downwasting dominated deglaciation of the Cordilleran Ice Sheet on the Columbia Plateau, and in the Okanogan, Methow, and Pasayten valleys.

LACK OF LATE-GLACIAL ALPINE PHASE

Granodiorite erratics of Pasayten-stock provenance on the floor of the north-facing cirque at the head of the Pasayten West Fork indicate that the last active ice occupying Pasayten Valley flowed south (upvalley), clearly a manifestation of ice-sheet glaciation. Moreover, striations are preserved not only on the beveled-off ridges surrounding the cirque, but on the cirque headwall where they very likely would not have endured a late-glacial rebirth of cirque glaciers. Furthermore, renewed alpine glaciation would have produced a rugged headwall, would have removed the cirque-floor erratics, and very likely would have produced looped moraines downvalley. Clearly, the most recent glaciation of this valley was by the Cordilleran Ice Sheet. Because the West Fork cirque currently harbors a large permanent snowfield, it should have been one of the first cirques in the North Cascades to have experienced renewed alpine glaciation, had the climate been favorable. Pasayten-stock erratics on the floors of cirques in the heads of the Slate Creek, Trout Creek North Fork, and Robinson Creek indicate that nearby valleys similarly remained free of late-glacial alpine ice. Striations midway up the headwall of a cirque at the head of South Creek, an upper Twisp tributary, indicate that the Methow-Chelan divide, like the Methow-Pasayten-Skagit divide, lacked a late-glacial alpine phase. That two areas 20 mi apart exhibit parallel late-glacial records suggests that after disappearance of the Cordilleran Ice Sheet the entire Methow region experienced no renewed valley glaciation. The absence of valley moraines in the Methow, Pasayten, Skagit, and Chelan drainage basins is complementary evidence that the Cordilleran Ice Sheet postdated the most recent significant alpine phase. Except for small glaciers confined to the highest cirques during the Neoglacial interval, there has been no regeneration of alpine ice.

COMPARISON TO NEW ENGLAND SECTOR OF LAURENTIDE ICE SHEET

At about the same time as the Cordilleran Ice Sheet occupied the northern Cascades (Schafer and Hartshorn, 1965, p. 119), the New England sector of the Laurentide Ice Sheet overtopped uplands that had been previously occupied by local alpine glaciers. Striated cirque headwalls, erratics of northwestern provenance positioned on cirque floors as well as on upland surfaces, and an absence of valley-glacier moraines indicate that alpine glaciers in the Presidential Rnage, like alpine glaciers in the northern Cascades, were not reborn following ice-sheet deglaciation (Goldthwait, 1968). The abundance of stratified drift and dead-ice topography southeast of the upland provance indicates, moreover, that the terminal 100 mi of the Laurentide Ice Sheet in New England, like the Cordilleran Ice Sheet southeast of the highest Cascades, disappeared mainly by downwasting and stagnation.

Thus, parallel modes and sequence of deglaciation occured <u>simultaneously</u> on opposite sides of the continent, areas affected by two physically independent ice sheets--ice sheets of unequal volume and nourished by different sources of moisture. Therefore, climate change, not peculiarities of glacier dynamics, was the cause of glacier fluctuations and of the somewhat puzzling deglacial sequence. An abrupt, popular, and final amelioration of climate affected the 40°-50° latitudes of the North American continent about 14,000 years B. P.

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VITA

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