Combined Ice and Water Balances of Gulkana and Wolverine Glaciers, Alaska, and South Cascade Glacier, Washington, 1965 and 1966 Hydrologic Years

By MARK F. MEIER, WENDELL V. TANGBORN, LAWRENCE R. MAYO, *and* AUSTIN POST

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COMBINED ICE AND WATER BALANCES OF GULKANA AND WOLVERINE GLACIERS, ALASKA, AND SOUTH CASCADE GLACIER, WASHINGTON, 1965 AND 1966 HYDROLOGIC YEARS

By MARK F. MEIER, WENDELL V. TANGBORN, LAWRENCE R. MAYO, and AUSTIN POST

ABSTRACT

Glaciers occur in northwestern North America between lat 37° and 69° N. in two major mountain systems. The Pacific Mountain System, near the west coast, receives large amounts of precipitation, has very mild temperatures, and contains perhaps 90 percent of the glacier ice. The Rocky Mountain or Eastern System, on the other hand, receives nearly an order of magnitude less precipitation, has temperatures that range from subpolar to subtropic, and contains glaciers that are much smaller in both size and total area.

As a contribution to the International Hydrological Decade program on combined balances at selected glaciers, the U.S. Geological Survey is conducting studies of ice and water balance on four glaciers in the Pacific Mountain System: Wolverine and Gulkana Glaciers in Alaska, South Cascade Glacier in Washington, and Maclure Glacier in California. Similar data are being collected by other organizations at five glaciers in western Canada, including two in the Rocky Mountain System, and at one glacier in the Rocky Mountain System in northern Alaska.

Gulkana, Wolverine, South Cascade, and Maclure Glaciers have dissimilar mass balances, and each is fairly representative of the glaciers for its particular region. Gulkana Glacier (lat 63°15′ N., Alaska Range, Alaska) normally has an equilibrium line at an altitude of 1,800 m (meters), an activity index of about 6 mm/m (millimeters per meter), a winter balance of about 1.0 m, and an annual exchange of about 2.2 m. (Balance values are given in terms of water-equivalent measure; the winter balance of 1 m, for example, indicates a volume of ice equal in mass to a volume of water 1 m in depth covering the area of the glacier.) The normal approximate parameters for the other glaciers studied are as follows: Wolverine Glacier (lat 60°24′ N., Kenai Mountains, Alaska)—equilibrium-line altitude 1,200 m, activity index 9 mm/m, winter balance 2.5 m, and annual exchange 5.5 m; South Cascade Glacier (lat 48°22′ N., North Cascades, Wash.)—equilibrium-line altitude 1,900 m, activity index 17 mm/m, winter balance 3.1 m, and annual exchange 6.6 m; and Maclure Glacier (lat 37°45′ N., Sierra Nevada, Calif.)—equilibrium-line altitude 3,600 m, activity index 23 mm/m, winter balance 2.3 m, and annual exchange 4.6 m.

Mass balances of these four glaciers and their drainage basins are measured annually by standard glaciological techniques. In addition, the hydrologic balance is calculated using streamflow and precipitation measurements. Combining these independent measurements results in fairly well defined values of water and ice balance for the glaciers and drainage basins. A revision of the standard International Hydrological Decade mass-balance system permits combination of annual and stratigraphic terms.

The annual balance of South Cascade Glacier at the end of the 1965 hydrologic year was slightly positive (+0.07 m averaged over the glacier), but continued ablation and deficient accumulation in October 1965 resulted in slightly negative net balances for both the glacier and the drainage basin. Factors tending to produce this near-zero balance were the above-average late-winter balance (3.48 m) and the numerous summer snowfalls. Ice ablation averaged about 39 mm of water per day during the main melt season. Runoff during the summer ablation season was lower than the 1958–64 average.

The South Cascade Glacier annual balance in 1966 (−0.94 m) was considerably more negative mainly owing to the deficient winter snowpack (the late-winter balance was only 2.52 m) and the warm dry summer. Ice ablation averaged about 44 mm of water per day during the melt season. The loss in storage of this and other glaciers in the North Cascades increased the runoff of many valley streams by approximately 50 percent during August and September.

The 1966 Gulkana Glacier annual balance was slightly positive (+0.06 m); on the basis of past observations and the rapid terminus retreat of this glacier, this value is considered unusual. Accumulation (late-winter balance) was probably near the average of recent years. The ablation season was quite short, ending over most of the glacier on August 15; this factor alone probably accounts for the slightly positive balance. Ice ablation averaged approximately 26 mm of water per day during the melt season. Runoff from this glacier was not measured in 1966 but is estimated on the basis of ice-balance and precipitation data to be about 2 m, which is 5–10 times the amount measured in nearby lower altitude basins.

In the Kenai Mountains of Alaska, the 1966 Wolverine Glacier annual balance was negative (−0.26 m); the winter balance (1.83 m) was based on precipitation records at nearby Seward and was probably deficient. Ice ablation was heavy during the summer (about 50 mm of water per day average).

Collection of data on ice and water balance for Maclure Glacier began in late 1966 and will be given in subsequent reports.

INTRODUCTION

The International Hydrological Decade (IHD), 1965–74, was launched to promote international cooperation in research and in the training of specialists and technicians in scientific hydrology. Its purpose is to enable all countries to make a fuller assessment of their water resources and a more rational use of them as demands increase from expanding population, industry, and agriculture. The IHD project concerned with the combined heat, ice, and water balances at selected glacier basins marks an important step in broadening the understanding of snow hydrology, high-mountain and glacier hydrology, and the relation of glacier variations to changes in climate.

Snow is difficult to treat using the usual equations of hydrology because snow accumulation and wastage are complex and not sufficiently understood. Little study has gone into the understanding of snow distribution, the meteorological causes of changes in the snowpack and its physical, thermal, and hydraulic properties, and the hydrology of basins where substantial amounts of water are stored seasonally or from year to year in the form of snow. Drainage basins dominated by glaciers are ideal sites for the study of these problems because there the collection, storage, movement, and release of melt water are dominated by the unique properties of snow and ice. This is an example of thermodynamic hydrology in which the heat balance influences the ice balance and, in the process, largely determines the water balance (UNESCO/IASH, 1970).

The specific objective of this IHD project on heat, ice, and water balances is to obtain sufficient information to define these balances and understand how they change with time at glacier basins in widely differing environments in many parts of the world. To extend the coverage from individual drainage basins to global patterns of atmospheric circulation, chains of glacier-basin stations over the world were recommended for the IHD by the International Commission of Snow and Ice of the International Association of Scientific Hydrology (IASH). One chain of stations extends along the western mountains of the Americas from Arctic Alaska to the Antarctic Peninsula. Another chain extends from the Tien Shan and Pamir Mountains westward through Europe to the west coast of North America, at latitudes between 35° and 55°. A third chain extends from the Polar Urals westward through Scandinavia, Iceland, and Canada to Arctic Alaska, approximately at the Arctic Circle. The McCall, Gulkana, and Wolverine Glaciers in Alaska, the Berendon, Place, and Sentinel Glaciers in the Coast Mountains of British Columbia, the South Cascade Glacier in Washington, and the Maclure Glacier in California form the north-south profile in North America. In addition, McCall and Gulkana Glaciers are the westernmost members of the Arctic Circle east-west chain, and South Cascade Glacier (together with the Place and Sentinel Glaciers) forms the western end of the 35°–55° east-west chain.

This paper first presents a general discussion of the regional and local settings of the glaciers. This is followed by the results for ice and water balances from South Cascade Glacier for the hydrologic years ending on September 30, 1965 and 1966, and results from Gulkana and Wolverine Glaciers for the hydrologic year 1966. Ice and water balances for Maclure, South Cascade, Wolverine, and Gulkana Glaciers will be presented in subsequent years by one publication per hydrologic year. When heat-balance results are compiled, they will be incorporated with the presentations of ice and water balance.

In this report, the introduction, the section on regional setting, and the descriptions of individual glaciers were written by M. F. Meier and Austin Post; precipitation data for Alaska and adjacent Canada were contributed by L. R. Mayo. The results from South Cascade Glacier were supplied by W. V. Tangborn, field leader of that project. The results from Gulkana and Wolverine Glaciers were supplied by L. R. Mayo, field leader of those projects.

REGIONAL PHYSICAL AND CLIMATIC SETTING

GEOLOGY

The glaciers in western North America are in a great chain of highlands—the North American Cordillera. This chain comprises two vast mountain systems that roughly parallel the Pacific coast from California and Colorado through Canada and Alaska; these systems are separated by a complex of basins, plateaus, and isolated mountain ranges.

On the west is the Pacific Mountain System, comprising principally the Aleutian and Alaska Ranges and Kenai, Chugach, Wrangell, and St. Elias Mountains in Alaska; the Coast Mountains in southeastern Alaska and western British Columbia; the Cascade Range in Washington and Oregon; and the Sierra Nevada in California (pl. 1). The Pacific Mountain System consists mainly of eugeosynclinal sediments and volcanic rocks which have been strongly and almost continuously deformed from the Mesozoic to the present time and into which enormous granitic batholiths were intruded, especially during the late Mesozoic. Vulcanism, warping, and faulting continue to the present day.

Farther inland is the Rocky Mountain or Eastern System. This includes the Brooks Range in northern Alaska, the Mackenzie and Selwyn Mountains near the Yukon–Northwest Territories border, and the Rocky Mountains along the Alberta–British Columbia border and south through Montana, Wyoming, and Colorado to New Mexico. In general, these mountains consist of miogeosynclinal sediments that comprise both carbonate and clastic rocks and range in age from Paleozoic to Mesozoic. Most of the mountain-producing deformation, folding, and thrust faulting occurred in the Late Cretaceous. From southern Montana to New Mexico, uplift followed by erosion has exposed Precambrian rocks. Thick sequences of volcanic rocks occur in northwestern Wyoming and southern Colorado.

Between the Pacific and Rocky Mountain or Eastern Systems only a few mountain ranges are sufficiently high and continuous to contain large numbers of glaciers. These include the Wood River Mountains in western Alaska, mountainous areas in north-central British Columbia, and the Selkirk, Monashee, and Purcell Mountains in southeastern British Columbia. These mountains are of varied geology; deformed sedimentary and volcanic rocks predominate except for large areas in southern British Columbia which are dominated by granitic intrusions.

CLIMATE

The region shown on plate 1 extends from lat 34° N. to 72° N. and from the sea coast to points 1,500 kilometers inland; its altitude ranges from 84 meters below to 6,194 m above sea level. The Pacific Ocean adjacent to the Alaskan, Canadian, and Washington-Oregon coasts is relatively stable in temperature all year. Most of the precipitation that forms glaciers in this region is derived from the North Pacific. In winter, moisture-laden storms spawned from the semipermanent Aleutian low-pressure area move inland along the Pacific coast. Very heavy precipitation occurs when these storms impinge on and are lifted by the Pacific Mountain System; annual values of more than 5,000 millimeters have been measured near the coast in Washington, British Columbia, and Alaska. Annual runoff as high as 10,000 mm averaged over river basins on Baranof Island, Alaska, indicates that some coastal mountains receive much more precipitation than is recorded at present precipitation-gage locations. Strong precipitation-shadow effects occur where these storms move over high ridges.

The rather complex precipitation pattern shown near the Pacific coast on plate 1 is a generalization; most individual mountain ridges or mountainous islands show a strongly negative windward-leeward precipitation gradient. The storms from the Aleutian low also bring moisture to the ranges farther inland, but much less total moisture is available to those ranges. Movement of warm moist air into the interior is frequently blocked in winter by a persisting high-pressure area north and east of the Rocky Mountains as well as by the mountains themselves. Consequently, winter precipitation decreases markedly from the coast inland. In summer the Aleutian low shifts northward and a North Pacific high-pressure system develops. This brings fog but little precipitation to the coast of Oregon and California. Storms moving north of the high-pressure area bring to the Alaskan coast abundant summer precipitation, and precipitation rates in late summer and fall may be the highest of the year. Western Washington experiences a mixture of weather types but is generally dry in July. Farther inland, moisture is precipitated from cyclonic or convective storms moving eastward toward low-pressure areas east of Hudson Bay. In the inland mountains of Alaska and western Canada and in the Rocky Mountains, precipitation may be greater in summer than in winter. Even so, the amount of summer snowfall is low, and the annual precipitation is much less than along the coast. North and east of the coastal mountains, annual-precipitation values less than 500 mm are common.

Temperatures are moderate along the Pacific coast; Cordova, Alaska (lat 61° N.), and Eureka, Calif. (lat 41° N.), have approximately the same average yearly maximum temperature (30°C). Winter temperatures are colder along the north coast but not extremely so; the average January temperature of Cordova is −5°C. Inland, on the other hand, the range of temperatures is much more extreme; average yearly high and low temperatures at Snag, Yukon Territory, at lat 62° N. near the Alaska-Yukon border, are about +30°C and −50°C.

ICE BALANCE AND ALTITUDE

Figure 1 shows the variation of net (or annual) balance as a function of altitude for certain glaciers. For each glacier the equilibrium-line altitude, or ELA, is given; this is the altitude, in meters above sea level, at which the balance equals zero. The whole altitude range of the ablation area is shown for each glacier, but for Nisqually, Saskatchewan, Seward-Malaspina, and Taku Glaciers, the whole altitude range of the accumulation area is not shown. The gradient of net balance in the vicinity of the equilibrium-line altitude is termed the activity index [1] (Meier, 1962), or AI, and is a measure of the rate at which mass is transferred from higher to lower altitudes; this is also shown for each glacier. In the region studied, equilibrium-line altitudes tend to increase with continentality and (or) a decrease in precipitation totals and tend to increase with decreasing latitude. In general, high activity indices tend to indicate a maritime high-precipitation environment and low activity indices a dry continental environment. However, the activity index is to a slight extent also a function of glacier size (small glaciers tend to show higher activity indices).

Data for the following glaciers were taken from the indicated sources: Dinwoody (Meier, 1951), Maclure (Scully, written commun., 1969), Grasshopper (Alford and Clark, 1968), Saskatchewan (Meier, 1960), Peyto,

[1] Activity index and all other terms in this report referring to ice, snow, and water balances are given in terms of water-equivalent measure. For example, an ice balance of 1 m for a glacier indicates ice equal in mass to a volume of water 1 m in depth covering the area of the glacier.

Place, and Woolsey (Østrem, 1966), Gulkana (Mayo, this rept.), Blue (LaChapelle, written commun., 1969), Berendon (Mathews, written commun., 1969), Taku (Nielsen, 1957), Lemon Creek (LaChapelle, written commun., 1961), Nisqually (Hodge, written commun., 1970), South Cascade (Tangborn, written commun., 1970), McCall (Keeler, 1958), and Wolverine (Mayo, this rept.).

GLACIERS OF THE PACIFIC MOUNTAIN SYSTEM

Most of the glacial ice and the largest glaciers in western North America occur in the Pacific Mountain System. A wide variety of glacier types and glacier

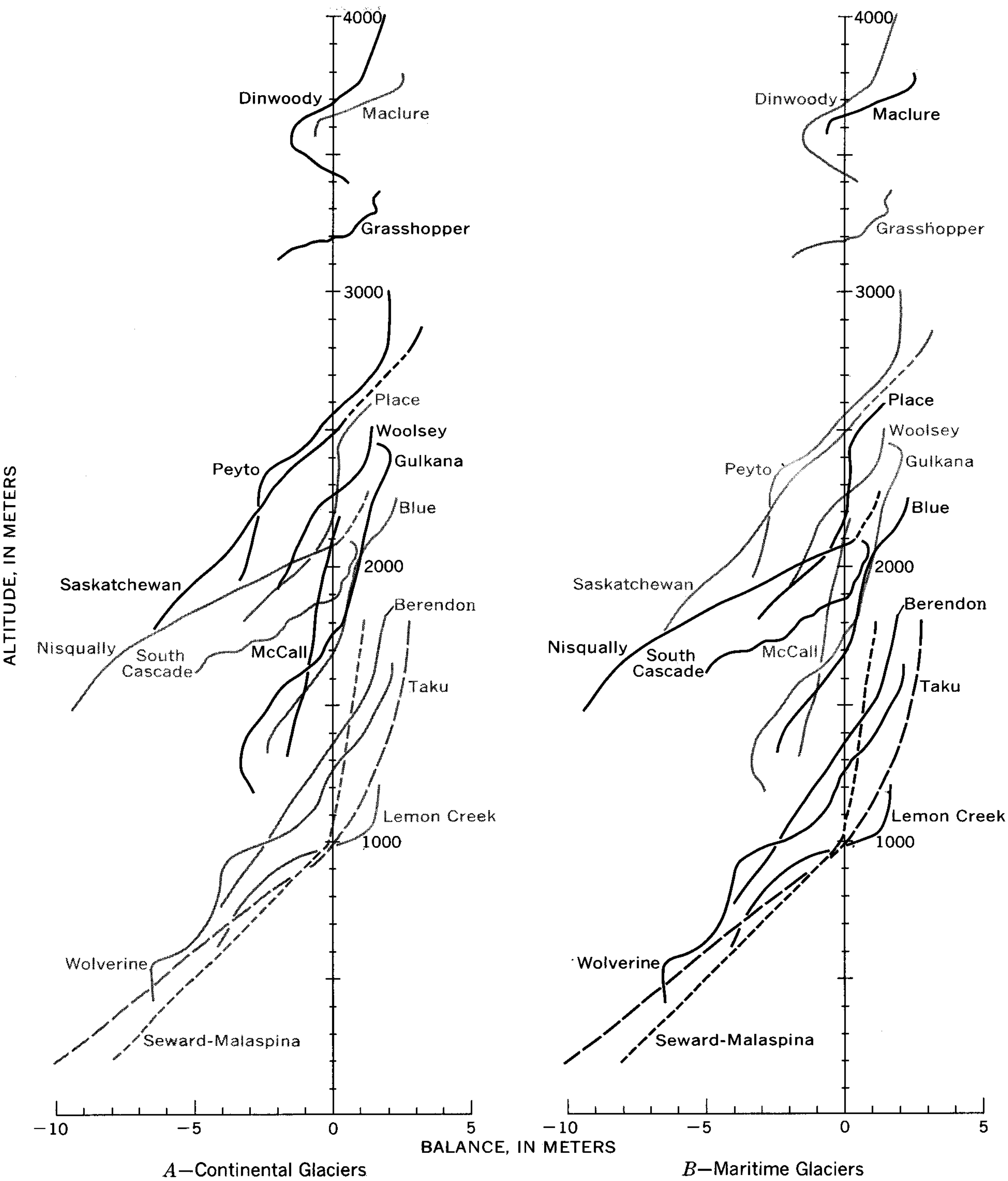

FIGURE 1.—Ice balance versus altitude for representative glaciers.

environments occurs in this mountain system, as can be noted from the large range in equilibrium-line altitudes and activity indices of the maritime glaciers shown in figure 1. Glaciers occupy about 73,000 km^2 in Alaska alone. The coastal glaciers are probably temperate except for those parts of glaciers that are at very high altitudes. Glaciers that are on the inland side of the St. Elias and Chugach Mountains may, in part, be colder than 0°C. Completely polar glaciers probably do not occur in western North America.

The coastal Kenai Mountains (fig. 2) are fairly low, but snowfall is so heavy that large icefields occur and equilibrium-line altitudes are the lowest of the major glacierized areas in western North America. Chenega Glacier (fig. 2), a major outlet of the Sargent Icefield, has an ELA of only 550 m. Glaciers northwest of coastal icefields have less snowfall and higher equilibrium-line altitudes. For example, the ELA of Wolverine Glacier is about 1,200 m and its AI (fig. 1) is about 9 mm/m. Most of the glaciers in this area are stable or are retreating slightly.

Northeast of the Kenai Mountains the Chugach Mountains are covered with large icefields and hundreds of mountain glaciers. Many large valley glaciers debouch into the sea; an example is the Columbia Glacier, which is 64 km long and has an area of 1,070

FIGURE 2.—Chenega Glacier, Kenai Mountains, Alaska. This is the largest outlet glacier of the Sargent Icefield. This very active tidewater glacier discharges large quantities of ice into an arm of Prince William Sound. The glacier retreated rapidly early in the century and opened Nassau Fiord; in recent decades the glacier terminus has been stable. Mountain peaks, all less than 2,500 m in altitude, rise as nunataks above the icefield. Farther inland, scattered glaciers, including the Wolverine, demonstrate by their higher altitude and smaller size the precipitation-shadow effects of the coastal icefields. Wolverine Glacier is in the upper left background.

km² and an ELA of about 850 m. Some of these glaciers are retreating, but most are nearly stable; a few, such as the Harvard and Meares, are advancing slightly. No surging glaciers have been identified in the Kenai Mountains and only two in the Chugach Mountains west of the Copper River.

North of the Kenai and Chugach Mountains and extending well into interior Alaska is the long arc of the Alaska Range. The central part of the range contains the highest peak in North America, Mount McKinley (alt 6,194 m), and is the center of an extensive icefield covering about 4,300 km² and containing the largest glaciers of the Alaska Range—Kahiltna Glacier being 75 m long and covering approximately 475 km². East of Mount McKinley are two other mountainous areas in which ice covers 2,000 km² and 1,400 km². A continuous structural valley traversing the northern part of the range (the Denali fault zone) is the locus of a large number of surging glaciers, such as the Muldrow, Susitna, Black Rapids, and Gakona. The Gulkana is a moderate-size valley glacier in the eastern part of the Alaska Range (fig. 3). The rather continental climate of the Gulkana area is indicated by a higher ELA, 1,800 m, and a low AI, 6 mm/m (fig. 1). Except for glaciers that periodically surge, most glaciers in the Alaska Range are retreating markedly. Extensive stream icings (aufeis deposits) form in winter in the valleys below most Alaska Range glaciers.

The eastern Chugach Mountains and the Wrangell Mountains (just south of the eastern extremity of the Alaska Range) merge into the St. Elias Mountains north of Icy Bay. Here an area of more than 50,000 km² is covered with ice; these are the most extensive icefields on the North American continent. Noteworthy

FIGURE 3.—The eastern Alaska Range in interior Alaska. Gulkana Glacier is on the left. The peaks shown here are as much as 3,000 m in altitude. The Richardson Highway follows the shore of Summit Lake (alt 979 m), well above the regional timberline (alt 750 m). A conspicuous trimline above the Gulkana Glacier terminus marks the 1875 ice position.

glaciers include the Bering, the largest glacier in North America (200 km long and 5,800 km^2 in area); the Seward-Malaspina, the type example piedmont glacier (5,200 km^2 in area); and the Hubbard, a valley glacier 140 km long. On the seaward side of the range, equilibrium-line altitudes are very low and activity indices are rather high. For instance, the Seward-Malaspina Glacier has an ELA of 1,000 m and, in spite of its large size, has an AI of about 10 mm/m (fig. 1). On the inland side of the range, the equilibrium-line altitudes are considerably higher and the activity indices are presumably lower, although none has been measured. On the extreme northeastern fringe of the range, subpolar ice (having temperatures considerably below the freezing point) is known to occur in the lower reaches of the glaciers. Many surging glaciers occur in the St. Elias Mountains; most other glaciers there are probably stable or are retreating slightly, although some tidal glaciers (such as the Hubbard) are advancing. A few other tidewater glaciers are receding exceptionally fast.

A vast system of icefields and isolated glaciers extends along the crest of the Coast Mountains of southeastern Alaska and British Columbia from the Yukon border south to the State of Washington. The Juneau Icefield, about 5,100 km^2 in area, is best known. Glaciers associated with it include the strongly advancing large Taku outlet glacier (ELA 1,000 m, AI about 13 mm/m; fig. 1) and the much smaller retreating Lemon Creek Glacier (ELA 1,000 m, AI about 18 mm/m). Another large icefield is situated north of the Stikine River. Farther south are many large valley glaciers and innumerable mountain glaciers. The Berendon Glacier near Stewart, British Columbia, has an ELA of about 1,400 m and an AI of about 7 mm/m (fig. 1). In southern British Columbia, the Place Glacier has a much higher ELA (2,200 m) but about the same activity index. Many of the smaller glaciers of the Coast Mountains were advancing in the early 1960's, but only a few had continued to do so by 1969; most large glaciers (with the notable exception of the Taku) have been retreating.

The Pacific Mountain System continues southward into the United States as the Cascade Range and smaller outliers, the Olympic Mountains and the Coast Range, to the west. Blue Glacier in the Olympics has an ELA of 1,700 m and an anomalously low AI of 5 mm/m (fig. 1). South Cascade Glacier in the North Cascades of Washington has an ELA of 1,900 m and a more typical AI of 17 mm/m. The North Cascades are particularly rugged (fig. 4); many glaciers in this range advanced in the period 1950–58. From southern Washington to northern California the glaciers occur only on high volcanoes. Nisqually Glacier on Mount Rainier, Washington, advancing in response to increased snowfall since 1946, reflects the maritime climate in this area in its AI of 20 mm/m (fig. 1). Farther south warmer summers and lower precipitation cause the equilibrium-line altitude to rise abruptly. Several small cirque glaciers occur in the Sierra Nevada (fig. 5); these show both maritime and some continental characteristics. Maclure Glacier has an ELA of 3,600 m and an AI of 23 mm/m (fig. 1). Most glaciers in the southern Cascades and in the Sierra Nevada are stable or are retreating slowly.

GLACIERS OF THE ROCKY MOUNTAIN OR EASTERN SYSTEM

The Rocky Mountain System contains thousands of mountain glaciers but no major icefields. Many tiny cirque glaciers occur in the Brooks Range, but moderate-size valley glaciers occur only in the Romanzof Mountains near the eastern end of the Brooks Range. In this high-latitude environment the glaciers are subpolar and small stream icings form below the termini of the large valley glaciers. McCall Glacier has the lowest AI (2 mm/m; fig. 1) of any glacier measured in western North America. Small mountain glaciers in the Mackenzie and Selwyn Mountains of Canada are undoubtedly subpolar but have not been studied in detail.

Considerable information is available on glaciers in the Canadian Rockies along the Alberta–British Columbia border. Small icefields covering up to 280 km^2 occur near the headwaters of the Columbia, Athabasca, and Saskatchewan Rivers. Saskatchewan Glacier (21 km long) has an ELA of 2,500 m and an AI of 9 mm/m (fig. 1); the nearby Peyto Glacier has an ELA of 2,600 m and the same activity index. In the Rocky Mountains of the United States, only small cirque glaciers occur except in the Wind River Mountains of Wyoming. The glaciers in Montana, Wyoming, and Colorado are found at very high altitudes (approaching 4,000 m in Colorado). However, in some areas such as the Wind River Range, the local mesoclimate results in large amounts of orographic and convective precipitation and abundant wind drifting so that considerable snow accumulation is possible; the high-altitude (ELA 3,700 m) Dinwoody Glacier in the Wind River Mountains has an AI of 13 mm/m, and the slightly lower altitude Grasshopper Glacier in the Beartooth Mountains has an AI of about 22 mm/m (fig. 1). Almost all the glaciers in the Rocky Mountains of Canada and the United States are retreating markedly.

GLACIERS OF THE INTERIOR SYSTEM

In the mountain ranges between the Rocky Mountain System and the Pacific Mountain System, many scattered small mountain glaciers exist. The only area where glaciers have been studied is in the Selkirk and Monashee Mountains. Woolsey Glacier in the Selkirk Mountains has typically intermediate values of ELA, 2,300 m,

FIGURE 4.—Ladder Creek and Neve Glaciers, Snowfield Peak, in North Cascades National Park, Washington. This scene displays the rugged nature of the North Cascades where glacierized peaks rise abruptly from heavily timbered valleys that are less than 1,000 m above sea level. The Cascade divide is crossed by trail at Park Creek Pass (alt 1,845 m) top left; Goode Mountain (alt 2,800 m) is to the left, and Buckner Mountain (alt 2,780 m) with Boston Glacier (7 km² in area) is to the right of the pass. Timberline is at about 1,860 m.

and AI, 9 mm/m (fig. 1). Tiny vestigial glaciers occur in the intermontane ranges of the United States, such as the Wallowa, Salmon River, and Wasatch Mountains in northeastern Oregon, Idaho, central Utah, and eastern Nevada. Whether some of these are true glaciers or not is a matter of definition. In northwestern Alaska, small glaciers are present only in the Wood River and Kigluaik Mountains.

DESCRIPTIONS OF INDIVIDUAL GLACIERS

GULKANA GLACIER

Gulkana Glacier is a branched valley glacier in interior Alaska on the south-facing flank of the eastern Alaska Range. These mountains are generally high, having individual peaks up to 3,000 m in altitude, and are very rugged. Bedrock is a complex of Precambrian metamorphic rocks, Mesozoic intrusive and volcanic rocks, and sedimentary rocks ranging from Paleozoic to Quaternary in age. Some of the larger glaciers of the area, such as the Gakona, Canwell, and Black Rapids, are subject to periodic surging; there is some evidence that the Gulkana Glacier may have surged 20 or more years ago. Gulkana Glacier is easily accessible from the Richardson Highway near Isabel Pass (pl. 2*A*, map).

Ice and snow in four adjacent cirques form the accumulation area of Gulkana Glacier. These cirque glaciers

FIGURE 5.—The Sierra Nevada, California. In the foreground is Mount Dana and Dana Glacier. The Lyell Glaciers are situated on Mount Lyell, top right, and Maclure Glacier is partly in shadow farthest to the right. During Pleistocene glaciation the highest peaks, up to 4,000 m in altitude, jutted from an extensive icecap that covered the High Sierras, and presented a scene quite similar to the Sargent Icefield (fig. 2). Timberline is about 3,300 m.

converge in a simple south-flowing tongue that forms the ablation area, 4 km in length; the terminus, which is lightly covered by debris, has an altitude of approximately 1,160 m (pl. 2*A*, photo). The ELA is about 1,800 m. Gulkana Glacier drainage basin (31.6 km^2) contains Gulkana Glacier (21.8 km^2), several smaller glaciers and perennial snowfields (2.6 km^2), and a fairly large area of ice-cored moraine (1.5 km^2). In this report, the snow and ice balances of all active glaciers and perennial snowfields (69 percent of the basin area) are reported together. Ablation from the ice-cored moraine is reported in the basin totals only. Drainage from the glacier flows over an outwash plain and in the past has flowed into Summit Lake, 11.5 km south of the glacier; in recent years the stream has been directed westward to the Delta River (pl. 2*A*, map).

Gulkana Glacier has the highest latitude of the four discussed here (63°15′ N.). Although it is within the Pacific Mountain System, the Chugach Mountains and the Copper River Lowland lie between the glacier and the Pacific Ocean. The local climate verges on the continental, as is indicated by the following low values, measured in recent years: AI, 6 mm/m; winter balance, 1.0 m; and annual exchange, 2.2 m. In mean altitude, accumulation, ablation, and recent activity, Gulkana Glacier is considered to be fairly representative of the glaciers in this region.

This glacier has been studied by University of Alaska research groups frequently since International Geophysical Year investigations were initiated in 1957. The recent history of Gulkana Glacier shows advances in 1580, 1650, 1830, and 1875. The maximum advance, in 1830, extended more than 2.5 km beyond the present terminus position. Since 1875 this glacier has been steadily retreating (Reger, 1968). Measurements of snow and ice balances in Gulkana Glacier basin were begun in 1966 as part of the United States contribution to IHD studies of combined balance on selected glaciers.

WOLVERINE GLACIER

Wolverine Glacier is a valley glacier in the Kenai Mountains, Kenai Peninsula, in south-central Alaska. The glacier is 12 km southwest of Kings Bay, a fiord extending inland from the Gulf of Alaska (pl. 2*B*, map). Although the Kenai Mountains are fairly low, extensive icefields (fig. 2) and many hundreds of small glaciers are present. The large ice masses result from extremely heavy snowfall; their mean ELA averages about 1,000 m —the lowest in western North America. The bedrock is primarily eugeosynclinal deposits of slate, graywacke, and pillow basalts intruded by granodiorite. Access to Wolverine Glacier is usually by skiplane landing on the glacier or by floatplane landing on Paradise Lakes 3 km to the southwest.

The accumulation area of Wolverine Glacier is in a gently sloping basin 4 km wide; from this basin the ice descends by a steep icefall to a valley tongue 5 km long and tapers from a width of 1.5 km to a sharp-pointed terminus at an altitude of 370 m (pl. 2*B*, photo). The glacier is almost free of debris, is moderately crevassed, and is the most active glacier of those described in this report. Wolverine Glacier drainage basin (24.9 km^2) includes Wolverine Glacier and several perennial snowfields (18.0 km^2), two small areas of ice-cored moraine (0.13 km^2), and numerous small lakes in an unglacierized tributary basin. The glacier and perennial snowfields constitute 72 percent of the basin area. The stream issuing from the glacier flows down a narrow gorge to Nellie Juan River, which empties into Kings Bay. The glacier was approximately 1 km longer 50–100 years ago.

Wolverine Glacier lies in the precipitation shadow of the Sargent Icefield. Even so, it has a very maritime climate characterized by fairly high precipitation totals. The winter balance, 2.5 m, and annual exchange, 5.5 m, are high; the ELA, 1,200 m, is low; and the AI, 9 mm/m, is moderate. Wolverine Glacier is considered to be fairly representative of the small glaciers of the Kenai Mountains and the coastal side of the Chugach and St. Elias Mountains.

The U.S. Geological Survey studies initiated in 1966 are the first scientific investigations made on this glacier. Reconnaissance measurements of snow and ice balances in Wolverine Glacier basin were begun in 1966 as part of the United States contribution to the IHD.

SOUTH CASCADE GLACIER

South Cascade Glacier is a valley glacier on the western slope of the main divide of the North Cascades in north-central Washington State (pl. 2*C*, map). These rugged mountains, consisting of a complex of high ragged ridges and deep narrow valleys, were heavily glaciated in Pleistocene time, and the peaks and valleys are strikingly sculptured by ice erosion. More than 700 glaciers, ranging in size from ice patches 0.1 km^2 in area to glaciers covering up to 7 km^2, are on the high peaks and ridges in the west and central part of the mountains (Post and others, 1971) where winter precipitation is very heavy (up to 4,500 m annually). East of the Cascade divide precipitation is much less; the glaciers are small and are restricted to the highest peaks. Bedrock in the vicinity of South Cascade Glacier includes banded migmatite, quartz diorite, biotite schist, and metaquartz diorite.

South Cascade Glacier is in the heart of the North Cascades in the Glacier Peak Wilderness Area. Access to the glacier is generally by foot over rugged brushy terrain through the canyon of the South Fork Cascade

River. A special agreement between the Geological Survey and the U.S. Forest Service permits limited access by helicopters and aircraft for scientific purposes.

Glaciers in the North Cascades are so varied in size and configuration that no single glacier can be considered typical. South Cascade Glacier is larger, has a lower gradient, and lies at a lower altitude than most other glaciers in the region. Perhaps its most distinctive characteristic is a low accumulation basin which actually lies below the altitude of trees on some adjoining slopes. The basin occupies a small north-facing valley, and the glacier now terminates at the edge of a small deep lake. The lake drains to the South Fork Cascade River, a tributary of the Skagit River which discharges into Puget Sound 100 km to the west. The area covered by the trunk or main glacier and by snow or ice in the drainage basin that remains permanently connected to the main glacier is 2.9 km^2. The South Cascade Glacier drainage basin (6.1 km^2) is approximately 54 percent glacierized; the main glacier accounts for 47 percent and the disconnected patches of snow and ice 7 percent of the total drainage area. The AI of South Cascade Glacier, 17 mm/m, is relatively high; ELA, 1,900 m, is moderately low; the winter balance, 3.1 m, is high; and the annual exchange, 6.6 m, is also high.

In recent times South Cascade Glacier shows evidence of advances in 2700–2900 B.C., the 16th century, and the middle and late 19th century (Meier, 1964). The maximum advance, when the terminus was 1.4 km beyond its present position, occurred about 1600. Since 1928, when the glacier filled the basin of the present lake (pl. 2*C*, photos), the glacier has thinned and retreated almost continuously. Between 1928 and 1944, the annual balance was estimated to have averaged approximately -1.7 m; between 1945 and 1965, it averaged approximately -0.6 m (Tangborn, 1968). The dynamic response characteristics of South Cascade Glacier have been analyzed by Nye (1963, 1965). Detailed glaciological studies have been conducted on this glacier by the Geological Survey since the International Geophysical Year (Meier and Tangborn, 1965).

MACLURE GLACIER

Maclure Glacier is one of several small cirque glaciers near the crest of the Sierra Nevada in Yosemite National Park, Calif. The High Sierras embraces an area approximately 600 km long north to south and 130 km in width. Unlike the North Cascades and other mountains farther north, the Sierra Nevada chiefly consists of a high, rolling, moderately dissected upland. Here massive granitic rocks have been unroofed by erosion. Much of the area displays spectacular effects of Pleistocene glaciation—Yosemite Valley is one of the world's finest examples of glacial erosion. The present glaciers are tiny vestiges of ice confined to protected nooks in the highest parts of the range. Maclure Glacier occupies a high shallow cirque on the north side of Mount Maclure and is the easternmost of the Lyell Glaciers (pl. 2*D*). Access is by foot from Tuolumne Meadows or by helicopter.

Maclure Glacier is 0.5 km long and covers an area of 0.2 km^2. Below the clean ice of the glacier, a thick moraine, probably ice cored, presumably represents the Neoglacial extent of this glacier. The local climate shows maritime characteristics (Maclure Glacier's AI is 23 mm/m, its winter balance is 2.3 m, and its annual exchange is 4.6 m) in spite of a very high ELA of 3,600 m. Runoff from the 0.97-km^2 drainage basin containing Maclure Glacier flows into Tuolumne River, which flows west into the San Joaquin Valley of California.

It is interesting to note that the first quantitative glacier studies on the continent were evidently conducted on this glacier in 1872 by the famous naturalist John Muir. Among his studies were observations of glacier movement, which amounted to about 2 cm per day near the center of the glacier. Since then the glacier has been observed by many visitors in Yosemite National Park, and its growth and decline have been measured by the U.S. National Park Service at frequent intervals. Studies of ice and water balances were initiated by the Geological Survey in the summer of 1966 as part of the United States contribution to the IHD. These data will be presented in subsequent reports.

MEASUREMENTS

ICE-BALANCE TERMS

Most ice-balance data for glaciers are obtained by the use of stakes, pits, or probes, referenced to or from a marker horizon that is usually a summer surface. This is termed the stratigraphic system of measurement (UNESCO/IASH, 1970). However, summer surfaces may be transgressive with time so that a comparison of stratigraphic ice-balance data with water-balance data is difficult. This difficulty may be circumvented by use of the annual system, which references all measurements to values taken at two instants in time, normally the beginning and end of a hydrologic year. The annual system, however, is not convenient in regard to field measurement. Some scheme is needed to relate these two systems and take advantage of the best qualities of each.

We use a combined system that relates the annual and stratigraphic systems on the basis of identification of the material involved—snow and frozen water of the year under consideration, previously deposited ice and firn, and new snow deposited after the summer of the year in question. This system also relates changes in these materials to hydrologic and meteorologic quantities.

The terms used in this report are illustrated on a graph (fig. 6) showing ice-balance quantities, reported in millimeters or meters of water equivalent, averaged over the glacier as a whole. Symbols for these terms use the following conventions: A bar over the letter symbol indicates an average taken over the whole glacier or drainage basin; a letter in parentheses following a symbol indicates that only the material snow (*s*), ice and old firn (*i*), or new firn (*f*) is considered; if no parentheses follow a symbol, the total mass (undifferentiated) is considered; *the subscript 0 refers to measurements* made on or about the beginning of the hydrologic year, which runs from t_0 (Oct. 1) to t_1 (Oct. 1); the subscript *a* refers to quantities measured over one hydrologic year; the subscript 1 refers to measurements made only at the end of a hydrologic year; and the subscript *n* refers to measurements made close to but not at the end of the hydrologic year. At t_0, the amount of snow (including superimposed ice) above a summer surface overlying older ice and firn is the initial snow balance $\bar{b}_0(s)$. At t_1, the amount of newer snow overlying a new summer surface is the final snow balance, $\bar{b}_1(s)$.

The amount of the old ice and firn lost during the hydrologic year as measured from t_0 to t_1 is the annual ice balance, $\bar{b}_a(i)$. The net change in glacier mass from t_0 to t_1 is the annual balance, $\bar{b}_a$. The glacier actually reaches a minimum mass at times t_0' and t_1', generally close to but not the same as t_0 and t_1. The net change in glacier mass from t_0' to t_1' is the total mass net balance, $\bar{b}_n$. An initial balance increment, $\bar{b}_0$, relates the balance at t_0' to that at t_0. At some time t_0'', after t_0', ablation generally ceases for a considerable length of time during the winter. The change in mass of old ice and firn from t_0 to t_0'' is the initial ice balance, $\bar{b}_0(i)$. As new snow accumulates near the end of the hydrologic year, the older snow becomes firn. At some time t_1'', after t_1', ablation again ceases for a period during the ensuing winter.

The term "net balance" has been used in the past to describe the difference between the amount of ice lost and the amount of snow added to storage as firn. This concept can be defined more exactly in two ways. The change in mass of old firn and ice at t_0'' to new firn at t_1'' is the firn and ice net balance, $\bar{b}_n(fi)$; the amount of new firn accumulated at t_1'' is the net firnification, $\bar{b}_n(f)$. Alternatively, the change in mass of old firn and ice at t_0 to new firn at t_1 is the firn and ice annual balance, $\bar{b}_a(fi)$; the amount of new firn accumulated at t_1 is the annual firnification, $\bar{b}_a(f)$. This alternate scheme should not be used on glaciers, such as South Cascade, in which appreciable changes in the amount of new firn may occur after the end of a hydrologic year.

Other useful parameters are the maximum balance during the hydrologic year, $\bar{b}_x$ (generally estimated or computed from other data), and the measured late-

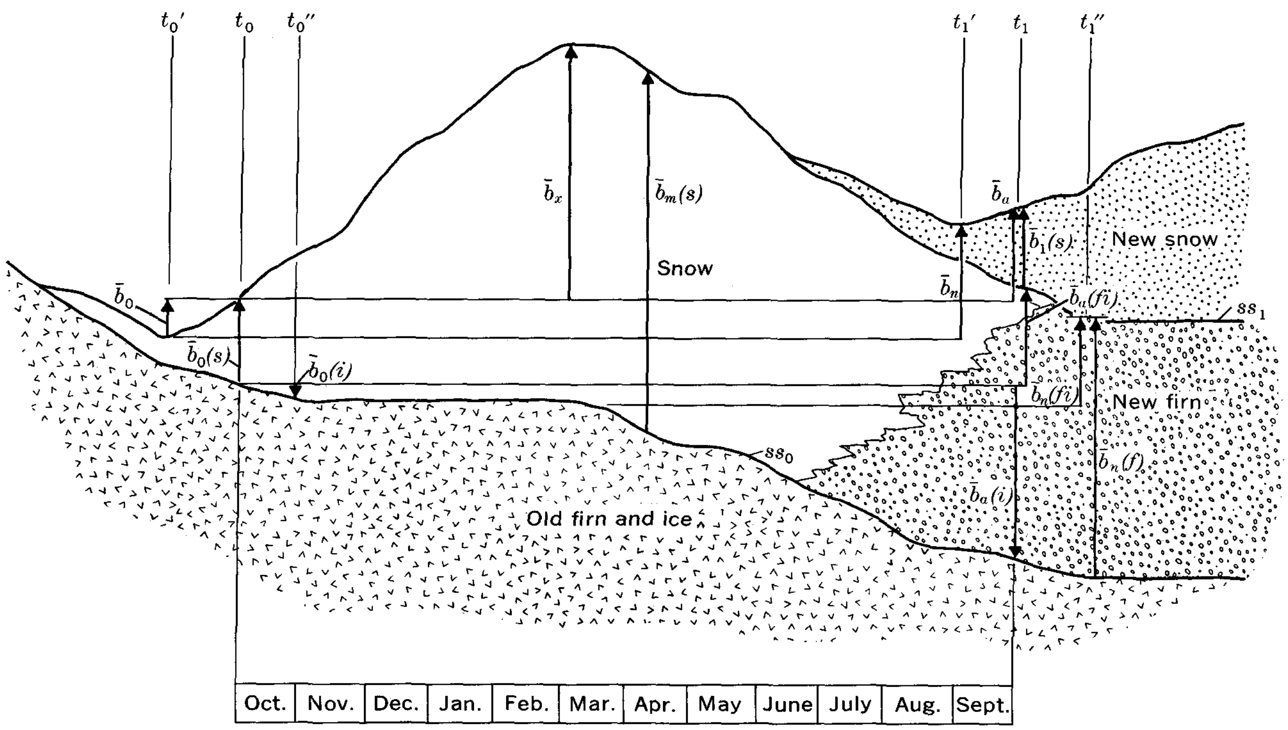

FIGURE 6.—The combined system for ice-balance terms. This diagram portrays the ice-mass changes with time, averaged over a whole glacier. See text for description of terms.

winter snow balance, $\bar{b}_m(s)$. Not all these terms are measured, computed, or even needed for all glaciers in this program, but all must be defined to relate the results from different glaciers properly. Other terms relating to accumulation, ablation, balance, exchange, and glacier areas are used as given in UNESCO/IASH (1970).

EVALUATION OF ERRORS

Errors arising from measurements of ice, snow, and water balances are difficult to treat analytically owing to the inherent difficulty of measuring large masses of moving and compacting snow and ice and to sampling problems. Reliable measurements of standard errors of balance values at a single stake have been made. But even these error measurements may be subject to large unknown influences if there are long time periods when no observations are made or if there are unexplained vertical movements of ablation stakes; in addition there are always difficulties and unknown errors in making continuous and accurate density measurements. Interpolating between widely separated points over large areas of glacier ice and snow is normally done using snowline information as a guide. This improves the result but also drastically increases the difficulties of assigning meaningful error values to each measurement. The standard errors given in this report should be considered as little more than approximations; they represent a combination of error measurements at points and subjective error estimates for interpolations or values not susceptible to error measurement. We have used the rule that the error of an indirectly measured quantity (which is the sum or product of two or more directly measured quantities) is equal to the square root of the sum of the squares of the errors of the directly measured quantities. Standard error values are given in the same dimensions as the balance values to which they apply (meters of water equivalent).

1965 HYDROLOGIC YEAR

SOUTH CASCADE GLACIER

Field Program

Most of the instrumentation needed for measurements of ice and water balance had been installed by January 1, 1965, the beginning of the IHD. As a consequence, nearly complete records of streamflow at the outlet of the drainage basin, air temperature at two locations, and precipitation catch in a gage at one location were obtained for the 1965 hydrologic year (table 1). Ice-balance, wind-speed, and additional precipitation data were obtained during the summer months.

Instrument locations together with selected snowline positions are shown on plate 3*A*. The distribution of snow cover over the basin and the late-winter snow balance, $b_m(s)$, were measured on May 12, 1965 (pl. 3*B*), just 15 days before the time of maximum balance. The minimum-balance condition was reached on November 2, 1965 (t_1'; pl. 3*C*).

Climate

The weather in the North Cascades during the 1965 hydrologic year was marked by slightly above average winter snowfall, a cool spring, and a warm autumn (pl. 3*D*, 1st and 2d graphs). Except for part of June the summer was cloudier but had less precipitation than

TABLE 1.—*Instrumentation at South Cascade Glacier during the 1965 hydrologic year*

[Continuous record ———, estimated record - - - -, occasional measurement • •]

Station	Altitude (meters)	Measurement	1965 Hydrologic year												
			1964			1965									
			Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.
1	1,610	Streamflow	———	———	———	- - - -	———	———	———	- - ———	———	——— - - ———	———	———	———
1	1,610	Air temperature	———	———	———	- - - -	———	———	———	- - ———	———	——— - - ———	———	———	———
1	1,610	Precipitation	———	———	———	———	———	———	———	———	———	———	———	———	———
1	1,610	Wind speed	———	———			—			———	———	———	———	——— —	———
1	1,610	Balance								•	•	•			•
2	1,840	Air temperature	———	———	- - - -	- - - -	———	———	- - - -	———	———	———	———	———	———
2	1,840	Precipitation									———	———	———	———	
2	1,840	Wind speed										———	———	———	
P-0	1,730	Balance								•	•	•	•	•	•
P-1	1,860	Balance	•							•		• •	• •	• •	•
P-1	1,860	Precipitation								•		• •	• • •	• • •	•
P-3	2,045	Balance										•	•	•	•
P-3	2,045	Precipitation								•	•	• •	•	•	•
Basin photographed										•	•	• •	•	•	•
Basin occupied by personnel			—				•			—	———	———	———	———	—

average. The accumulation season at South Cascade Glacier continued until almost the end of May, and short periods of snow accumulation occurred every month except July (pl. 3*D*, 1st graph). The mean winter air temperature was −2.9°C, the mean summer temperature +7.3°C, and annual mean temperature +1.3°C.

Ice Balance

Unusually small basin-storage changes occurred during October and November of 1964, as shown by the hydrologic-balance curve (pl. 3*D*, 3d graph). The near-zero balance until November 22 was due to the mild temperatures and low snow accumulation during these two months. Such subtle, seemingly minor climatic aberrancies during the normal accumulation period can significantly alter the ultimate glacier balance.

Table 2 gives snow, ice, and water balances and related values for the South Cascade Glacier during the 1965 hydrologic and balance years. The standard error for each measurement is shown adjacent to each param-

TABLE 2.—*Snow, ice, and water balances, South Cascade Glacier, 1965 hydrologic and balance years*

[Parameter values and errors in meters except where indicated. Date: Hydrologic year, Oct. 1, 1965 (t_0), to Sept. 30, 1966 (t_1)]

	Glacier		Basin		Date	Term	Explanation
	Value	Error	Value	Error			
Parameters for annual mass balances							
$\bar{b}_a$	0.07	0.10	0.04	0.18	Hydrologic year	Annual balance	Net change in glacier mass from t_0 to t_1; approximately equal to the difference between precipitation as snow and melt-water runoff for one hydrologic year.
$\bar{b}_n$	−.17	.12	−.09	.10	Nov. 1, 1964 (t_0') to Nov. 2, 1965 (t_1').	Total mass net balance	Net change in glacier mass from t_0' to t_1'; change in mass during one balance year.
$\bar{b}_n(fi)$	−.19	.13	−.10	.10	Nov. 22, 1964 (t_0'') to Nov. 22, 1965 (t_1'').	Firn and ice net balance	Change in mass of old firn and ice during a single melt season; the mass between two consecutive summer surfaces.
Parameters relating annual and net mass balances							
$\bar{b}_0$	0.03	0.01	0.02	0.01	Oct. 1 to Nov. 1, 1964	Initial balance increment	Balance change between first time of minimum balance (t_0') and t_0; relates balance-year quantities to hydrologic-year quantities.
$\bar{b}_0(s)$	.03	---	.02	---	Oct. 1, 1964	Initial snow balance	New snow accumulated on summer surface (ss_0) at t_0.
$\bar{b}_0(i)$	−.08	.05	−.04	.05	Oct. 1 to Nov. 21, 1964	Initial ice balance	Ice melt in the ablation area after t_0 and before ice melt begins the following spring; measured by ablation stakes.
$\bar{b}_1(s)$	0	---	0	---	---	Final snow balance	New snow accumulated on summer surface (ss_1) at t_1.
$\bar{b}_a(i)$	−.60	.05	−.30	.10	Hydrologic year	Annual ice balance	Ice and firn melt in the ablation area *for the hydrologic year.*
$\bar{b}_n(f)$	.60	.08	.30	.15	Nov. 22, 1965	Net firnification	The increment of new firn in the accumulation area at t_1''; measured after ablation ceases in the autumn.
Parameters for snow accumulation and ablation							
$\bar{b}_x$	3.67	0.18	2.48	0.20	May 27, 1965	Maximum balance	Maximum value of the balance (in relation to balance at t_0) for the hydrologic year, similar to the "winter balance" or the "apparent accumulation."
$\bar{b}_m(s)$	3.48	.15	2.40	.20	May 12, 1965	Late-winter snow balance	Balance measured to the summer surface (ss_0) in late winter or spring; measured in pits or by probing.
$\bar{c}_a$	4.00	.18	2.66	.25	Hydrologic year	Annual accumulation	Accumulation of snow between t_0 and t_1.
$\bar{a}_a$	−3.93	.20	−2.62	.20	do	Annual ablation	Ablation of snow, ice, and firn between t_0 and t_1.
Parameters for glacier dimensions							
S	[1] 2.90	0.04	6.11	0.02	---	Area	
AAR	[2] .53	[2] .03	[2] .30	[2] .05	Sept. 30, 1965	Accumulation-area ratio	A rough index of annual balance, measured at time t_1, neglecting new snow overlying ss_1.
ELA	1880	20	---	---	do	Equilibrium-line altitude	Do.
ΔL	−12	3	---	---	Hydrologic year	Advance or retreat	Average horizontal-distance change of terminus in direction of glacier flow.
Parameters for water balances							
p_a	---	---	2.05	0.50	Hydrologic year	Measured annual precipitation	Value measured with a gage at one point.
$p_a(r)$	---	---	.33	.05	do	Measured annual precipitation as rain.	Do.
$\bar{p}_a^*$	4.25	0.30	3.43	.30	do	Calculated annual precipitation	Calculated average for glacier or basin, using annual basin runoff and balance.
$\bar{r}_a$	4.32	.28	3.39	.10	do	Measured annual runoff	

[1] Glacier area, in km², is larger than previously published value because small connected snow and ice areas are included with the trunk glacier in this report.
[2] Dimensionless.

eter value and is based on calculations of the reliability of the field measurement. For some parameters the standard error is estimated. For example, runoff, $\bar{r}_a$, is measured only for the entire drainage basin, but a value for the glacier alone can be derived on the basis of annual ablation, $\bar{a}_a$, and measured precipitation as rain, $p_a(r)$. The much larger standard error for glacier runoff is then estimated in accordance with ablation errors and errors arising from an approximation of liquid precipitation runoff from the glacier.

The maximum balance, $\bar{b}_x$, was 3.67 m averaged over the glacier, or 2.48 m averaged over the drainage basin (table 2). This is a measure of the maximum water-equivalent depth of seasonal snow reached in late winter or early spring, a quantity which has been referred to as the apparent accumulation (Meier, 1962) or erroneously referred to as just the accumulation. The annual accumulation, $\bar{c}_a$, on the glacier was 4.00 m. The maximum balance was reached on May 27, 1965, 15 days after detailed measurements were made of the late-winter balance, $b_m(s)$, shown on plate 3*B*. Balance change between May 12 and May 27, 1965, was estimated on the basis of recorded precipitation, air temperature, and runoff. For May 12, 1965, the measured snowpack density, in megagrams per cubic meter, was 0.495 on the glacier (P–1, alt 1,860 m) and 0.523 off the glacier (site 1, alt 1,610 m).

The late-winter balance, $b_m(s)$, net balance, b_n, and the summer change in storage, $b_n - b_m(s)$, are shown as functions of altitude on plate 3*E*. Note that all three curves become more negative at the highest altitudes. This is presumably due to wind removal of snow and accounts for the patches of bare ice exposed at high altitudes. The balance curves for the basin are near zero at the altitude of South Cascade Lake because the lake (0.24 km^2 or 4 percent of the basin area) is incapable of supporting a load of snow except for a small area close to the shore. As snow accumulates on the lake during the winter, it displaces water that runs off directly.

The total mass net balance, $\bar{b}_n$, or change in ice storage was -0.17 m averaged over the glacier and occurred between the minimum balances on November 1, 1964, and November 2, 1965 (table 2). The equivalent 1958–64 average net balance was -0.60 m. Factors tending to produce a less negative balance were the above-average winter snowpack, the late spring, and the summer snowfalls. The low altitude of the previous year's snowline was a significant factor in reducing ice melt in July. Large areas of ice are usually exposed in July and undergo heavy ablation owing to the low albedo of ice and to the intense midsummer radiation. The high-albedo 1964 firn covering much of the ice area significantly reduced the total ice melt during the summer. Ablation continued throughout October and caused a net change in basin storage of -4 mm/day during that month. This was an anomalous condition, since the ablation season usually ends shortly after October 1. The result was a positive annual balance of 0.07 m on September 30 and a negative net balance on November 2 (pl. 3*D*, 3d graph). On the glacier, the standard error of the net balance was about 0.12 m and the error in the annual balance about 0.10 m.

Precipitation

One recording precipitation gage is in continuous operation in this drainage basin (site 1, alt 1,610 m, pl. 3*D*, 1st graph). Because of its low altitude and the inherent inaccuracy of precipitation gages in mountainous areas, however, this gage is considered only an index of the average basin precipitation.

An estimate of the total basin precipitation is made by treating the entire basin as a large gage where annual precipitation equals annual runoff plus annual storage changes plus net evaporation, or

$$\bar{p}_a{}^* = \bar{r}_a + \bar{b}_a + \bar{v}$$

where $\bar{p}_a{}^*$ = calculated precipitation, $\bar{r}_a$ = measured annual runoff, $\bar{b}_a$ = measured annual balance, and $\bar{v}$ = net evaporation-condensation balance. We assume that $\bar{v}$ is negligible compared to the other terms; measurements with lysimeters on the glacier in past years have shown that evaporation and condensation on snow are nearly equal and tend to cancel each other out. Net evapotranspiration undoubtedly occurs over the small part of the basin which is vegetated or becomes bare of snow for a small part of each year. This water loss is difficult to estimate but must be of the order of a few millimeters averaged over the whole drainage basin for a whole year. Assuming zero net evaporation, the calculated basin precipitation is 3.43 m.

Cumulative basin precipitation, $\Sigma\bar{p}^*$ (pl. 3*D*, 3d graph), is calculated by multiplying cumulative precipitation recorded at the gage since t_0, Σp, by a ratio of the calculated annual basin precipitation, $\bar{p}_a{}^*$, divided by the annual gage precipitation, p_a:

$$\Sigma\bar{p}^* = \frac{\bar{p}_a{}^*}{p_a}\Sigma p,$$

where $\dfrac{\bar{p}_a{}^*}{p_a} = \dfrac{3.43}{2.05} = 1.67.$

Hydrologic Balance

An estimation of the daily basin balance (here designated as the hydrologic balance; see pl. 3*D*, 3d graph) can be made by subtracting cumulative runoff from cumulative basin precipitation. The standard error of the hydrologic balance curve is large (over 10 percent) because of the large basin-precipitation error. An independent check of the calculated hydrologic balance against the measured maximum late-winter balance on May 27 shows a difference of 7 percent (hydrologic balance = 2.31 m, $\bar{b}_x = 2.48$ m).

1966 HYDROLOGIC YEAR

SOUTH CASCADE GLACIER

Field Program

Instrumentation used in 1966 was virtually the same as in 1965, and the same sites were used (pl. 4*A*). Air temperature at two locations, wind speed, precipitation, and streamflow records are virtually complete for the whole hydrologic year (table 3). The distributions of snow and ice for late winter and for the end of the balance year are shown on plate 4*B* and *C*.

Climate

The climate in the North Cascades during the 1966 hydrologic year was characterized by below-average amounts of precipitation, particularly during the winter season, and by above-average temperatures (pl. 4*D*, 1st and 2d graphs). In June and July, however, temperatures were below normal and precipitation was slightly above average. Snow fell at altitudes above 1,800 m on several occasions in June. With the exception of March and June, high pressure dominated the weather pattern in the Pacific Northwest during the hydrologic year. The mean winter air temperature was −3.5°C, the mean summer temperature was +6.8°C, and the annual mean temperature was +0.8°C.

Ice Balance

The map of measured late-winter balance (pl. 4*A*) as of May 12, 1966, shows a typical pattern of snow accumulation but less than normal amounts; $\bar{b}_m(s) = 2.52$ on the glacier and 1.82 on the whole drainage basin. At this time the snowpack density, in megagrams per cubic meter, was 0.498 on the glacier (P–1) and 0.615 off the glacier (site 1). The maximum balance, $\bar{b}_x$, was attained on May 22; the value for $\bar{b}_x$ is less than that for $\bar{b}_m(s)$ because of considerable melting at the beginning of the hydrologic year (table 4). The annual accumulation, $\bar{c}_a$, was 2.59 m on the glacier or 1.99 m averaged over the drainage basin.

The map of total mass net balance, $\bar{b}_n$ (pl. 4*C*), shows the snow and ice cover on the basin at t_1', the date of minimum balance (Oct. 16, 1966). The change in ice storage of the glacier during the 4-month period since May 22 is very large, amounting to over 3.5 m for the glacier and 2.3 m for the total basin (pl. 4*D*, 3d graph). The basin-storage change during this period represented about 70 percent of the total annual basin runoff, $\bar{r}_a$ (3.25 m). The remaining runoff during the hydrologic year was due to precipitation as rain ($p_a(r)$, 14 percent) and snow and ice melt between October 1, 1965, and May 22, 1966 (16 percent).

The annual balance, $\bar{b}_a$ (−0.94 m on the glacier), was a slightly greater loss in mass than the average since 1958. The annual loss in ice storage in the basin contributed 13 percent of the annual runoff from the basin and about 23 percent of the runoff from the total glacierized area. The greater-than-average loss in mass can be attributed to both the deficient winter snowpack, which resulted in an early exposure of glacier ice, and the warmer-than-average ablation season. This loss of mass eliminated most of the winter snow on the glacier except at the very highest altitudes (pl. 4*E* and fig. 7).

TABLE 3.—*Instrumentation at South Cascade Glacier during the 1966 hydrologic year*

[Continuous record ———, estimated record - - - -, occasional measurement ••]

Station	Altitude (meters)	Measurement	1966 Hydrologic year												
			1965			1966									
			Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.
1	1,610	Streamflow	———	———	———	———	———	———	———	———	———	———	———	———	———
1	1,610	Air temperature	———	———	———	———	———	———	———	———	———	———	———	———	———
1	1,610	Precipitation	———	———	———	———	———	———	———	———	———	———	———	———	———
1	1,610	Wind speed	———	———	———	———	———	———	———	———	———	———	———	———	———
1	1,610	Balance						•		•	••				
2	1,840	Air temperature	———	———	- - -	- - -	———	———	———	———	———	———	———	———	———
2	1,840	Precipitation								——	——	——	———	——	— •
2	1,840	Wind speed								•	——	———	———		
P-0	1,730	Balance								•	••	•	••	••	•
P-1	1,880	Balance	•					•		•	•	•	••••		•
P-1	1,880	Precipitation								•	••••	•	•••	••••	•
P-3	2,045	Balance							•			•	••	•	
Basin photographed									•		•	••	••	••	
Basin occupied by personnel			—				—	—		—	——	———	———	——	—

TABLE 4.—*Snow, ice, and water balances, South Cascade Glacier, 1966 hydrologic and balance years*

[Parameter values and errors in meters except where indicated. Date: Hydrologic year, Oct. 1, 1965 (t_0), to Sept. 30, 1966 (t_1)]

	Glacier		Basin		Date	Term	Explanation
	Value	Error	Value	Error			
					Parameters for annual mass balances		
$\bar{b}_a$	−0.94	0.10	−0.45	0.12	Hydrologic year	Annual balance	Net change in glacier mass from t_0 to t_1; approximately equal to the difference between precipitation as snow and melt-water runoff for one hydrologic year.
$\bar{b}_n$	−1.03	.10	−.50	.12	Nov. 3, 1965 (t_0') to Oct. 16, 1966 (t_1').	Total mass net balance	Net change in glacier mass from t_0' to t_1'; change in mass during one balance year.
$\bar{b}_n(fi)$	−1.07	.10	−.52	.12	Nov. 24, 1965 (t_0'') to Nov. 5, 1966 (t_1'').	Firn and ice net balance	Change in mass of old firn and ice during a single melt season; the mass between two consecutive summer surfaces.
					Parameters relating annual and net mass balances		
$\bar{b}_0$	0.08	0.03	0.04	0.04	Oct. 1 to Nov. 2, 1965	Initial balance increment	Balance change between first time of minimum balance (t_0') and t_0; relates balance year quantities to hydrologic year quantities.
$\bar{b}_0(s)$	0	--------	0	--------	--------	Initial snow balance	New snow accumulated on summer surface (ss_0) at t_0.
$\bar{b}_0(i)$	.20	.05	.10	.05	Oct. 1 to Nov. 23, 1965	Initial ice balance	Ice melt in the ablation area after t_0 and before ice melt begins the following spring; measured by ablation stakes.
$\bar{b}_1(s)$	0	--------	0	--------	--------	Final snow balance	New snow accumulated on summer surface (ss_1) at t_1.
$\bar{b}_a(i)$	−1.03	.08	−.49	.10	Hydrologic year	Annual ice balance	Ice and firn melt in the ablation area for the hydrologic year.
$\bar{b}_n(f)$	.09	.05	.04	.02	Nov. 5, 1966	Net firnification	The increment of new firn in the accumulation area at t_1''; measured after ablation ceases in the autumn.
					Parameters for snow accumulation and ablation		
$\bar{b}_x$	2.39	0.16	1.77	0.22	May 22, 1966	Maximum balance	Maximum value of the balance (in relation to balance at t_0) for the hydrologic year, similar to the "winter balance" or the "apparent accumulation."
$\bar{b}_m(s)$	2.52	.15	1.82	.20	May 12, 1966	Late-winter snow balance	Balance measured to the summer surface (ss_0) in late winter or spring; measured in pits or by probing.
$\bar{c}_a$	2.59	.18	1.99	.20	Hydrologic year	Annual accumulation	Accumulation of snow between t_0 and t_1.
$\bar{a}_a$	−3.53	.20	2.44	.20	do	Annual ablation	Ablation of snow, ice, and firn between t_0 and t_1.
					Parameters for glacier dimensions		
S	[1] 2.84	0.04	6.11	0.02	--------	Area	
AAR	[2] .30	[2] .05	[2] .14	[2] .03	Sept. 30, 1966	Accumulation-area ratio	A rough index of annual balance, measured at time t_1, neglecting new snow overlying ss_1.
ELA	2140	20	--------	--------	do	Equilibrium-line altitude	Do.
ΔL	−16	3	--------	--------	Hydrologic year	Advance or retreat	Average horizontal-distance change of terminus in direction of glacier flow.
					Parameters for water balances		
p_a	--------	--------	1.77	0.35	Hydrologic year	Measured annual precipitation	Value measured with a gage at one point.
$p_a(r)$	--------	--------	.46	.06	do	Measured annual precipitation as rain.	Do.
$\bar{p}_a^*$	3.09	0.33	2.80	.22	do	Calculated annual precipitation	Calculated average for glacier or basin, using annual basin runoff and balance.
$\bar{r}_a$	4.02	.30	3.25	.15	do	Measured annual runoff	

[1] Square kilometers.
[2] Dimensionless.

Precipitation

Precipitation remains one of the most difficult parameters to measure accurately. Standard gages in high mountain environments serve, at best, only as indices of the total precipitation. For example, the well-shielded recording gage at site 1 (alt 1,610 m) measured just 63 percent of the estimated basin precipitation for this hydrologic year. This was due, in part, to this gage's location at the lowest point in the basin. But the gage is inaccurate mainly because about 80 percent of the precipitation occurred as wind-driven snow which is only partially intercepted by the gage.

The actual basin precipitation, $\bar{p}_a^*$, was estimated as the sum of the annual runoff, $\bar{r}_a$ (3.25 m), and the annual balance for the basin, $\bar{b}_a$ (−0.45 m). The ratio of this calculated precipitation (2.80 m) to the measured precipitation, p_a (1.77 m), is 1.58 as compared with 1.67 in the 1965 hydrologic year. The cumulative basin pre-

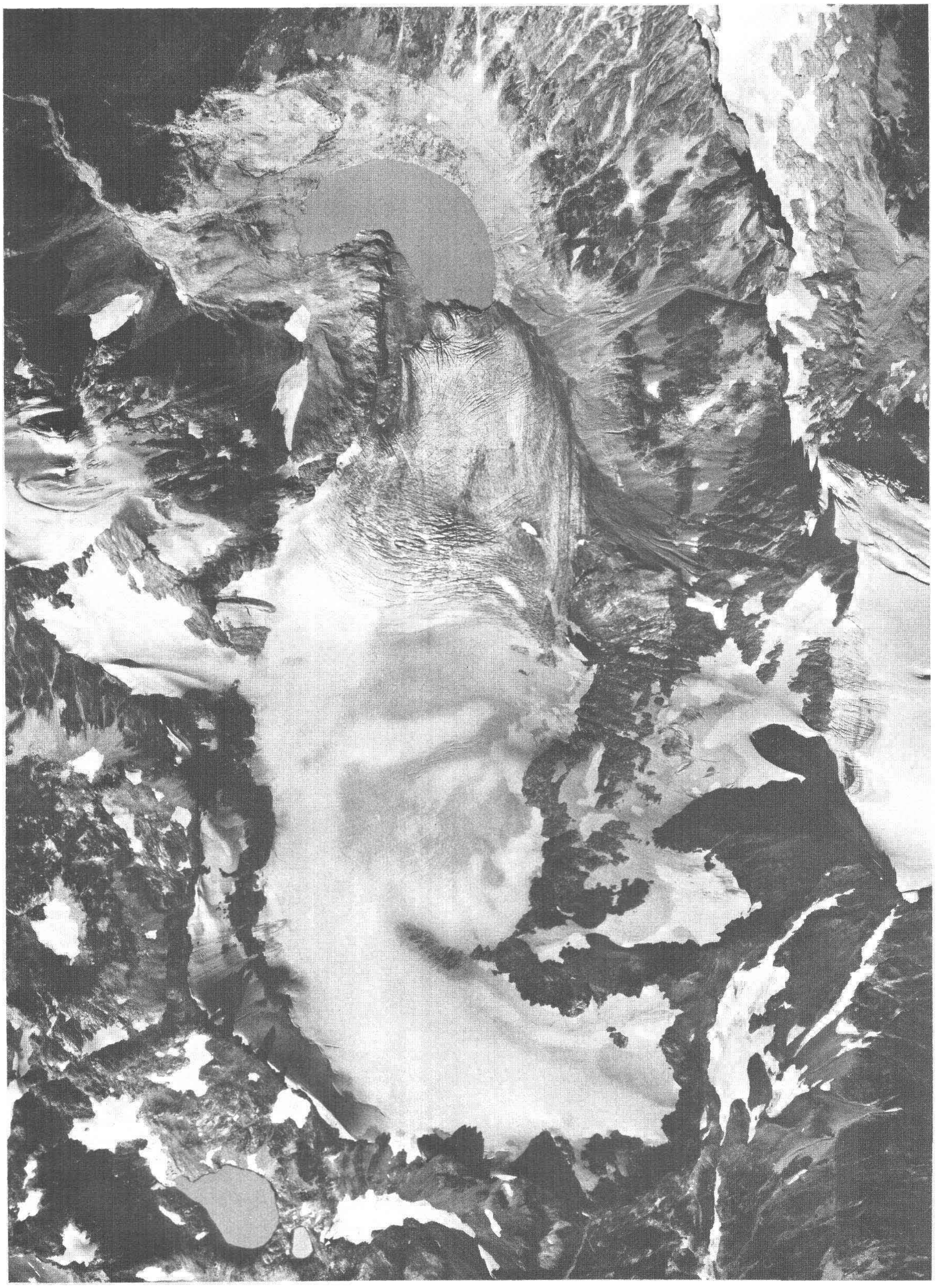

FIGURE 7.—Vertical air photograph of South Cascade Glacier and basin on September 22, 1966. The thin veneer of snow remaining on the glacier at this time is rapidly disappearing. The transient accumulation-area ratio is 0.40, which can be compared with the values of approximately 0.30 on September 30, 1966, and 0.15 on October 16, 1966, the end of the balance year. Although little snow remains on the glacier, accumulation patterns are still obvious in this photograph. The strong influence of the prevailing western storm wind is indicated by drift patterns and by the predominance of snow remaining on the western side of the glacier (north to top of photograph).

cipitation, $\Sigma\bar{p}^*$ (pl. 4*D*, 3d graph), was determined by multiplying the cumulative measured precipitation, Σp_a, by this fixed ratio. The hydrologic-balance curve (pl. 4*D*, 3d graph) is the difference between cumulative precipitation and cumulative runoff.

Contribution of Glacier to Runoff

The runoff from small glacierized basins, such as this one, contributed a significant volume of water to the Cascade River during the months of August and September. During these two months most of the discharge from the South Cascade Glacier basin originated from the melting of glacier ice. The negative balance (-0.94 m) resulting from this high degree of ice melt thus increased the runoff of the low altitude streams to a value far above that produced by precipitation alone. For the South Cascade Glacier basin (6.1 km^2), ice melt contributed 85 percent of the total runoff in August and September; for the Cascade River basin (435 km^2) during the same period melting of ice in South Cascade Glacier alone contributed 7 percent of the total runoff, and ice melt from all glaciers (16 km^2) in the basin contributed approximately 35 percent of the flow. The total runoff from South Cascade Glacier for the 1966 hydrologic year was fairly high (4.02 m).

GULKANA GLACIER

Field Program

Gulkana Glacier was visited in late September 1965, when preliminary ice-balance measurements were made at two sites (table 5). During the summer (May–September 1966) the snow balance was measured, an additional balance site established, a hut constructed, and a streamflow gaging station installed (pl. 5*A*). Continuous streamflow data, however, were not produced until almost the end of the hydrologic year, so none are reported here; nor are any continuous temperature or precipitation data available. These data will appear in reports covering subsequent hydrologic years.

Climate

The weather during the 1966 hydrologic year at several nearby low-altitude weather stations had no outstanding variations from a normal pattern, and the glacier balance was apparently in the normal range. The large new snow accumulation during August and September 1966 occurred during the time when the maritime coastal regions were receiving more than usual precipitation. One measurement made on Gulkana Glacier on September 26, 1965, indicates that roughly 0.10 m water equivalent of new snow, the initial snow balance, $\bar{b}_0(s)$, was present on the glacier at the beginning of the 1966 hydrologic year.

Ice Balance

The snowpack in the basin on May 18, 1966 (pl. 5*B*), averaged 0.82 m water equivalent ($\bar{b}_m(s)$, table 6). This was 5–10 times more than the winter snowfall recorded at several weather stations in the Copper River basin to the south and the Tanana River valley to the north. The relation of late-winter snow balance to altitude (pl. 5*C*) cannot, however, be interpreted to be the same as the change of precipitation with altitude. Some of the increase with altitude is caused by storms which rain on the lower elevations but snow higher up. In addition, a large part of the snow which falls high in Gulkana Glacier basin is blown away by strong south winds; this snow either evaporates or is redeposited on Canwell Glacier to the north. Little or no snow is blown into Gulkana Basin by these same winds because no high windblown mountains are present immediately to the windward south side of the basin. The average snowpack density on May 18 was about 0.36 megagrams per cubic meter.

The snowpack reached its maximum balance, $\bar{b}_x$, about May 20, 1966, at the lowest point in the basin; maximum balance was probably reached about July 15, 1966, high in the basin. Possibly the highest areas accumulated snow continually throughout the year. About 55 percent of the snowpack measured May 18 on the

TABLE 5.—*Instrumentation at Gulkana Glacier during the 1966 hydrologic year*

[Continuous record ———, estimated record ----, occasional measurement • •]

Station	Altitude (meters)	Measurement	1966 Hydrologic year												
			1965				1966								
			Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.
I	1,140	Streamflow													—
A	1,400	Snow and ice balance	•								•	•		•	•
B	1,670	Snow and ice balance	•								•			•	••
C	1,940	Snow and ice balance									•				•
Basin photographed											—	•	••	•	••
Basin occupied by personnel			•								—	—		—	—

TABLE 6.—*Snow and ice balances, Gulkana Glacier, 1966 hydrologic year*

[Parameter values and errors in meters except where indicated. Date: Hydrologic year, Oct. 1, 1965 (t_0), to Sept. 30, 1966 (t_1)]

	Glacier		Basin		Date	Term	Explanation
	Value	Error	Value	Error			
Parameter for annual mass balance							
$\bar{b}_a$	0.06	0.19	0.03	0.15	Hydrologic year	Annual balance	Net change in glacier mass from t_0 to t_1; approximately equal to the difference between precipitation as snow and melt-water runoff for one hydrologic year.
Parameters relating annual and net mass balances							
$\bar{b}_a(fi)$	−0.21	0.10	−0.19	0.10	Hydrologic year	Firn and ice annual balance	Change in mass of firn and ice from t_0 to t_1.
$\bar{b}_0(s)$	.10	.05	.08	.04	Oct. 1, 1965	Initial snow balance	New snow accumulated on summer surface (ss_0) at t_0.
$\bar{b}_0(i)$	0	--------	0	--------	--------	Initial ice balance	Ice melt in the ablation area after t_0 and before ice melt begins the following spring; measured by ablation stakes.
$\bar{b}_1(s)$	.37	.10	.30	.08	Sept. 30, 1966	Final snow balance	New snow accumulated on summer surface (ss_1) at t_1.
$\bar{b}_a(i)$	.76	.12	.57	.10	Hydrologic year	Annual ice balance	Ice and firn melt in the ablation area for the hydrologic year.
$\bar{b}_a(f)$	.55	.10	.38	.08	do	Annual firnification	New firn formed on the glacier from t_0 to t_1; not definable if snowmelt continues after t_1.
Parameter for snow accumulation							
$\bar{b}_m(s)$	1.00	0.15	0.82	0.15	May 18, 1966	Late-winter snow balance	Balance measured to the summer surface (ss_0) in late winter or spring; measured in pits or by probing.
Parameters for glacier dimensions							
S	[1] 21.8	0.2	31.6	0.2	Sept. 30, 1966	Area	Area of glaciers, excluding the ice-cored moraines.
AAR	[2] .59	[2] .02	[2] .41	[2] .02	do	Accumulation-area ratio	A rough index of annual balance, measured at time t_1, neglecting new snow overlying ss_1.
ELA	1770	30	--------	--------	do	Equilibrium-line altitude	Do.
ΔL	−44	10	--------	--------	Hydrologic year	Advance or retreat	Average horizontal-distance change of terminus in direction of glacier flow.

[1] Square kilometers.
[2] Dimensionless.

glaciers did not melt, and this residual snow was identifiable as firn in September 1966. High in the eastern cirques, summer avalanches and storms caused considerable accumulation. Small patches of firn remained on the glacier at altitudes as low as 1,600 m. New snow definitely began accumulating by August 15 at 1,940 m and by September 15, 1966, at 1,400 m.

Ablation of glacier ice began early in May 1966 from snowfree (windblown) morainal ridges low on the glacier (pl. 5*A*). The melting became most rapid in July and continued until the ice became fully snow covered in late September 1966 (pl. 5*D*). Ice ablation measurements were made on September 23, after the ice was snow covered. Some ablation of the previous year's firn occurred in July and August, and this is included as glacier-ice ablation. Near the glacier terminus, a maximum ablation of 4.0 m occurred on bare ice; however, a 10–20 cm layer of supraglacier moraine reduced the annual ice ablation locally to about 2.5 m. Isolated areas of ice ablation occurred up to an altitude of 2,300 m. The amount of glacier-ice ablation exceeded the formation of firn, so the annual firn and ice balance was −0.21 m (pl. 5*E*, table 6).

Heavy snowfall in August and September covered the entire basin with new snow by September 20, 1966 (pl. 5*D*). From measurements taken September 23, it is estimated that the basin had an average final snow balance, $\bar{b}_1(s)$, of 0.30 m at the end of the hydrologic year. This heavy snow accumulation exceeded the initial snow balance and more than made up for the basin's negative firn and ice balance, $\bar{b}_a(fi)$, of −0.19 m; as a result the basin gained 0.03 m storage, $\bar{b}_a$, during the 1966 hydrologic year (table 6).

The continuous formation of kettle holes and the exposure of ice after mudflows indicates that ablation from the ice-cored moraine was occurring. The amount of melting (pl. 5*E*) was estimated to be greater than 1 m in localities having 20 percent bare ice exposed and 0.1– 0.5 m where the ice is buried below alluvium and till.

The terminus of the glacier is stagnant and consists of nearly level ridges of ice. Ablation of some of this ice caused a terminal retreat of about 44 m. The glacier has retreated 530 m since 1954 and about 2 km since 1910 (Reger, 1964, pl. 1), an average retreat rate of 36 m per year.

Precipitation and Runoff

The annual precipitation and runoff during the 1966 hydrologic year are estimated to be from 1.5 to 2 m on the basis of the data for snow and ice balance and on an estimate of 0.5 m of additional ungaged precipitation. The precipitation and runoff from the low forested regions of interior Alaska are only 10–20 percent of the estimated values for Gulkana Basin; therefore, the glacier-covered Alaska Range contributed an important part of the water runoff from the area.

WOLVERINE GLACIER

Field Program

Wolverine Glacier was selected as a site for a combined balance study of the IHD in 1966, but permanent installations were not installed until 1967. Only limited data for ice and snow balance are available for the 1966 hydrologic year (table 7, pl. 6*A*).

Climate

The weather recorded at Seward, Alaska, 45 km southwest of Wolverine Glacier, departed significantly from the long-term means. Precipitation from October 1965 through July 1966 was 0.76 m, which was only about 50 percent of that normally recorded. The August and September precipitation totaled 0.81 m and was about 200 percent of that usually recorded. Temperatures were in the near-normal range. Precipitation data from Wolverine Glacier, as yet unpublished (obtained in subsequent years), indicate that a close correlation exists between Seward and Wolverine Glacier precipitation. Therefore, the 1966 ice balances on Wolverine Glacier were probably below average, like the precipitation at Seward.

Ice Balance

The late-winter snow balance, $\bar{b}_m(s)$, on Wolverine Glacier was measured at eight points from the head to the terminus on April 23, 1966. Although no basin snow map was constructed, the altitude distribution of the point data is shown (pl. 6*B*) and was used to compute the glacier and basin average late-winter snow balances of 1.83 and 1.55 m, respectively (table 8). No data are available for Wolverine Glacier prior to April 1966, so the changes in snow balance (pl. 6*C*) for the first half of the 1966 hydrologic year are only estimated.

The basin was often photographed from aircraft during the summer (table 7). These photographs provide the basis for most of the results reported here. A small number of stakes, pits, and probing points provided some point data on the snow and ice balance.

Part of the terminus was blown bare of snow on April 23, 1966, where 8.0 m water equivalent of ice melted from the glacier during the 1966 hydrologic year. An additional 0.25 m of ice ablation at the terminus occurred after the end of the hydrologic year. The previous year's firn was exposed between the transient snowline and the glacier ice at the end of the ablation season. Ablation of this firn is included in table 8 as part of the annual ice balance, $\bar{b}_a(i)$. The average equilibrium line reached an altitude of 1,250 m. Isolated ablation areas occurred at altitudes as high as 1,500 m.

Approximately 30 percent of the snow on the glacier in April remained in September 1966 and was buried under the new snow. The residual snow thus became the increment of new firn in the accumulation area, $\bar{b}_a(f)$. Several small perennial snowfields at an altitude of 800 m were the lowest areas of firn formation. These deposits are due to wind drifting.

New snow began accumulating high in the basin by late August 1966. A series of severe storms crossed the region in September, and by the end of the hydrologic year approximately 0.30 m water equivalent of new snow covered the glacier (pl. 6*C*). The lower edge of the new snow was at an altitude of 1,000 m on the glacier and at 1,100 m on the mountain slopes by September 21, 1966.

The amount of ice and old firn ablation, $\bar{b}_a(i)$, exceeded the formation of new firn, $\bar{b}_a(f)$, during the 1966 hydrologic year. The firn and ice annual balance, $\bar{b}_a(fi)$, thus

TABLE 7.—*Instrumentation at Wolverine Glacier during the 1966 hydrologic year*

[Continuous record ——— , estimated record ---- , occasional measurement ••]

Station	Altitude (meters)	Measurement	1966 Hydrologic year												
			1965			1966									
			Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.
A	650	Snow and ice balance							•		•			•	•
B	1,110	Snow and ice balance							•		•				•
C	1,350	Snow and ice balance							•						•
Basin photographed									•		•	• •	•	• •	•
Basin occupied by personnel									•		—			•	—

TABLE 8.—*Snow and ice balances, Wolverine Glacier, 1966 hydrologic year*

[Parameter values and errors in meters except where indicated. Date: Hydrologic year, Oct. 1, 1965 (t_0), to Sept. 30, 1966 (t_1)]

	Glacier		Basin		Date	Term	Explanation
	Value	Error	Value	Error			
Parameter for annual mass balance							
$\bar{b}_a$	−0.26	0.34	−0.16	0.26	Hydrologic year	Annual balance	Net change in glacier mass from t_0 to t_1; approximately equal to the difference between precipitation as snow and melt-water runoff for one hydrologic year.
Parameters relating annual and net mass balances							
$\bar{b}_a(fi)$	−0.36	0.23	−0.26	0.18	Hydrologic year	Firn and ice annual balance	Change in mass of firn and ice from t_0 to t_1.
$\bar{b}_0(s)$	.20	.20	.15	.15	Oct. 1, 1965	Initial snow balance	New snow accumulated on summer surface (ss_0) at t_0.
$\bar{b}_0(i)$	.05	.05	.03	.03	Oct. 1 to 14, 1965	Initial ice balance	Ice melt in the ablation area after t_0 and before ice melt begins the following spring; measured by ablation stakes.
$b_1(s)$	.30	.10	.25	.10	Sept. 30, 1966	Final snow balance	New snow accumulated on summer surface (ss_1) at t_1.
$\bar{b}_a(i)$	.99	.15	.72	.12	Hydrologic year	Annual ice balance	Ice and firn melt in the ablation area for the hydrologic year.
$\bar{b}_a(f)$	.63	.20	.46	.15	do	Annual firnification	New firn formed on the glacier from t_0 to t_1; not definable if snowmelt continues after t_1.
Parameter for snow accumulation							
$\bar{b}_m(s)$	1.83	0.30	1.55	0.30	Apr. 23, 1966	Late-winter snow balance	Balance measured to the summer surface (ss_0) in late winter or spring; measured in pits or by probing.
Parameters for glacier dimensions							
S	[1] 18.0	0.2	24.9	0.2	Sept. 30, 1966	Area	Area of glaciers, excluding the ice-cored moraines.
AAR	[2] .56	[2] .05	[2] .40	[2] .05	do	Accumulation-area ratio	A rough index of annual balance, measured at time t_1, neglecting new snow overlying ss_1.
ELA	1250	30	--------	--------	do	Equilibrium-line altitude	Do.
ΔL	−4	2	--------	--------	Hydrologic year	Advance or retreat	Average horizontal-distance change of terminus in direction of glacier flow.

[1] Square kilometers.
[2] Dimensionless.

was −0.36 m over the glacier (table 8, pl. 6*D*). The total mass net balance, $\bar{b}_n$, was very likely equal to the firn and ice annual balance.

Runoff

The annual runoff from the heavily glacierized Nellie Juan River basin, to which Wolverine Creek is tributary, has been gaged from 1960 to 1965 and averaged 2.1 m per year. The 1966 annual loss in ice storage from Wolverine Basin, 0.26 m, was equal to 12 percent of the 1960–65 annual runoff of the Nellie Juan River and illustrates to what extent long-term loss in ice storage may affect the larger rivers of the Kenai Peninsula. During a drought year the glaciers could contribute nearly all the riverflow.

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