The Geology of North America
Vol. G-1, The Geology of Alaska
The Geological Society of America, 1994

Chapter 27

Late Cenozoic glaciation of Alaska

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INTRODUCTION

Glaciers cover only about 5 percent of Alaska today, but they spread over as much as half of the state during the most widespread advances of the late Cenozoic (Fig. 1). Both modern and ancient glaciers have been most extensive in southern Alaska, where they were close to moisture sources in the North Pacific Ocean and the Gulf of Alaska (Fig. 2). Glaciers were smaller farther to the north because the nearest water bodies were ice-covered for much of the year and broad continental platforms were emergent during times of glacioeustatic sea-level lowering. The repeated glaciations of late Cenozoic time had great impact even on nonglaciated parts of Alaska, where they caused formation of proglacial lakes, construction of outwash terraces, loess deposition, and isostatic depression of coastal lowlands. Because the Alaskan glacial record interrelates with so many other physical processes and climate-related features, it provides the fundamental stratigraphic framework for the late Cenozoic history of much of the state.

The positions of readily accessible glaciers along the southeastern Alaskan coast were recorded during 18th century explorations of La Perouse, Cook, Vancouver, and others, and detailed studies of those glaciers and their fluctuations began near the close of the 19th century (Reid, 1896; Gilbert, 1904; Tarr and Martin, 1914). Compilations of the statewide glacial record have been published by Capps (1932), Péwé and others (1953), Karlstrom and others (1964), Coulter and others (1965), and Péwé (1975).

Ancient as well as modern glaciers in Alaska were primarily alpine in character. Glaciers that originated in mountains of northern and central Alaska were free to expand individually until they attained equilibrium, but the much larger glacier systems south of the crest of the Alaska Range coalesced to form a vast complex that constituted the northern part of the Cordilleran Ice Sheet (Hamilton and Thorson, 1983; Clague, 1989).

The Alaskan glacial record has been strongly influenced by tectonism. Volcanoes have been active throughout late Cenozoic time along zones of active plate convergence and in some intraplate areas, and lava flows and tephra layers are important means of dating and correlating Alaskan glacial advances. Tectonic deformation of drift sheets was especially active in southern parts of the state, severely disrupting glacial records but also providing a mechanism for correlating drifts on the basis of comparable amounts and/or types of deformation. Conversely, well-dated glacial successions can provide a framework for assessing long-term rates and recurrence intervals of faulting, volcanic eruptions, and other tectonic activities.

Glacial records from northern and southern Alaska differ in fundamental respects. The glacial successions in northern and central parts of the state extend back into late Tertiary time and commonly consist of five to seven distinct major glacial episodes. Farther to the south, glacial records tend to be shorter or more fragmentary because of (1) erosion or burial of older drifts during the latest Pleistocene ice advance, (2) dissipation of sediments where glaciers flowed into the sea, (3) obliteration of glacial features by tectonic activity, and (4) accelerated erosion and deposition owing to high relief caused by tectonism. The longer, climatically responsive records of northern and central Alaska therefore provide a baseline for interpreting less complete glacial records elsewhere in the state.

Most previous mapping of glacial deposits in Alaska has utilized formally named geologic-climate units, such as "glaciations" and "stades," as defined by the Code of Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature, 1970). These units subsequently were abandoned by the North American Commission on Stratigraphic Nomenclature (1983, p. 849) because "inferences regarding climate are subjective and too tenuous a basis for the definition of formal geologic units." I have retained the term "glaciation" but use it informally in cases where a glaciation was formally named prior to the recent revision of the stratigraphic code. I also informally use the terms "drift sheet" (allostratigraphic units) and "glacial episode" and "glacial phase" (diachronic units). For a more thorough discussion of stratigraphic nomenclature, see Hamilton and others (1986b).

Much of this chapter is based on regional summaries in Hamilton and others (1986a). Emphasis is on studies carried out since the major statewide compilation by Péwé (1975). Space constraints prohibit discussion of the Holocene glacial record and of modern glaciation in Alaska, which have been reviewed recently by Calkin (1988) and Krimmel and Meyer (1989), respectively.

Time divisions employed here generally follow those of

Hamilton, T. D., 1994, Late Cenozoic glaciation of Alaska, *in* Plafker, G., and Berg, H. C., eds., The Geology of Alaska: Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-1.

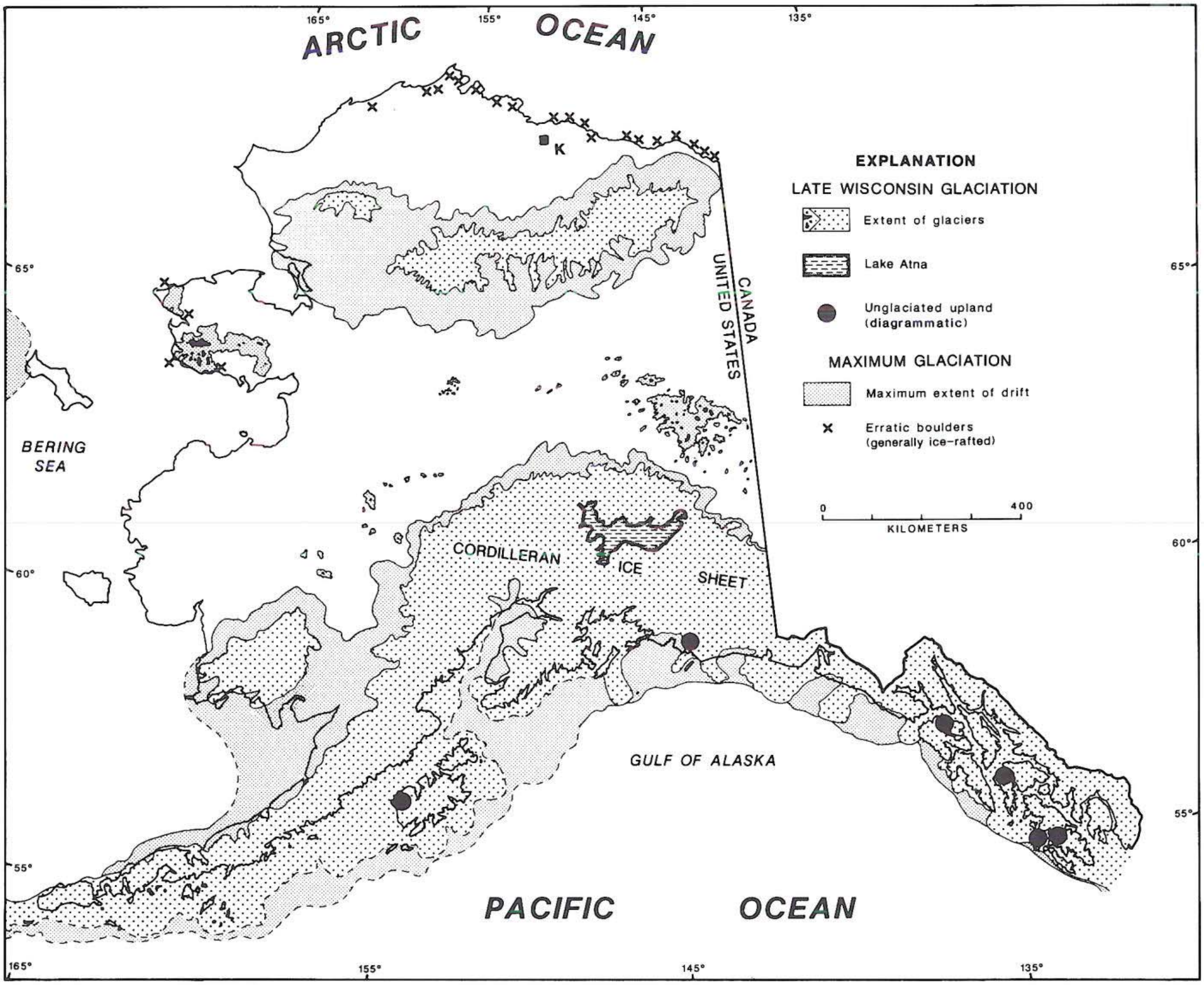

Figure 1. Maximum known advances of glaciers of late Wisconsin age and limits of older drift in Alaska (shown by dashed lines where inferred). Erratic boulders along Arctic Ocean margin represent the Flaxman Member of the Gubik Formation. K = Kuparuk gravel of Carter (1983).

Richmond and Fullerton (1986). The Pliocene-Pleistocene boundary is set at 1.65 Ma by international agreement. The early/middle Pleistocene boundary is the boundary between the Matuyama Reversed- and Brunhes Normal-Polarity Chrons, which has a generally accepted age of about 730 ka (Mankinen and Dalrymple, 1979), but which may be as old as 790 ka (Johnson, 1982). The boundary between the middle and late Pleistocene is the beginning of marine-oxygen-isotope stage 5, which Martinson and others (1987) have dated at about 125 ka. The age assigned to the Pleistocene-Holocene boundary, 10 ka, is the "nice round number" of Hopkins (1975).

For semantic simplicity, I use the general term "diamict" (Harland and others, 1966) for till-like deposits rather than the more familiar "diamicton" and "diamictite" for unlithified and indurated deposits, respectively. Till-like deposits in Alaska include abundant slightly to moderately lithified sediments of late Tertiary to early Pleistocene age, with properties intermediate between those of diamicton and diamictite.

TERTIARY GLACIATION

Glacial deposits of inferred Tertiary age occur widely in Alaska (Table 1). Many of these deposits have been faulted, tilted, uplifted, or otherwise tectonically deformed; other deposits are related to ancient valley or drainage systems that subsequently were abandoned. Depositional morphology seldom has been pre-

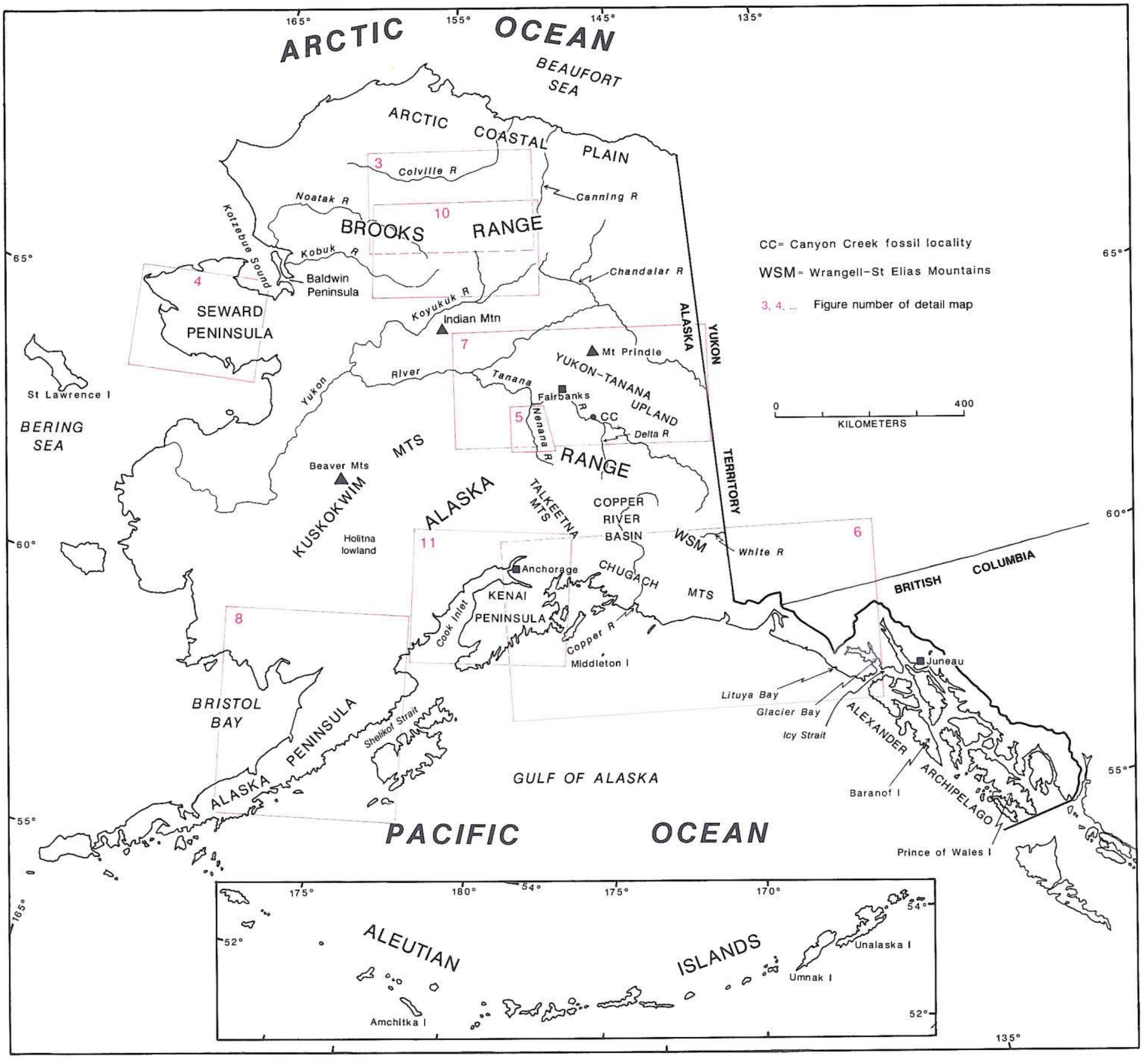

Figure 2. Index map of Alaska, showing locations of detailed maps and other localities mentioned in text.

TABLE 1. CORRELATION OF ALASKAN GLACIAL DEPOSITS

Age/Event	Arctic Coastal Plain[1]	Brooks Range[2]		Seward Peninsula[3]	Yukon-Tana Upland[4]	Kuskokwim Mountains[5]	North-Central Alaska Range[6]	
Late Wisconsin		Itkillik II	Late Phase 13-11.5 ka Main phase 24-13 ka	● 11.5 ka Mount Osborn (Esch Creek)	Salcha (Convert)	Tolstoi Lake	Riley Creek (Donnelly)	IV 10.5–9.5 ka III 12.8–11.8 ka II 15–13.5 ka I 26–17 ka
Penultimate Glaciation	Flaxman Member of Gubik Formation	Itkillik I	● >40 ka Phase B Phase A	● >40 ka Salmon Lake (Mint River) Stewart River*	Eagle (American Creek)	Bifurcation Creek	Healy (Delta)	(Late phase) (Early phase)
Middle Pleistocene		Sagavanirk-Tok River	Late phase Early phase	✧ 0.47 ± 0.05 Ma Nome River	Mount Harper (Little Champion)		Lignite Creek Bear Creek*	
			❑ 0.7 Ma	❑ 0.7 Ma				
Early Pleistocene		Anaktuvuk River (Sleepy Bear)		Sinuk	Charley River (Prindle)	Beaver Creek	Browne	
				❑ 2.2 Ma			Teklanika River	
Late Pliocene	Outwash	Gunsight Mountain		Unnamed drift	Unnamed drift			
	2.4-3.5 Ma			▲ 2.8 Ma			▲ 2.8 Ma	
Older Tertiary	Kuparuk gravel						Nenana gravel	

served, but glacial origin of the deposits generally is indicated by (1) erratic boulders, (2) striated and faceted clasts, or (3) ice-rafted sand and stones in marine deposits.

Arctic Coastal Plain

The informally termed "Kuparuk gravel" (Carter, 1983), crops out east of the Colville River (see Fig. 1). It overlies Paleocene deposits and is truncated by a bluff that formed during a late Pliocene marine transgression (Carter and others, 1986b). It also is cut by fluvial terraces that are correlated with the late Tertiary Gunsight Mountain glacial interval of Hamilton (1979). Clasts as much as 1.5 m in diameter are common in the gravel. They are composed of resistant rock types common in nearby parts of the Brooks Range, suggesting that a Tertiary ice advance from that source must have reached to within 30 km of the modern coast (Carter, 1983).

Central Brooks Range

Intensely eroded drift of Gunsight Mountain age has been recognized beyond both flanks of the Brooks Range (Hamilton, 1979, 1986a, 1986b). Distinctive rock types such as the middle Paleozoic Kanayut Conglomerate (Nilsen and Moore, 1984) indicate sources deep within the range, and the lobate drift border demonstrates that some of the principal mountain valleys already were major conduits for alpine trunk glaciers by Gunsight Mountain time.

Drift remnants of Gunsight Mountain age are traceable for at least 330 km along the foothills north of the Brooks Range, where they define a series of large piedmont lobes that extended as much as 58 km beyond the range front (Fig. 3). Till remnants or erratic boulders occupy erosion surfaces that generally stand about 100 m above modern drainages and are associated with a regional terrace system along the Colville River. Rare arcuate

TABLE 1. CORRELATION OF ALASKAN GLACIAL DEPOSITS (continued)

Northwestern Alaska Range[7]	Wrangell and St. Elias Mountains[8]	Upper Cook Inlet[9]	Bristol Bay[10]	Gulf of Alaska[11]
Farewell 2	● 11.3 ka Macauley (Kluane) ● 29 ka	Elmendorf moraine 13.7-11.7 ka Naptowne	Brooks Lake: Iliuk Newhalen Iliamana (Alegnagik) Kvichak (Okstukuk)	Yakataga Formation: Eurhythmic ● 12,430 ± 100 Raven House
● 34,340 ± 1940 Farewell I (2 phases)	● >49 ka xxxxxxxxxxxxxxx Icefield	● >49 ka Knik	● >40 ka Mak Hill (Manokotak and Gnarled Mountain moraine belts)	❑ 40 ± 10 ka High terrace (drift unit) ● >72 ka
Selatna		Eklutna	Johnston Hill (Ekuk; Halfmoon Bay?)	
Lone Mountain	Unnamed tills	Caribou Hills*	Unnamed drift (Nichols Hill)	Intense ice rafting
		Mount Susitna Diamicts		
Big Salmon Fork	▲ 2.7 ± 0.6 Ma Diamicts			❑ 2.5 Ma
	▲ 8.4 ± 0.7 Ma Diamicts	Kaloa deposits (late Miocene and younger)		Minor ice rafting

Place names below headings refer to informally named drift units. ( ) = local name used elsewhere in region; * = correlation uncertain; xxx = Old Crow tephra; ● = radiocarbon age; ✧ = Ar-Ar age; ▲ K-Ar age; ❑ = estimated age (see text for explanation).

Principal sources of data: 1. Carter, 1983; Carter and others, 1986a; 2. Hamilton, 1986a; 3. Kaufman and Hopkins, 1986; Kaufman and Calkin, 1988; 4. Weber, 1986; 5. Kline and Bundtzen, 1986; 6. Thorson, 1986; Ten Brink and Waythomas, 1985; 7. Kline and Bundtzen, 1986; 8. Denton, 1974; Denton and Armstrong, 1969; 9. Reger and Updike, 1983, 1989; Schmoll and Yehle, 1986; 10. Detterman, 1986; Lea and others, 1989; 11. Mann, 1986; Plafker, 1981; Armentrout, 1983.

morainal forms lack primary relief; they are defined principally by boulder concentrations and arcuate stream courses. Drift having nearly featureless topographic expression is more common and generally is associated with concentrations of relict boulders in the beds of streams. Elsewhere, the former drift sheet is represented only by scattered boulders of the highly resistant Kanayut Conglomerate, which are most common where loess cover of Quaternary age has been partly or completely stripped by erosion.

South of the Brooks Range, drift provisionally correlated with glacial advances of Gunsight Mountain age forms a broad (4 to 5 km), arcuate, probable end moraine that extends for about 70 km along the south margin of the Koyukuk River basin (Hamilton, 1986a). The moraine bears a thick cover of loess and lacustrine silt of Quaternary age, but its glacial origin is demonstrated by concentrations of erratic boulders along the beds and banks of rivers.

A broad U-shaped valley remnant in the west-central Brooks Range, which contains small nested end moraines of early Pleistocene and younger age, is interpreted as having been eroded during Gunsight Mountain time. Alluvial-fan deposits on the valley floor beneath the nested moraines contain cones of an extinct spruce that also occurs in the late Pliocene Beaufort Formation of arctic Canada (Matthews, 1990) but is not known to occur in Pleistocene deposits. The stratigraphic setting of the fossiliferous fan deposits indicates a probable late Tertiary age for the Gunsight Mountain glacial interval.

The Seward Peninsula

Glacial deposits older than the middle Pleistocene Nome River drift are poorly differentiated on the Seward Peninsula (Kaufman and Hopkins, 1986). Evidence for probable Tertiary glaciation is strongst in the California River area (Fig. 4), where surface erratics and erratic-bearing diamicts occur beyond the limits of moraines and drift assigned to the Sinuk glaciation of probable early Pleistocene age. Although not definitive, stratigraphic relations suggest that the pre-Sinuk diamict underlies marine transgressive deposits of late Pliocene age and is younger than a 2.8-Ma basaltic lava flow (Kaufman and Hopkins, 1986; D. M. Hopkins, personal communication, 1990); it therefore probably is a late Pliocene deposit.

North-central Alaska Range

Erratic boulders in the Nenana Gravel, a thick (1,000 to 1,300 m), weakly indurated conglomerate of late Miocene and early Pliocene age, have been described by Carter (1980) and Thorson (1986). Near the Delta River valley, boulder-bearing beds in the upper part of the Nenana Gravel dip northeast at 40 to 60° and contain clasts as much as 2 m in diameter, which include lithologies foreign to their present drainage basins (Carter, 1980). The Nenana Gravel also coarsens upward in exposures near the Nenana River valley; boulders and faceted cobbles are common near the top of the unit (Thorson, 1986). Other boulders are residually concentrated along ridge crests by erosional stripping of the Nenana Gravel. A borrow pit near the Nenana River exposes bouldery diamicton that contains grooved, striated, and faceted clasts and underlies lignite-bearing oxidized sand assignable to the Nenana Gravel (Thorson, 1986). The diamict is interpreted as lodgment till of probable late Tertiary age. A potassium-argon age determination of 2.79 ± 0.25 Ma provides a minimum age for the Nenana Gravel (Thorson, 1986).

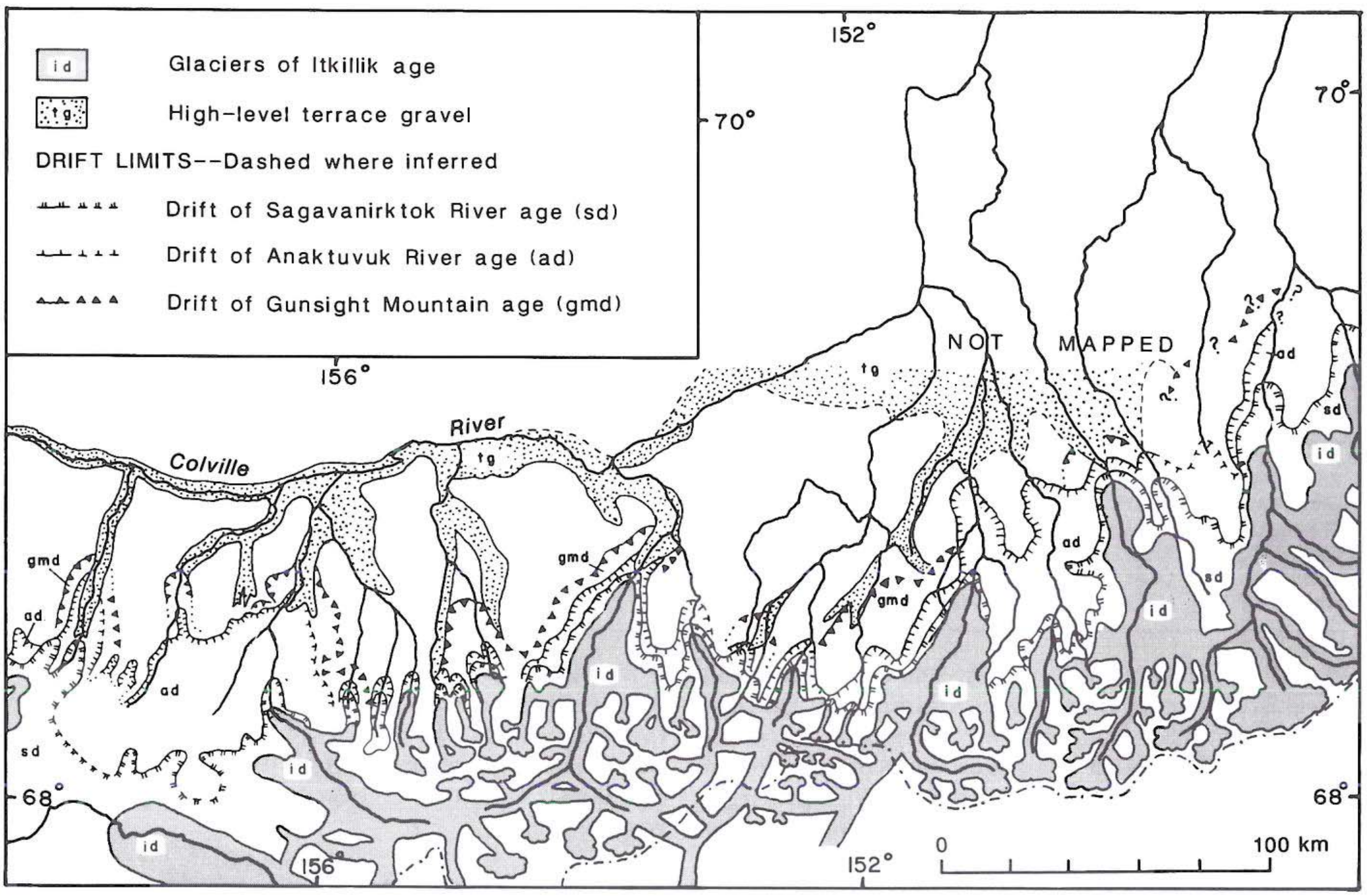

Figure 3. Glacier limits and high-level terrace gravel, northern Brooks Range and Arctic Foothills. Slightly modified from Hamilton (1986a).

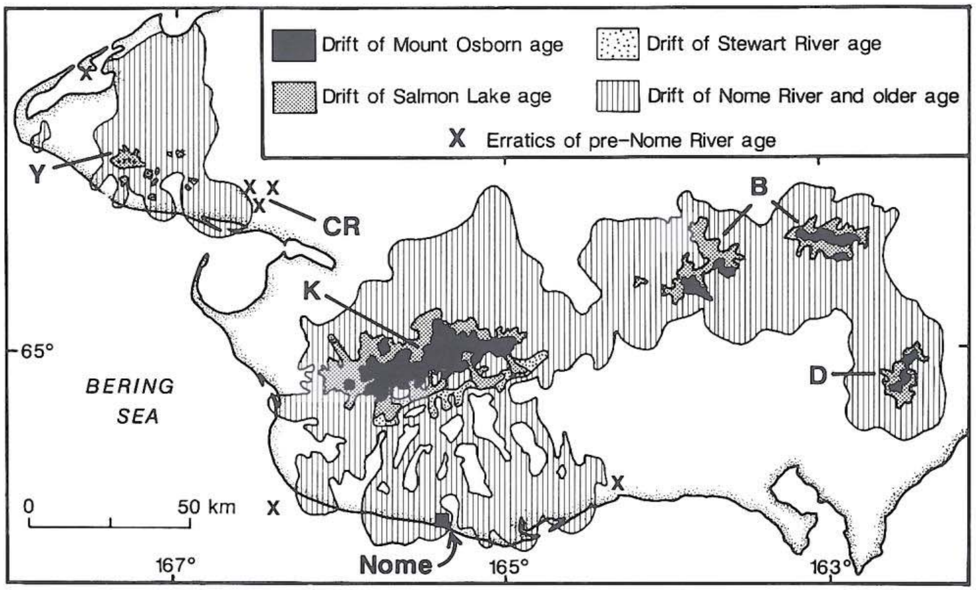

Figure 4. Seward Peninsula, showing ice advances of Mount Osborn, Stewart River, and greater age. B = Bendeleben Mountains, D = Darby Mountains, K = Kigluaik Mountains, Y = York Mountains. CR = California River locality. Modified from Kaufman and Hopkins (1986).

Teklanika River drift, a bouldery till of latest Tertiary or early Pleistocene age (Thorson, 1986), is exposed in a series of bluffs along the Teklanika River (Fig. 5). Diamict as much as 22 m thick contains striated and faceted boulders of Alaska Range lithologies in an oxidized sand-granule matrix and overlies the strongly oxidized, weakly indurated Nenana Gravel. The bluffs intersect the outer flank of a broad, gravelly morainal ridge with a thick cap of eolian sediments that defines a glacier lobe centered on the Nenana River valley and much larger than any younger glacier lobes. Because it generally was undeflected by foothills north of the Alaska Range, and because its deposits occur at altitudes much higher than those of younger drifts, the drift lobe probably was deposited before the Nenana River had incised its present valley and before erosion had etched the foothills into their present relief (Thorson, 1986).

Northwestern Alaska Range

Diamicts of the Big Salmon Fork glacial interval in foothills 15 to 25 km northwest of the Alaska Range consist of subrounded cobbles and boulders, many faceted and striated, in a silty sand matrix (Kline and Bundtzen, 1986). Associated conglomerates of poorly sorted, coarse cobble gravel may be outwash. Bedding dips northwest at 20 to 35° along the crests of hogbacks and homoclinal ridges. Clasts differ in composition from those of nearby younger glacial deposits, indicating significant changes in source areas since deposition of the diamicts and associated sediments.

Belts of lag boulders that lie beyond the limits of morainal topography in small creeks elsewhere in the foothills may also indicate very old and extensive glacier advances of late Tertiary age. Any moraines dating from these advances have either been destroyed by erosion or buried beneath younger sediments.

The Wrangell Mountains

Thick sequences of interbedded diamicts, fluvial sediments, and lava flows in fault-bounded blocks north of the Wrangell Mountains were initially described by Capps (1916) and later studied by Denton and Armstrong (1969). At least 12 separate diamicts in one of the structural blocks were identified as tillites of continental glaciers, because of (1) their massive, nonsorted, and nonstratified character; (2) presence of faceted, pentagonal, polished, and striated clasts that included allegedly exotic lithologies; and (3) identification of diagnostic glacial surface textures on sand grains (Denton and Armstrong, 1969). Potassium-argon ages on associated lavas indicated that the diamicts were deposited between 2.7 and at least 10 Ma. In one structural block, 10 diamicts underlie a flow dated at 8.4 ± 0.7 Ma, and flows dated at about 9.8 ± 0.3 to 10.2 ± 0.3 Ma are closely associated with diamicts in two other blocks.

Although Denton and Armstrong (1969) rejected alternative origins for the diamicts, Plafker and others (1977) argued that all or most of those units are actually lahars that originated on the Wrangell volcanoes. Rare glacially worked clasts in the deposits could have been shaped by alpine glaciers and later redeposited in lahars; Plafker and others (1977) found no clasts exotic to the upper parts of the contiguous Wrangell volcanoes. Recent studies by Eyles and Eyles (1989) have confirmed that the diamicts were probably emplaced as debris flows within alluvial-fan and fan-delta complexes in an actively subsiding fault-bounded basin. Eyles and Eyles (1989) concluded that the

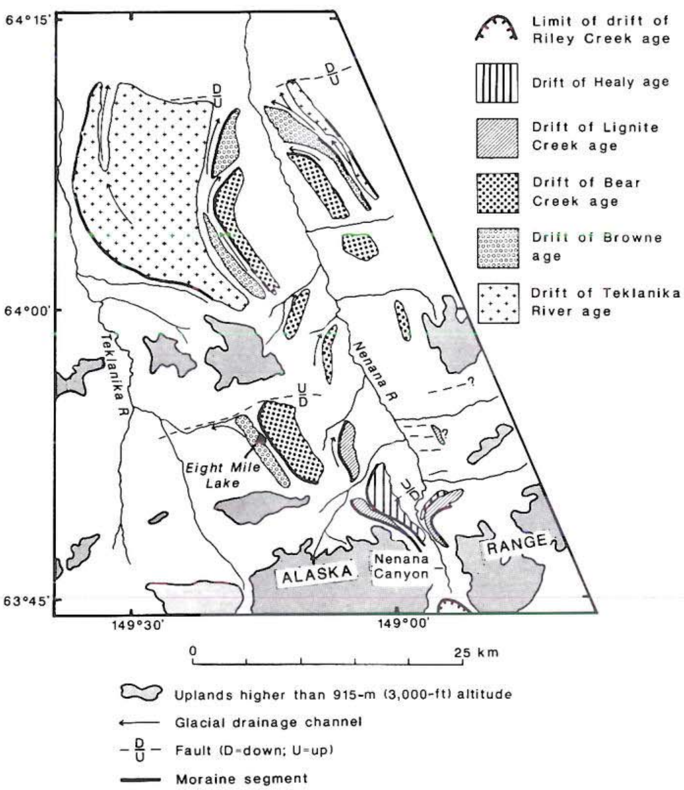

Figure 5. Major drift limits in the Nenana River valley, north-central Alaska Range. Modified from Thorson (1986).

glacially striated and faceted clasts within the diamicts were derived from deposits of local mountain glaciers rather than deposits of more extensive ice sheets, as proposed by Plafker and others (1977).

Upper Cook Inlet

Diamicts interpreted as tillites of late Miocene to Pliocene age occur in drill holes that penetrate the Beluga and Sterling Formations on the Kenai Peninsula (Boss and others, 1976, cited in Schmoll and others, 1984). Boss and others (1976) believe that tillite distribution was similar to that of late Pleistocene glacial deposits and, therefore, that the Kenai Mountains generated glaciers similar to the large piedmont lobes that covered most of the Kenai Peninsula during late Pleistocene time.

Diamicts are exposed over a distance of about 3.5 km at Granite Point on the west side of Cook Inlet (Schmoll and others, 1984, p. 15; Schmoll and Yehle, 1983). These diamicts, informally termed the "Kaloa deposits" (Schmoll and others, 1984), have a total stratigraphic thickness of about 110 m and are disconformably overlain by younger diamicts of Pleistocene age. The beds dip 1 to 6 degrees, and appear to occupy the limb of an anticline. Schmoll and others (1984) divided the Kaloa deposits into four major units: two thick massive diamicts with stone lines and two thinner units that contain interbeds of sandstone, siltstone, claystone, and (in the lower of the two units) coal. Pollen from the principal coal bed indicates that the diamicts probably are late Miocene in age.

The Mount Susitna glaciation (Karlstrom, 1964) is represented only by scattered erratics and by very old ice-scoured bedrock surfaces around upper Cook Inlet. Correlations with the central Alaska Range suggest a late Pliocene or early Pleistocene age for this glacial advance (Reger and Updike, 1989).

The Gulf of Alaska

Tertiary glaciation around the Gulf of Alaska is recorded in glaciomarine deposits exposed in the Yakataga Formation and recovered from marine piston cores. The Yakataga Formation, of Miocene through Holocene age, consists of interbedded marine, glaciomarine, and glaciofluvial deposits that locally exceed 5 km thickness (Plafker and Addicott, 1976; Plafker, 1981; Armentrout, 1983; Plafker and others, this volume, Chapter 12). It is exposed through an area of more than 30,000 km^2 along the Gulf of Alaska coast, including several islands on the continental shelf (Fig. 6). Glaciomarine deposition started either in the middle Miocene (Marincovich, 1989) or the late Miocene (Lagoe and others, 1989) in response to tectonic uplift of the Alaskan coast and to global climatic cooling, but its initiation was time transgressive owing to northwestward motion of the Yakutat terrane relative to the uplifted and glaciated coastal mountains (Plafker, 1987). Armentrout (1983) recognized three relatively cool paleoclimatic episodes associated with moderately abundant glacial sediments in lower parts of the Yakataga Formation; warmer conditions and normal marine sedimentation subsequently prevailed during most of the Pliocene. Diamicts later became abundant again at a time estimated by J. M. Armentrout (written communication, 1986) as about 2.5 Ma on the basis of correlations with onset of widespread glaciation in the North Pacific and North Atlantic regions (e.g., Carney and Krissek, 1986; Shackleton and others, 1984; Dowsett and Poore, 1990). Glaciomarine deposits continued to dominate the sediment record into the Holocene.

Piston cores taken by the Deep Sea Drilling Project (DSDP) on the deep sea floor in the Gulf of Alaska and adjoining parts of the North Pacific Ocean include abundant ice-rafted pebbles and sand. Recent revisions in diatom stratigraphy enable the onset of widespread ice rafting in the Gulf of Alaska to be dated at shortly after 2.48 Ma (Rea and Schrader, 1985). Rare erratic pebbles lower in the cores were deposited sometime prior to 3 Ma. Other DSDP cores taken north of the eastern Aleutian Islands record similar histories in which glaciers evidently first reached tidewater shortly after 2.48 Ma (Rea and Schrader, 1985).

Basal exposures of the Yakataga Formation on Middleton Island, which are described in a later section, are in part of latest Pliocene age (Mankinen and Plafker, 1987).

EARLY AND MIDDLE PLEISTOCENE GLACIATIONS

Glacial deposits of early to middle Pleistocene age are widespread in northern and central Alaska. High rates of Quater-

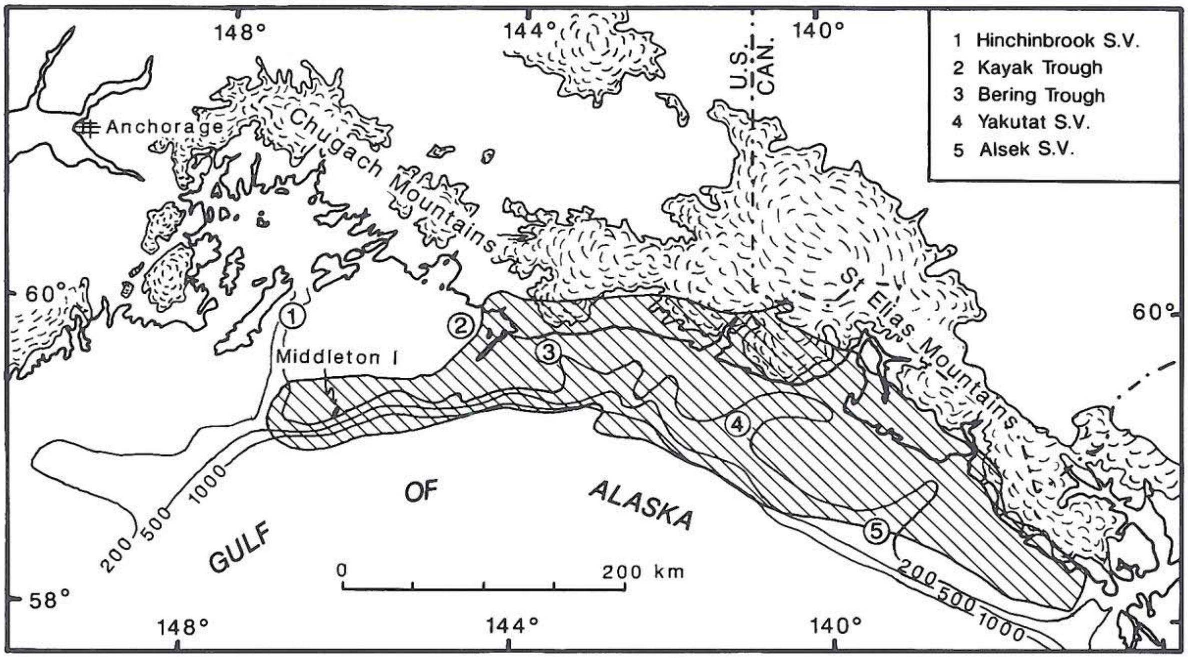

Figure 6. Extent of late Cenozoic Yakataga Formation across Gulf of Alaska continental shelf. Modern glaciers are shown by concentric dashed patterns. S.V. = Sea valleys crossing shelf. From Plafker and Addicott (1976).

nary tectonism farther to the south have caused comparable glacial deposits to be largely buried or obliterated by erosion, and drift successions older than late Pleistocene are uncommon in southern Alaska.

Two separate drift complexes are evident in many regions. The older, assigned to the early Pleistocene on the basis of paleomagnetic determinations, morphology, and position within regional drift successions, commonly was formed by glaciers that were much larger than any that formed subsequently. Early Pleistocene glaciation was followed by a long interval in which streams broadly eroded the drift sheets and incised glacial valleys to levels close to those of the present day. The younger drift complex, provisionally assigned to the middle Pleistocene, occupies modern valley floors. It forms extensive deposits in parts of northwestern Alaska, but elsewhere generally extends only short distances beyond the end moraines of late Pleistocene glaciers.

The central Brooks Range

Drift deposited during the Anaktuvuk River and Sagavanirktok River glaciations (Detterman and others, 1958; Keroher and others, 1966) is assigned respectively to the early and middle Pleistocene on the basis of limiting paleomagnetic determinations (Hamilton, 1986a) and recently revised correlations with the Seward Peninsula and Kotzebue Sound glacial sequences (Table 1).

During Anaktuvuk River time, valley glaciers in the central Brooks Range extended as much as 70 km beyond the margins of the range (Fig. 3). Glaciers flowing into the northern foothills formed paired lateral moraines; glaciers flowing south coalesced into broader piedmont lobes in the basins of the Kobuk, Koyukuk, and Chandalar Rivers. Large glaciers occupied the western Brooks Range and extended west into Kotzebue Sound. Some glacial valley systems of Anaktuvuk River age have been disrupted by tectonism, and deformation intensifies eastward toward the Canning River displacement zone of Grantz and others (1983). Drift of Anaktuvuk River age is distinguished from all younger glacial deposits by its association with ancient landscapes 50 m or more above the levels of modern streams and by the presence of tors, altiplanation terraces, and pediments on surfaces that were glaciated during Anaktuvuk River time but remained unglaciated thereafter.

In foothills north of the Brooks Range, drift of Anaktuvuk River age typically forms the outer sectors of massive compound lateral moraines several hundred meters thick (Hamilton, 1986a). End moraines typically have broad crests and gentle flanking slopes smoothly graded by solifluction (Table 2); they are covered by thick and continuous silt through which only a few large erratic boulders protrude. Drainage networks are well integrated. Kettle-like basins occur on some drift sheets, but they generally contain shallow thaw ponds that have developed in thick, postglacial, ice-rich silt. Drift sheets contain inset terraces 40 to 60 m above modern streams that continue downvalley into regional terrace systems 30 to 50 m high. Many kettles have developed on postglacial stream terraces, indicating that some glacier ice must have persisted for long periods after general deglaciation.

South of the Brooks Range, broad piedmont lobes of Anaktuvuk River age commonly terminated in proglacial lakes

TABLE 2. PHYSICAL CHARACTERISTICS OF PLEISTOCENE DRIFTS, NORTHERN AND CENTRAL ALASKA*

		Width of moraine crests (m)	Maximum flanking slopes (°)	Maximum boulder protrusion (cm)	Maximum depth of oxidation (m)
Central Brooks Range	Itkillik II	3–5 [2–3]	18–23 [16–22]	40–80 [20–30]	0.3
	Itkillik I	5–20 [5–15]	15–20 [14–20]	25–50 [≤20]	1.0–1.2
	Sagavanirktok River	75–200	2–3.5	100	>8
	Anaktuvuk River	500	1.2	10	>30
Seward Peninsula	Mount Osborn	24	14–16	70	0.2–0.3
	Salmon Lake	25	9–12	45	0.5
	Stewart River	50	6–7	30	
	Nome River	430	3.5	3	"Deep"
Yukon-Tanana Upland	Salcha	6	15–30		0.33
	Eagle	4–6	17–25		0.64
	Mount Harper	100	13–25		>0.53
	Charley River				>8
Nenana River Valley (Alaska Range)	Riley Creek	10–15	17–30		IV 0.2–0.3 III 0.3–0.5 II 0.7–0.8 I 1.0
	Healy	10–20	30		0.3
	Lignite Creek		15–20		15–20
	Bear Creek	<100	5–6		
	Browne	<50	15–25		

*Data from southern valleys of central Brooks Range are in brackets.
Note: Measurement techniques are not standardized and can vary between workers. See Kaufman and Calkin (1988) for recommended techniques.

(Hamilton, 1986a). Subdued, arcuate end moraines generally are covered by deposits of eolian and lacustrine silt more than 15 m thick. Knolls and ridges along many moraine crests were shaped by wave action; they contain washed gravel mixed with larger clasts derived from the original drift. River bluffs expose well-jointed, compact till and stony glaciolacustrine silt that are strongly oxidized along joint planes.

The Sleepy Bear glaciation, the oldest glacial advance recorded by Reger (1979) at Indian Mountain (see Fig. 2), probably is correlative with glacier advances of Anaktuvuk River age. End moraines having subdued morphology are overlapped by lacustrine sediments that formed when large glaciers from the Brooks Range crossed the Koyukuk River and created an extensive proglacial lake during Anaktuvuk River time (T. D. Hamilton, unpublished field mapping, 1987). The largest glacier of Sleepy Bear age occupied a valley floor that was incised 100 to 150 m prior to the next younger ice advance.

Glacier flow patterns of Sagavanirktok River age were similar to those of the Anaktuvuk River glaciation (Hamilton, 1986a), but glaciers were markedly smaller (Fig. 3). Paired lateral moraines in foothills north of the Brooks Range curve unbroken into valley centers, where they generally stand no more than 15 to

20 m above modern floodplains and extend only 3 to 8 km beyond the outermost terminal moraines of late Pleistocene age. Moraine crests are narrower than those of Anaktuvuk River drift, flanking slopes are steeper, and surface boulders are more abundant (Table 2). Kettles are rounded and partly filled by ice-rich silt; outwash terraces are capped by 1.5 to 2 m of frost-churned silty gravel. Outwash trains from late Pleistocene ice advances farther upvalley commonly contain kettles where they intersect end moraines of Sagavanirktok River age, indicating that subsurface glacier ice must still have been present when the younger glaciers advanced.

Glaciers of Sagavanirktok River age filled structural troughs along the south flank of the Brooks Range that had formed during a long interval of tectonism and nonglacial erosion that followed the Anaktuvuk River glaciation (Hamilton, 1989); they commonly terminated in lakes (Hamilton, 1986a). Drift is not as broadly eroded as is Anaktuvuk River drift, and it occurs close to modern valley floors. Some river bluffs expose compact, jointed, sandy to gravelly till, but silty glaciolacustrine diamict predominates elsewhere. Multiple end moraines and stratigraphic relations suggest two major ice advances in some valleys.

Till of Sagavanirktok River age, exposed in a bluff along the Kobuk River, overlies interglacial deposits that are magnetically reversed at the base but magnetically normal above (Hamilton, 1986a); the deposits probably span the boundary between the Matuyama Reversed- and Bruhnes Normal-Polarity Chrons. Because this part of the Kobuk River valley had been eroded previously during Anaktuvuk River time, the paleomagnetic data suggest that the Anaktuvuk River and Sagavanirktok River ice advances are of early and middle Pleistocene age, respectively.

The Seward Peninsula and Kotzebue Sound

Drift deposits of the Sinuk glacial interval, Nome River glaciation, and Stewart River glacial interval have been described by Kaufman and Hopkins (1986), who considered them to be of late Tertiary, early Pleistocene, and middle Pleistocene age, respectively. However, recent radiometric age determinations and amino-acid and paleomagnetic analyses suggest that the Sinuk drift may be of early Pleistocene age and that the succeeding Nome River advance probably took place during middle Pleistocene time (Table 1).

Sinuk drift consists of: (1) surface deposits with little primary glacial relief, (2) isolated erratic boulders, and (3) subsurface drift in excavations, natural exposures, boreholes, and marine seismic profiles. Surface exposures generally are limited to erratic boulders that are concentrated in the beds of streams, lie scattered on bedrock, or are exposed by placer mining. Tors and altiplanation terraces occur in upland areas glaciated during the Sinuk glacial interval, but these features have not been observed on surfaces covered by younger ice advances (D. M. Hopkins, oral communication, 1990). This relation is similar to that on surfaces glaciated during Anaktuvuk River time in the Brooks Range, suggesting that the Sinuk and Anaktuvuk River advances could be correlative.

Drift of the succeeding Nome River ice advance extends well beyond the major glaciated uplands of the Seward Peninsula (Kaufman and Hopkins, 1986; Kaufman and others, 1988). Glaciers were particularly extensive in the southwest part of the peninsula, where they reached the coast around Nome (Fig. 4). This distribution pattern suggests that glaciers received considerably more nourishment during Nome River time than during subsequent glaciations. Much of the Bering platform may have remained submerged during glacier growth, either due to tectonism (Hopkins, 1973) or because local glaciation began before onset of global sea-level lowering (J. Brigham-Grette, written communication, 1990). Morainal ridges of Nome River age generally are broad and smooth, with gentle slopes and thick, continuous loess cover (Kaufman and Calkin, 1988; Table 2).

The Baldwin Peninsula in Kotzebue Sound consists of marine, estuarine, glaciomarine, and glacial sediments that were deformed by glaciers flowing westward out of the Kobuk River valley and the Selawik River valley, which parallels the Kobuk valley to the south, and possibly southward out of the Noatak valley (Huston and others, 1990). The Baldwin Peninsula moraine is correlated with moraines of Nome River age on the Seward Peninsula by Brigham-Grette and Hopkins (1989) because of morphologic similarities and because both drift complexes represent the last extensive glaciation around the east margin of the Bering and Chukchi Seas. If the Baldwin Peninsula moraine were formed by glaciers that filled valleys of the western Brooks Range, it may correlate with the Anaktuvuk River glaciation as inferred by Hamilton (1986a, b) and by Kaufman and Hopkins (1986). However, if the moraine were formed by local ice caps in the lower Kobuk–Selawik River area, it more likely would correlate with the Sagavanirktok River glaciation of the central Brooks Range as concluded earlier by Hamilton and Hopkins (1982). Glaciers of Sagavanirktok River age in the central Kobuk River valley discharged into a large glacial lake (Hamilton, 1984) that may have formed behind an ice dam farther down the Kobuk River.

Age estimates for the Nome River glaciation are based on radiometric age determinations, paleomagnetic measurements, and amino-acid geochronology. Glacial deposits north of the Bendeleben Mountains on the Seward Peninsula are overlapped by a magnetically normal basaltic lava flow for which a potassium-argon age of 0.81 ± 0.08 Ma had been reported (Kaufman and Hopkins, 1986), but recent Ar-Ar micrograin analyses indicate contamination by an older phase of inherited argon (D. S. Kaufman, written communication, 1991). By excluding the older phase, D. S. Kaufman and R. C. Walter have determined a provisional age of 470 ± 190 ka for the basalt. Fossil molluscan shells from marine transgressive deposits that underlie Nome River till near Nome have amino-acid ratios that suggest a middle Pleistocene age (Kaufman and others, 1989b) of somewhere between 290 and 580 ka depending on the paleotemperature history of the site (D. S. Kaufman, written communication, 1990). Till from the type locality for the Nome River glaciation is normally magnetized, confirming that this glacial advance took place during the Brunhes Normal-Polarity Chron, and therefore is no older than about 730 ka (D. S. Kaufman, written communication, 1990). The glacially deformed marine and glaciomarine units on Baldwin Peninsula also have normal magnetic polarity, and amino-acid ratios on fossil molluscan shells suggest correlation with other marine deposits in northern Alaska that formed about 500 to 600 ka (Huston and others, 1990). Although the above data do not demonstrate an exact age for the Nome River glaciation, they show that it must be a middle Pleistocene event and that it most likely took place sometime between about 500 and 400 ka.

The Stewart River glacial interval is represented by smooth moraines with gravelly crests and thin loess 5 to 10 km beyond the limits of younger glacial deposits on the south side of the Kigluaik Mountains (Fig. 4; Kaufman and Hopkins, 1986; Kaufman and Calkin, 1988; Kaufman and others, 1988). Kettles generally are dry, stable, flat-bottomed depressions, but steep-sided and unstable dead-ice terrain on a Holocene terrace inset across one end moraine indicates that buried glacier ice may still be present locally. Relative-age criteria suggest that the Stewart River ice advance is closer in age to the drift of the penultimate glaciation than to the preceding, much more extensive ice advance of Nome River time. Its correlations with the Brooks Range glacial sequence are unclear.

Yukon-Tanana Upland

The glacial history of the Mount Prindle area, 85 km northeast of Fairbanks (Fig. 7), was described by Weber and Hamilton (1984). They correlate their oldest ice advance, the Prindle glaciation, with the Anaktuvuk River glaciation of the Brooks Range because of: (1) its great extent relative to all other ice advances, (2) major drainage changes and valley incisions since Prindle time, and (3) presence of tors on uplands last glaciated during Prindle time. They also correlate drift of the succeeding Little Champion glaciation, which occurs on modern valley floors, with drift of Sagavanirktok River age in the Brooks Range. Weber (1986) subsequently showed that this glacial succession extended throughout the Yukon-Tanana Upland. She correlated the oldest regional Pleistocene ice advance—the Charley River glacial episode—with the Prindle glaciation and the succeeding Mount Harper glacial episode with the Little Champion glaciation (Table 1).

Ice caps of Charley River age probably developed on all mountains above 900 m altitude. Outlet glaciers radiated to distances as great as 56 km and formed the most extensive glacial features preserved in the Yukon-Tanana Upland (Fig. 7). Evidence for the Charley River glacial episode consists primarily of concentrations of large erratic boulders at the downvalley limits of highly modified U-shaped troughs that radiate from the higher peaks. Possible moraines of Charley River age form solifluction-covered benches along some valley sides.

All mountains rising above 1,200-m altitude were glaciated during Mount Harper time (Weber, 1986). Valley glaciers extended as far as 29 km and reached altitudes of about 600 m (Fig. 7). Most end moraines of Mount Harper age occur in lower valleys and are largely vegetated (Weber, 1986). Their crests commonly are broad and flat, and they typically are fronted by broad outwash fans. Original moraine forms commonly are obscured by solifluction and talus deposits. Boulder surfaces are etched deeply, and many granite clasts have completely disintegrated. Multiple end moraines in many valleys indicate that the Mount Harper glacial episode probably consisted of at least two major ice advances (Weber, 1986).

The Kuskokwim Mountains

At least 13 isolated highlands within the Kuskokwim Mountains were sufficiently massive to generate alpine glaciers during Quaternary time (Kline and Bundtzen, 1986). Bundtzen (1980) recognized four major glacial advances in the Beaver Mountains. The oldest drift, assigned to the Beaver Creek glaciation, resembles early Pleistocene glacial deposits of the Seward Peninsula and the Yukon-Tanana Upland.

During Beaver Creek time, glaciers from a probable ice cap breached stream divides at high altitudes and formed long, planed upland surfaces (Kline and Bundtzen, 1986). Distribution of till and erratic boulders suggest that outlet glaciers advanced 14 to 26 km down each of the four major valleys that radiate from the center of the Beaver Mountains. Morainal landforms are absent, and evidence for former glaciation consists primarily of scattered patches of till, isolated erratic boulders, faint trim lines, and ice-marginal meltwater channels. The drift is covered by 1 to 3 m of loess that supports numerous thaw lakes.

North-central Alaska Range

Thorson (1986) recognized at least three glacial advances of probable early and middle Pleistocene age within the Nenana River valley (Fig. 5). From oldest to youngest, these are the Browne glaciation (Wahrhaftig, 1958) and the Bear Creek and Lignite Creek glacial episodes of Thorson (1986). The Browne glaciation probably is early Pleistocene in age, and the Lignite Creek event is late middle Pleistocene; the age assignment of the Bear Creek glacial episode is uncertain. The Teklanika River advance (discussed earlier in this chapter) could be either latest Tertiary or early Pleistocene in age.

The Browne glaciation was defined by Wahrhaftig (1958) on the basis of large, erratic boulders of granite on upland surfaces and in stream beds. Thorson (1986) later described probable paired end moraines of Browne age east and west of the Nenana River that are flanked by arcuate meltwater channels (Fig. 5). Original constructional forms are well defined, and ice-marginal channels and proximal outwash aprons also retain primary morphology (Table 2). A meltwater channel system east of the Nenana River grades into a broad fan of outwash gravel that is truncated either by a fault scarp or by a monocline about 100 m high.

The outer moraine of the Bear Creek drift sheet consists of arcuate, erratic-littered ridges east and west of the Nenana River that are flanked by marginal stream courses (Thorson, 1986). The ridges cross-cut moraines of Browne age and are inset within them. A correlative moraine segment near Eight Mile Lake contains bouldery gravel in which some clasts retain original glacial facets and grooves. Because of their cross-cutting and inset relations and because they appear to have higher surface relief and better internal drainage than drift of Browne age, the Bear Creek moraines probably represent a distinct younger glacial episode (Thorson, 1986). The Bear Creek glacier must have advanced down the Nenana valley when it was incised about 5 to 100 m below drift of Browne age and when the axis of the valley had shifted slightly westward.

The Lignite Creek glacial episode was named for a series of arcuate morainal ridges and intervening ice-marginal drainage channels close to the modern floor of the Nenana River valley (Thorson, 1986). These features are less subdued morphologically than drift of Bear Creek age. The ridges of pebble-cobble gravel mixed with sparse rounded boulders stand about 30 m high. Ice-marginal channels of Lignite Creek age are cut into gently sloping pediments and fans that developed after glacier retreat from the moraines of the Bear Creek advance.

Farther to the east, the early and middle Pleistocene glacial record in the Alaska Range is extremely fragmentary owing to uplift and erosional stripping within the mountains and to probable deep subsidence of glacial deposits that accumulated beyond their north flank in the Tanana River valley. An indirect record of glaciation in the Alaska Range is provided by the thick loess deposits that are exposed in mining excavations near Fairbanks (Péwé, 1989) and in river bluffs elsewhere in central Alaska (e.g., Edwards and McDowell, 1989; Begét and others, 1991). Paleomagnetic studies of the loesses (Begét and Hawkins, 1989; Begét and others, 1990) and dating of associated tephras (Stemper and others, 1989) promise to yield useful age estimates on major Pleistocene glacial and climatic fluctuations within the north-central Alaska Range.

Northwestern Alaska Range

Kline and Bundtzen (1986) describe two drift complexes that are younger than deformed Tertiary deposits but predate glacial deposits assigned to the late Pleistocene. The oldest drift,

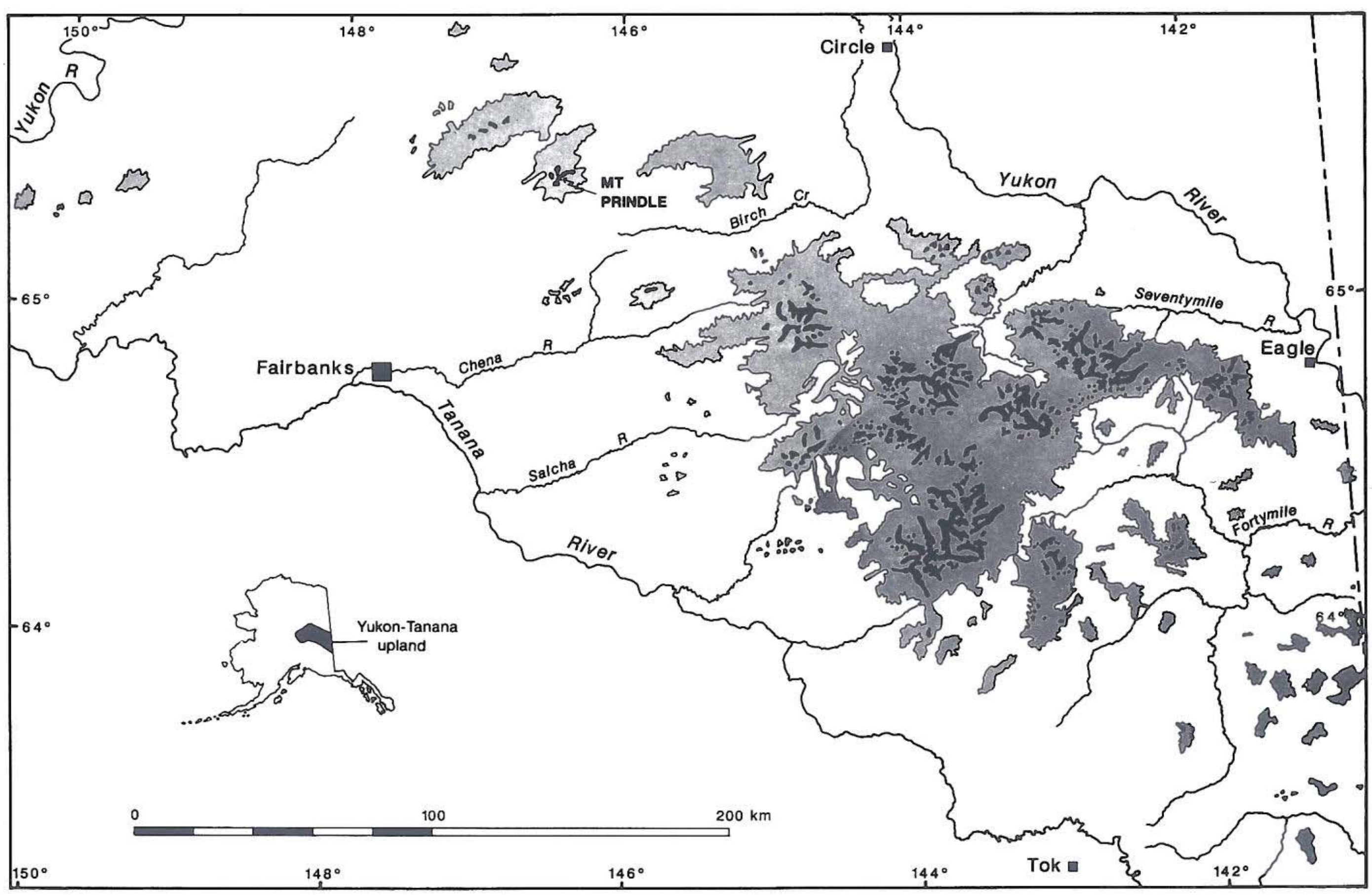

Figure 7. Approximate limits of maximum Pleistocene glaciation (light shading) and penultimate glaciation (Eagle glacial episode; dark shading) in the Yukon-Tanana Upland. Compiled by F. R. Weber, U.S. Geological Survey.

which formed during the Lone Mountain glaciation, is probably early Pleistocene in age. Younger glacial deposits, termed the Selatna drift by Fernald (1960), probably formed during the middle Pleistocene.

Glacial deposits that formed during Lone Mountain time are preserved only as erratic boulders and isolated patches of drift beyond the limits of younger drift sheets (Kline and Bundtzen, 1986). Drift remnants lack primary morainal morphology and have been smoothly graded by solifluction, but their glacial origin is indicated by erratic boulders derived from source areas deep within the Alaska Range. The deposits define several piedmont lobes that extended 40 to 50 km from the Alaska Range and terminated 5 to 10 km beyond the limtis of younger glacial advances.

Arcuate end moraines of Selatna age were deposited by piedmont lobes that extended as far as 40 km from the mountain front (Kline and Bundtzen, 1986). Subdued hummocky knob and kettle topography is common; kettles are partly or wholly drained, and most are integrated into postglacial drainage systems. Loess is more than 10 m deep locally. Vertical displacement along the Farewell fault system has offset some moraines by as much as 40 m. At least three nested moraine systems occur in some valleys, and morphologic differences suggest that they may span a long interval (Kline and Bundtzen, 1986).

Southern Alaska

The early and middle Quaternary glacial record of southern Alaska is extremely fragmentary because of late Quaternary tectonism and burial or destruction by the extensive glaciation of late Pleistocene age. Known drift deposits are described briefly here, but the attempted correlations in Table 1 are speculative.

Upper Cook Inlet. Glaciers from mountains to the north, east, and west flowed into upper Cook Inlet, where at various times they either coalesced or calved individually into tidewater (Hamilton and Thorson, 1983; Schmoll and Yehle, 1986; Reger and Updike, 1983, 1989). They generally did not leave clearly identifiable end moraines. Karlstrom (1964) defined five major Pleistocene ice advances in this region (Table 1). The oldest two advances, the Mount Susitna and Caribou Hills glaciations, are considered to be late Pliocene to early Pleistocene in overall age by Reger and Updike (1989), who believe that glaciers probably flowed down Cook Inlet to form an extensive ice shelf in the Gulf of Alaska during each of these events. Drift on Mount Susitna (see Fig. 11 for location) consists of scattered erratics on an abraded rock surface, and Reger and Updike (1983, 1989) describe similar ice-scoured upland surfaces elsewhere in upper Cook Inlet. The distribution of erratics and ice-modified surfaces indicates glacier flow patterns unrelated to modern valley systems.

The succeeding Caribou Hills glaciation was named for weathered drift in the southern Kenai lowland (see Fig. 11 for location), where glacial drainage channels and dissected lateral moraines and kame terraces still are recognizable (Reger and Updike, 1983, 1989). A possibly correlative deposit, the featureless, bouldery Mt. Magnificent ground moraine, occupies a glacially planed surface below the level of scattered mountain-top erratics near Anchorage (Schmoll and Yehle, 1986; Yehle and Schmoll, 1989). Throughout upper Cook Inlet, ice-scoured benches and truncated spurs occur at a distinct level between surfaces of Mount Susitna age and well-preserved features of younger ice advances (Reger and Updike, 1983, 1989).

Subdued moraines of the succeeding Eklutna glaciation, which occur within present drainage systems, were formed by the last Pleistocene ice masses to coalesce and entirely cover upper Cook Inlet (Reger and Updike, 1983). Ice may have overflowed westward across the Alaska Peninsula at that time (Hamilton and Thorson, 1983), forming moraines around the head of Bristol Bay that possibly correlate with Halfmoon Bay(?) drift described in the following section. Reger and Updike (1983, 1989) and Schmoll and Yehle (written communication, 1986) believe that glacier advances of Eklutna age must predate the last interglacial maximum (marine-isotope stage 5e) and therefore would be of middle Pleistocene age.

Bristol Bay region. Detterman (1986) described two drifts of possible early and middle Pleistocene age that are exposed on the Alaska Peninsula along the southeast shore of Bristol Bay. The oldest drift, which is unnamed, lacks glacial morphology and supports fully integrated drainage systems. Coastal bluffs expose moderately indurated till that unconformably underlies marine sediments and younger glacial deposits (Fig. 8). Upper surfaces of the drift and overlying marine deposits are oxidized, and both of those units are more indurated than overlying glacial sediments. The next-younger glacial unit, drift of the Johnston Hill glaciation, forms subdued morainal morphology (Detterman, 1986). Erratic lithologies suggest that glacial source areas were predominantly to the north and northeast of the Alaska Peninsula. Drift exposed in shore bluffs is weakly cemented and oxidized, and many clasts are partly disintegrated. Deposits within 40 m of present sea level have wave-cut scarps that formed when the sea stood higher against an isostatically depressed coast.

A possibly correlative glacial sequence in the Nushagak lowland along the north shore of Bristol Bay has been described by Lea and others (1989). The Nichols Hill drift unit, the oldest of the glacial deposits, was formed by a very extensive glacial advance from the Ahklun Mountains (Fig. 8). It is visible in coastal bluffs, but is nowhere exposed at the surface. The succeeding moraines, designated the Ekuk moraines by Lea and others (1989), were deposited by coalesced glacial lobes from the Ahklun Mountains (Fig. 8). Exposures in coastal bluffs show proglacial fluvial and intertidal deposits that have been strongly deformed by ice thrusting (Lea, 1990a). The Ekuk moraines may be correlative with the Halfmoon Bay(?) drift near the head of Bristol Bay, which was deposited from glaciers in the Alaska Range, and possibly also with Johnston Hill drift of the Alaska Peninsula (Table 1).

Copper River Basin. Glaciers have repeatedly flowed into the Copper River Basin from source areas in the Alaska and

Chugach Ranges to the north and south and in the Talkeetna and Wrangell Mountains to the west and east. The oldest recognized ice advances, several of which may have formed ice caps covering the entire basin (Nichols, 1989), probably date from the early and middle Pleistocene. A series of volcanic debris flows that originated in the Wrangell Mountains are interstratified with alluvium and glacial deposits in river bluffs along the Copper River and several of its tributaries (Nichols and Yehle, 1985; Richter and others, 1988). Radiometric dating of flow components or rocks from their source areas should provide a chronologic framework for much of the glacial record in the Copper River Basin.

Aleutian Islands. Extensive ice caps covered most of the Aleutian Islands during late Wisconsin time, and little evidence for earlier glaciation has been preserved (Black, 1983; Thorson and Hamilton, 1986). Glacial deposits older than late Pleistocene are recognized only on Amchitka Island, where indurated and faulted till is considered middle Pleistocene in age (Gard, 1980).

Gulf of Alaska. Exposures of the Yakataga Formation on Middleton Island document continental shelf sedimentation during latest Pliocene and early Pleistocene time—about 2.2 to 1.0 Ma

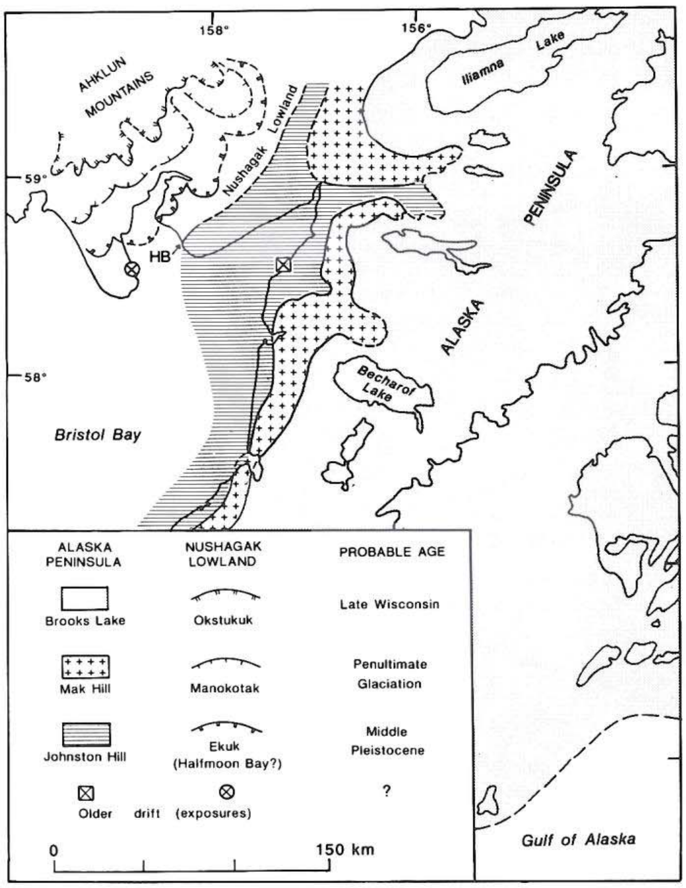

Figure 8. Glaciation of the Bristol Bay region. Drift sheets of Alaska Peninsula (shown by patterns) are from Detterman (1986) and Riehle and Detterman (1993); moraine belts of Nushagak lowland are from Lea and others (1989). Oldest drift of Nushagak lowland is the Nichols Hill drift of Lea and others (1989); oldest drift of Alaska Peninsula is unnamed. HB = outer limit of Halfmoon Bay(?) drift.

(Mankinen and Plafker, 1987). Thick marine diamicts contain faceted and striated boulders that originated in the Chugach and St. Elias Mountains; their sandy mud matrix is similar to glacially derived nearshore mud in the modern Gulf of Alaska (Plafker and Addicott, 1976; Plafker, 1981; Eyles, 1988). The diamicts contain striated boulder pavements that represent repeated expansion of a partly floating ice shelf to the edge of the continental shelf (Eyles, 1988; Eyles and Lagoe, 1989). Submarine channel-fill deposits in the basal part of the unit probably formed when glaciers were less extensive. Foraminiferal biofacies indicate that sea level fluctuated by as much as 100 m during deposition of the Yakataga Formation, and that these fluctuations probably were caused by glacial eustacy (Lagoe and others, 1989).

Abundant glacially derived sediments on the continental shelf around Middleton Island suggest that subsequent glacier advances must have reached this area, but the age of those advances is uncertain (Molnia, 1986). Marine sediment cores suggest nine ice-rafting maxima of middle Pleistocene age in the Gulf of Alaska (von Huene and others, 1976). The most intense ice rafting shown in these cores occurred about 400 ka, and ice rafting has been nearly continuous during the past 300 k.y.

LATE(?) PLEISTOCENE GLACIATIONS

Introduction

Glacial deposits of the younger ice advances in Alaska typically form rugged drift sheets, with primary depositional morphology only slightly to moderately altered. Two distinct components generally are recognized in northern and central Alaska. The younger drift is bracketed by radiocarbon ages of about 24 and 11.5 ka and therefore is equivalent in age to the late Wisconsin glaciation of midcontinental North America (Hamilton, 1982a, b) and to oxygen-isotope stage 2 of the marine sediment record. (For brevity, marine oxygen-isotope stages are termed "stage 1, stage 2," etc., through the remainder of this chapter.) Extensive ice advances of late Wisconsin age obliterated older glacial deposits in much of southern Alaska, where some areas may have remained glaciated throughout middle Wisconsin (stage 3) time.

Deposits of the next-older (penultimate), more extensive ice advance are beyond the range of radiocarbon dating, and their age assignment is controversial (Vincent, 1989; Fig. 9). Although generally inferred to be younger than the last interglacial maximum (stage 5e; e.g., Hamilton, 1986b; Heginbottom and Vincent, 1986), several recent studies have challenged that age assumption. Evidence against an early Wisconsin (stage 4) or late stage 5 age for the penultimate glaciation includes:

1. Redating of the Old Crow tephra, a widespread stratigraphic unit in Alaska and the Yukon (Westgate and others, 1983, 1985; Schweger and Matthews, 1985), from a formerly accepted age estimate of 86 ± 8 ka (Wintle and Westgate, 1986) to a present age estimate of 149 ± 13 ka (Westgate, 1988). The Old Crow tephra may have been deposited during the waning

phase of the penultimate glaciation in several parts of Alaska and the Yukon (Westgate and others, 1983; Hughes and others, 1989; Waythomas, 1990).

2. In the Yukon Territory, soils of apparent interglacial character developed on deposits of the Reid glaciation (Tarnoci, 1989), which generally is correlated with the penultimate glaciation of Alaska (e.g., Hamilton, 1986b; Heginbottom and Vincent, 1986).

3. Stratigraphic evidence suggests that the penultimate glaciation in parts of south-central and southwestern Alaska may be older than stage 5e. For example, drift of the penultimate Knik glaciation of the Upper Cook Inlet is overlain by the Goose Bay peat unit, which has an interglacial pollen flora (Ager and Brubaker, 1985) and a thermoluminescence age of about 175 ka (Reger and Updike, 1989).

4. In northern Alaska, morphologic "freshness" of drift of the penultimate glaciation, which has been used to argue for a post–stage 5e age, may be due in part to relict glacier ice that remained within these deposits and that continued to melt out during the Holocene (Hamilton, 1982b, 1986a).

5. "Early Wisconsin" deposits in several other regions of North America have been reassigned to greater ages following new geochronologic or stratigraphic studies (e.g., Curry, 1989; Easterbrook and others, 1981; Hughes and others, 1989; Pierce and others, 1976).

None of these lines of evidence, however, conclusively proves that penultimate glaciation through all of Alaska is of pre–stage 5e age, and equally strong arguments can be made for a younger age assignment. For example:

1. Deposits of the penultimate and late Wisconsin ice advances in the Brooks Range are components of a single drift sheet (Porter, 1964) whose outer limit is a more striking discontinuity than any within the drift complex (Hamilton, 1986a).

2. In areas of active tectonism, drift of the penultimate glaciation has been disrupted to a lesser degree than would be expected if it preceded stage 5e (i.e., if it formed prior to about 130 ka). For example, the Denali fault has undergone 50 to 60 m of right-lateral movement during the last 10 k.y. where it crosses the Delta River valley (Stout and others, 1973), yet during the interval that elapsed between the penultimate and late Wisconsin glaciations that glacial trough was not dislocated sufficiently to alter glacier flow patterns through it.

3. Little stratigraphic evidence is available for presence of interglacial deposits above drift of the penultimate glaciation. For example, the penultimate glaciation of the Nushagak lowland was followed by long intervals of thaw-lake and organic deposi-

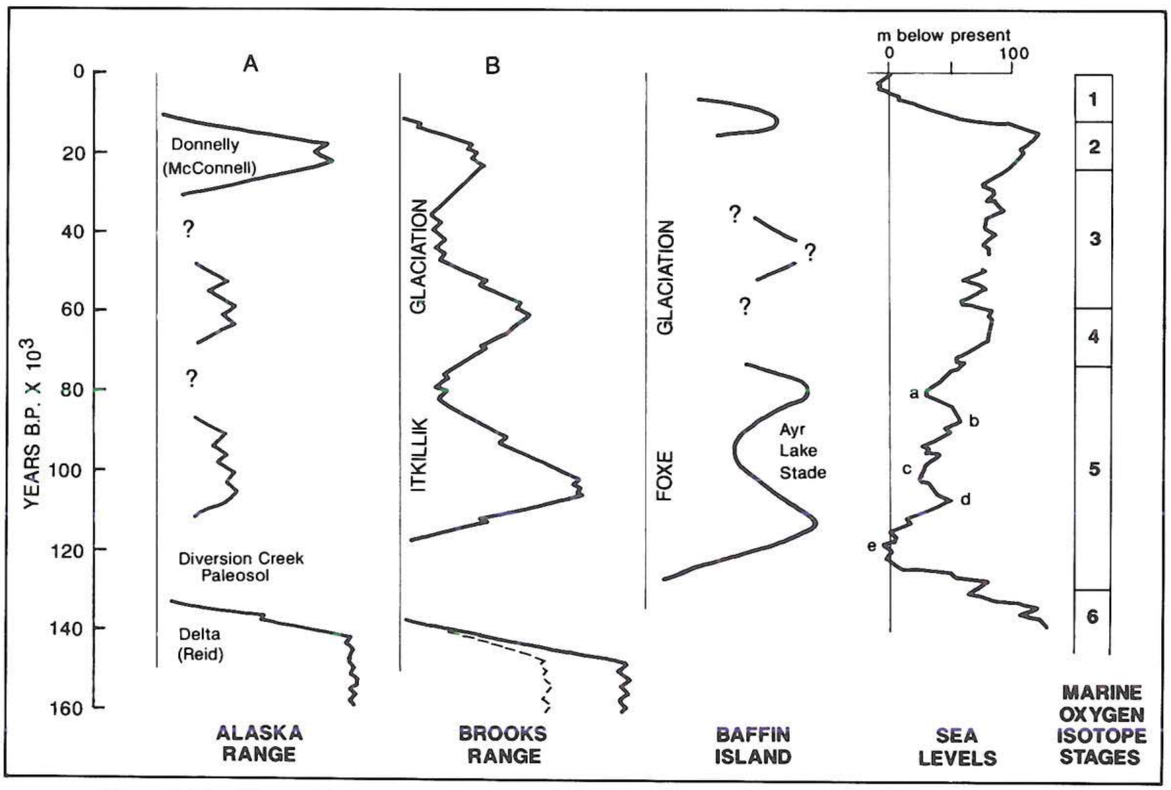

Figure 9. Possible models for youngest major Pleistocene glacier advances in Alaska. A, Alaska Range model. Penultimate glaciation preceded stage 5e; subsequent ice advances are obliterated by extensive late Wisconsin (stage 2) advance. B, Brooks Range model. Penultimate glaciation followed stage 5e and either partly overlapped or obliterated drift of late stage 6 (shown by solid and dashed lines, respectively). Baffin Island glacial record (from Andrews and Miller, 1984) and worldwide sea levels estimated from marine oxygen isotope record (from Shackleton, 1987) are shown for comparison.

tion under climates that were cooler than those of the Holocene (Lea and others, 1989).

4. On the arctic coast of Alaska, the Flaxman Member of the Gubik Formation, which records the breakup of an extensive ice shelf in the Canadian Arctic (see following section), is older than 52 ka but is younger than marine deposits believed to have formed during stage 5e (Carter and Ager, 1989; L. D. Carter, written communication, 1990).

5. Extensive glaciation in high latitudes late in stage 5 is demonstrated on Baffin Island, in the eastern Canadian Arctic, where stratified coastal exposures and marine sediment cores have been dated by radiocarbon and uranium-series methods and correlated by amino-acid ratios and pollen data (Andrews and Miller, 1984; Andrews and others, 1985). The most extensive glacier advance of the late Pleistocene on Baffin Island began shortly after stage 5e and terminated about 70 ka, and sediment cores from Baffin Bay show dominance of ice-rafted sediments from late stage 5 into stage 4. Uranium-series ages from Banks and Victoria Islands, in the western Canadian Arctic, suggest a similar glacial chronology (Causse and Vincent, 1989). Subsequent ice advances of late Wisconsin age are much more restricted in comparison.

Two models clearly are possible for the penultimate glaciation of Alaska (Fig. 9), and an additional possibility is that northern Alaska may follow a high-latitude model like that of Baffin Island, whereas southern Alaska may respond more like glaciers in more temperate regions of North America. According to Andrews and others (1985), circulation of warm water into Baffin Bay generated increased snowfall on Baffin Island during stages 5 and 4, causing glaciers to advance. Glaciers subsequently were so inhibited by absence of precipitation that maximum late Wisconsin advances did not occur until 12 to 8 ka, when retreat of the Laurentide ice sheet farther south allowed greater moisture to penetrate into the eastern Canadian Arctic (Andrews, 1987). If glaciers in northern Alaska were similarly controlled by moisture availability (Hamilton, 1981), in contrast to greater influence of temperature fluctuations farther south, there may be no single age for the "penultimate glaciation" of Alaska.

Arctic Coastal Plain

The Flaxman Member of the Gubik Formation consists of erratic-bearing glaciomarine sediments that occur locally along the Beaufort Sea coast to altitudes of about 7 m (Carter and others, 1986a, 1988; see Fig. 1). The erratic stones, of Canadian provenance, were transported to the Beaufort Sea coast by icebergs from an extensive ice sheet in the Canadian Arctic. Marine mollusk shells are enriched in ^{18}O relative to modern shells from the Beaufort Sea, confirming that more glacial ice was present in Flaxman time than occurs today. The Flaxman Member has been assumed to represent the distal glaciomarine facies of the Buckland advance of the northwestern sector of the Laurentide Ice Sheet (Vincent and Prest, 1987; Vincent and others, 1989), but the presence of spruce macrofossils suggests that it was deposited during an interstadial (Carter and Ager, 1989), and distinct erratic lithologies indicate a glacial source farther to the east (L. D. Carter, personal communication, 1990). Carter and Ager (1989) report that 11 thermoluminescence (TL) ages on sediments of the Flaxman Member range from 53 to 81 ka, and a uranium-series age on associated whale bone is 75 ka, but more recent unpublished TL determinations show wider age range and indicate that the TL method may be unreliable for the muddy glaciomarine sediments of the Alaskan arctic coast (L. D. Carter, personal communication, 1990).

A glaciomarine transgressive unit that locally exceeds 40 m thickness occurs on the Alaskan Beaufort shelf at water depths shallower than about 100 to 115 m (Dinter, 1985). It is composed of gravelly silt and sand derived from the Canadian Arctic and overlies a disconformity that probably represents the minimum sea-level stand of late Wisconsin time (Dinter, 1985). The transgressive unit evidently was deposited by sediment-laden icebergs that calved from the Laurentide Ice Sheet. Dinter (1985) suggested that the transgressive unit is late Wisconsin and Holocene in age, and that its outer limit marks the late Wisconsin minimum sea-level position. However, new core data and reinterpretation of seismic profiles indicate that the unit is probably older than the late Wisconsin (Dinter and others, 1990).

The central Brooks Range

The youngest major drift complex in the central Brooks Range is assigned to the Itkillik glaciation (Detterman and others, 1958), which is divisible into two major phases—Itkillik I and Itkillik II (Hamilton and Porter, 1975; Hamilton, 1986a). The outer border of Itkillik drift is the most striking discontinuity in the entire Brooks Range glacial succession; it separates subdued older glacial deposits from steeper, stonier, and more irregular drift with drainage anomalies and discontinuous vegetation.

During Itkillik I time, glaciers extended as far as 40 km north of the Brooks Range (Fig. 10), forming moraines having slightly flattened crests and irregular profiles (Table 2). Drainage nets are well integrated, but stream courses are largely controlled by original drift morphology. Streams cross moraines in bouldery riffles, and outwash terraces are 20 to 25 m high near moraine fronts. Nested sets of recessional moraines and ice-marginal drainage channels indicate that glaciers retreated dynamically from northern valleys rather than stagnating in place. Moraines in forested southern valleys generally have smooth, somewhat flattened, hummocky crests that support sparse vegetation; moist depressions are covered by dense muskeg.

The maximum advance of Itkillik I time (phase A) may have been followed by a younger major readvance (phase B) (Hamilton, 1986a, 1989); both advances are older than the effective range of radiocarbon dating.

Drift-mantled glacier ice persisted for long intervals in many northern valleys, and relict ice of Itkillik I age may still be widely present today (Hamilton, 1986a). Kettles that have turbid water, steep flanks of unstable gravel, and highly angular outlines remain active today in several valleys.

The largest glaciers of Itkillik II age extended about 25 km north of the Brooks Range (Fig. 10). Drift limits generally are marked by sharp lateral moraines, heads of conspicuously channeled outwash aprons, abrupt narrowing of stream incisions, and bulky end moraines and ice-stagnation deposits that nearly block valley centers. The deposits are slightly more irregular, steeper sided, stonier, and less vegetated than drift of Itkillik I age (Table 2), and surface drainage is poorly integrated. End moraines retain multiple crests and other minor relief features. Kettles are deep and steep sided; many have bare, unstable flanks, open tension cracks, turbid water, and active slumps indicative of continued enlargement. Solifluction occurs locally on lower slopes, but crests and upper slopes lack colluvium. Axial streams cross end moraines through steep, bouldery rapids, and outwash terraces are 10 to 20 m high. Recessional moraines are rare, indicating general downwastage rather than frontal retreat of glaciers. Rugged end moraines in southern valleys (Table 2) commonly are located at the margin of the Brooks Range and enclose elongate lakes or lacustrine plains. Drift of Itkillik II age formed between about 24 and 11.5 ka and correlates with the late Wisconsin fluctuations of the Laurentide Ice Sheet (Hamilton, 1982a, 1986a, b; Hamilton and others, 1987).

A major readvance of glaciers near the close of Itkillik II time formed distinctive end moraines as much as 15 to 20 km beyond the range front in the larger northern valleys (Hamilton, 1986a). Elongate lakes 20 to 60 km long formed in most northern valleys as glaciers retreated. Their upper limits rise southward by about 100 m, probably due to isostatic recovery following deglaciation. Radiocarbon ages demonstrate that the late Itkillik II readvance took place between 13 and 11.5 ka (Hamilton, 1982a, 1986a).

The Seward Peninsula

The Salmon Lake and Mount Osborn glaciations of the Seward Peninsula have relatively fresh morainal morphology (Table 2) and resemble Itkillik-age deposits of the Brooks Range (Kaufman and Hopkins, 1986). Salmon Lake drift, which is correlated with deposits of the Itkillik I phase, is older than 40 ka; the succeeding Mount Osborn advance probably is of late Wisconsin (Itkillik II) age.

Glaciers of Salmon Lake age averaged about 16 km long in the Kigluaik Mountains, where they filled trunk valleys and formed large lobate moraines beyond valley mouths (Kaufman

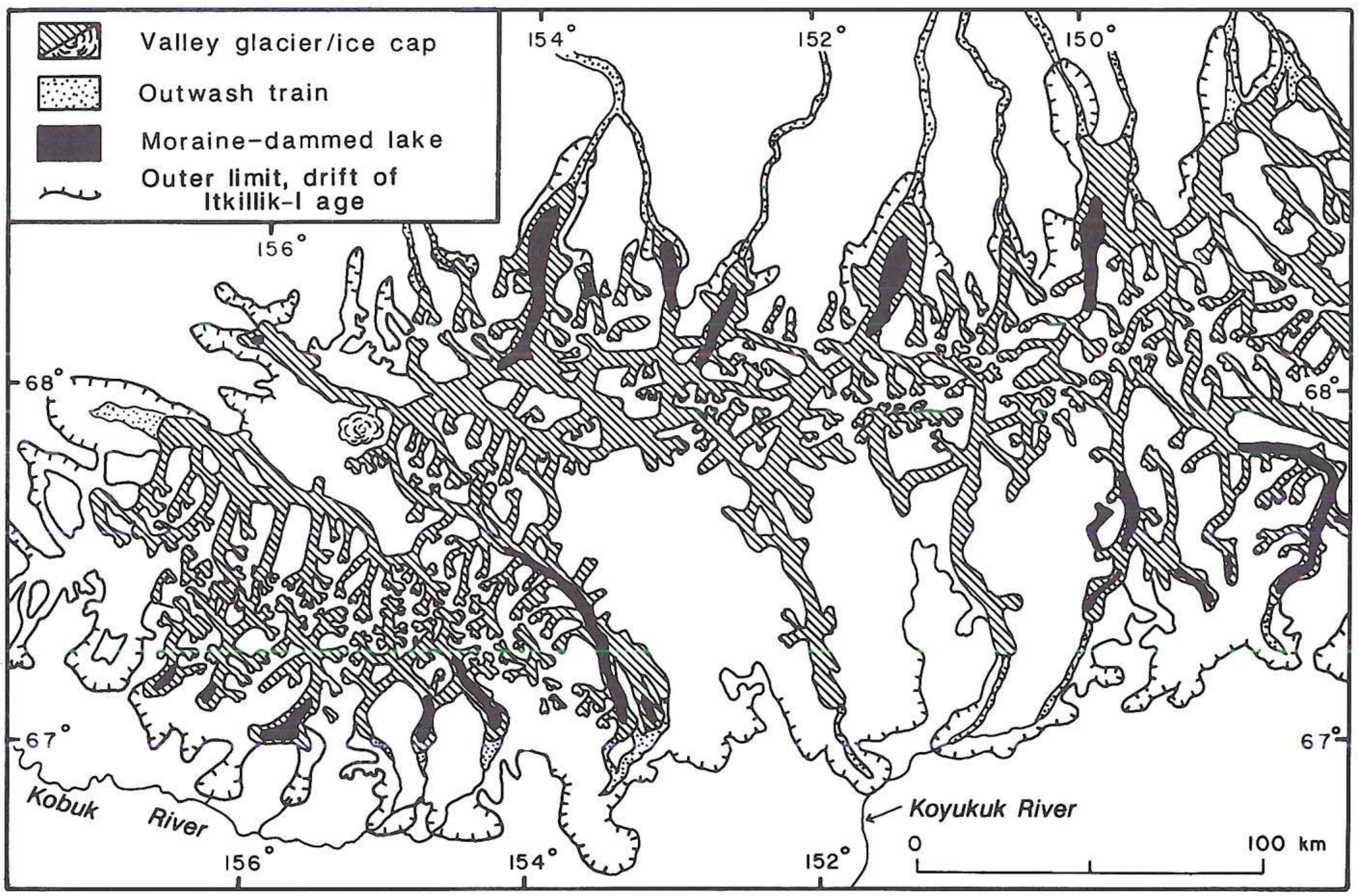

Figure 10. Distribution of glaciers, ice caps, and moraine-dammed lakes of Itkillik II age, central Brooks Range. Hachures show outer limits of preceding Itkillik I ice advance. From Hamilton (1986a).

and Hopkins, 1986; Kaufman and others, 1988); shorter glaciers in the Bendeleben and Darby Mountains were limited to tributary valleys and heads of trunk valleys (Fig. 4). Drift deposits have well-preserved glacial morphology, but have been significantly affected by periglacial processes (Kaufman and Calkin, 1988). In the York Mountains, drift of the Mint River glaciation (Sainsbury, 1967) is similar in appearance to drift of Salmon Lake age, and reconstructed equilibrium-line altitudes (ELAs) of both drifts are compatible (Kaufman and Hopkins, 1986). Morphologic criteria suggest that the Salmon Lake glaciation could be of either stage 4 or stage 6 age (Kaufman and Calkin, 1988).

During Mount Osborn time, small glaciers occupied high cirques and upper valleys in the Kigluaik Mountains, where they extended 2.5 to 3.0 km from cirque headwalls (Fig. 4). Moraines are sharply defined, with bouldery surfaces, and extend upvalley into prominent trimlines along valley walls (Kaufman and Calkin, 1988). Projection of reconstructed ELAs for Mount Osborn glaciers into the York Mountains indicates that snowline was at the appropriate altitude to generate the Esch Creek glaciation (Sainsbury, 1967), which therefore is considered to be correlative with the Mount Osborn advance (Kaufman and Hopkins, 1986). The highest north-facing cirques in the Kigluaik Mountains contain moraines that record a glacial stillstand or readvance late in Mount Osborn time, but a radiocarbon age from one cirque suggests that subsequent deglaciation was complete by 11.5 ka (Kaufman and others, 1988, 1989a).

The Yukon-Tanana Upland

Glacial deposits of Eagle and Salcha age in the Yukon-Tanana Upland were assigned to the late Pleistocene by Weber (1986). She correlated the Eagle glacial episode with ice advances of isotope stage 4 or late stage 5 age elsewhere in Alaska on the basis of tephrochronology, morphology (Table 2), and similar positions in glacial successions, and correlated the Salcha glacial episode with radiocarbon-dated glacial sequences of late Wisconsin age in the Alaska Range and the Brooks Range. Age-equivalent glacial advances in the Mount Prindle area were assigned by Weber and Hamilton (1984) to the American Creek and Convert glaciations, respectively.

Glaciers of Eagle age formed at and above altitudes of 1,460 m; they extended as much as 19 km through upper mountain valleys and generally terminated at about 900 m altitude (Fig. 7). End moraines generally occur above modern timberline; outwash trains form flat valley floors that are little dissected by modern streams. Almost all drift sheets contain multiple end moraines that indicate two distinct glacial advances.

Glaciers of Salcha age originated on the highest uplands, where they reoccupied the cirques and radiating valleys that had been shaped during the preceding glacial episode (Weber, 1986). Most glaciers were confined to upper valleys and terminated above modern timberline at altitudes above 1,200 m; a few of the longest ice tongues extended about 300 m lower. Moraine sequences in most valleys indicate four glacial advances. Moraines of the youngest two phases are weakly developed, and coalesce in some valleys.

The Kuskokwim Mountains

Drift deposits of Bifurcation Creek and Tolstoi Lake glaciations in the Kuskokwim Mountains have been described by Kline and Bundtzen (1986), who assigned them to the early Wisconsin and late Wisconsin, respectively.

Major cirque basins and U-shaped valleys in the Beaver Mountains were formed during ice advances of Bifurcation Creek age (Kline and Bundtzen, 1986). Glaciers extended about 10 km down valleys that radiated generally northward from the center of the massif and extended 4.5 to 6 km down south-trending valleys. Two and rarely three end moraines are present in almost all valleys. Cirque headwalls and valley flanks have been modified by stream incisions and by accumulation of colluvial rubble.

During Tolstoi Lake time, glaciers advanced as far as 6.5 km down northern valleys and 1.5 km or less down valleys that trend east or southeast (Kline and Bundtzen, 1986). South-facing cirques and valleys apparently did not contain ice at that time. Steep-fronted drift sheets were deposited in most major northern valleys, and multiple end moraines record three (rarely 4) distinct ice advances. Most cirque headwalls are unmodified except for talus cones, and cirque floors are undissected. A late Wisconsin age is inferred by Bundtzen (1980) for the drift on the basis of its topographic freshness, slight weathering of boulders, weak soil development, and unmodified cirques.

North-central Alaska Range

The youngest two major Pleistocene glacial advances formed conspicuous deposits in most valley systems of the north-central Alaska Range (Porter and others, 1983). The older of the two drifts is assigned to the Delta glaciation in the Delta River region (Péwé and Reger, 1989), whereas farther west in the Nenana River region, it is referred to the Healy glaciation (Wahrhaftig, 1958; Thorson, 1986). The Delta glaciation previously was believed to be of early Wisconsin age because Delta-age outwash at the Canyon Creek fossil locality underlies vertebrate fossils that had a uranium-series age estimate of about 80 ka (Weber and others, 1981; Hamilton and Bischoff, 1984; Weber, 1986). However, uranium-series ages on bones are now considered suspect (Bischoff and others, 1988; J. L. Bischoff, personal communication, 1990), and the age of the Delta glaciation is not longer constrained by the stratigraphy at Canyon Creek. The Healy terminal moraine of the Nenana River valley is a spatulate body of coarse, gravelly drift that consists of as many as ten individual ridges separated by undissected marginal channels (Thorson, 1986). Its sharply defined outer margin forms the limit of deposits with stony surfaces, original morainal morphology, and deranged drainage patterns. Drift of Healy age is older than the limit of radiocarbon dating; it was believed to postdate the last interglacial maximum (Ritter, 1982; Ten Brink, 1983;

Thorson, 1986) because it has not been severely weathered and because it is closer morphologically to drift of late Wisconsin age than it is to any older glacial deposits. Two main phases of Delta-age glaciation were suggested by Weber and others (1981) on the basis of drift morphology in the Delta River valley and stratigraphy of the Canyon Creek fossil locality, and two main phases of "early Wisconsin" glaciation also were reported by Ten Brink (1983) based on field studies in the Nenana valley region.

Expansion of succeeding glaciers of Donnelly or Riley Creek age to the north flank of the Alaska Range is closely constrained by two radiocarbon ages of 25.3 and 24.9 ka, and similar ages of about 26.4 and 23.9 ka provide maximum limits on the late Wisconsin McConnell glaciation in southwestern Yukon Territory (Hamilton and Fulton, 1994). Late Wisconsin glaciation in the north-central Alaska Range reportedly consists of four phases (Ten Brink and Waythomas, 1985), and outwash terraces formed during the oldest three phases are present in most valleys (Ritter and Ten Brink, 1986). During phase I, the maximum advance, glaciers constructed steep-fronted, bouldery, nested moraine systems in every major valley. They attained maximum positions at about 20 ka and began to retreat at about 17 ka. During stillstands and short readvances of phase II, dated at 15.0 to 13.5 ka, glaciers built a separate set of end moraines 2 to 10 km upvalley from the outermost moraines (Ten Brink and Waythomas, 1985). Phase II moraines generally are smaller than end moraines of phase I, and they contain more water-washed sediment. Phase II was followed by an episode of rapid glacier retreat and ice disintegration. Drift of phase III, a brief but strong glacier readvance at about 12.8 to 11.8 ka, includes abundant ice-disintegration features (Ten Brink and Waythomas, 1985). Moraines commonly cross-cut older moraines of the late Wisconsin succession, implying significant glacier retreat and readavance. This readvance is similar in age to the late Itkillik II readvance described previously in the central Brooks Range (Hamilton and Fulton, 1991). During Phase IV, the final late Wisconsin readvance, glaciers formed small moraines a few km beyond the limits of late Holocene ice advances in headward parts of the highest valleys. Ten Brink and Waythomas (1985) date phase IV at 10.5 to 9.5 ka, but this age assignment is based only on correlation with intervals of peat accumulation, which they believe represents wet and/or cool climate favorable for glacier growth.

Northwestern Alaska Range

The last major series of ice advances—the Farewell glaciation—in the western Alaska Range was subdivided by Fernald (1960) into two phases: Farewell 1 and Farewell 2. A radiocarbon age of 34,340 ± 1,940 B.P. from an organic layer near the base of overlying loess provides a minimum age limit on drift of the Farewell 1 advance, and Kline and Bundtzen (1986) believe that the Farewell 1 and Farewell 2 drifts probably are of early and late Wisconsin age, respectively.

During Farewell 1 time, glaciers formed piedmont lobes as much as 32 km beyond the Alaska Range (Kline and Bundtzen, 1986). Frozen peat and ice-rich silt as much as 10 m thick covers parts of the drift, and many ponds and lakes on its surface are thermokarst features. Multiple moraine systems in many valleys indicate two phases of glaciation.

Glaciers of Farewell 2 age advanced as much as 23 km from the range front where valleys drained ice fields south of the Alaska Range, but elsewhere they terminated close to the mouths of mountain valleys (Kline and Bundtzen, 1986). At least four glacial advances took place in most major valleys: the two earliest and most extensive formed arcuate moraines, the third was followed by widespread ice stagnation, and the fourth generally produced small moraines within mountain valleys. Correlations between these advances and the four glacial phases in the north-central Alaska Range are uncertain.

The St. Elias Mountains

Two drifts, assigned to pre-Macauley and Macauley glaciations, occur in stratigraphic sections and surface exposures along the White River valley. These units are correlative, respectively, with the Icefield and Kluane drift units in the northern St. Elias Mountains (Table 1). Both glacial events were assigned by Denton (1974) to the Wisconsin stage on the basis of radiocarbon ages, stratigraphic relations, and lack of significant weathering. However, drift of pre-Macauley age (Icefield glaciation) generally is correlated with drift of the Mirror Creek glaciation of southwest Yukon Territory, which is overlain by Old Crow tephra (Hughes and others, 1989). Therefore, it could be of pre-Wisconsin age. Deglaciation was followed by a nonglacial interval that extended from sometime prior to 49 ka until about 30 ka (Denton, 1974).

Macauley (Kluane) drift forms hummocky deposits having abundant kettles and sharp morphologic boundaries (Denton, 1974). Radiocarbon ages show that glaciers began to advance shortly after 29 ka and attained maxima sometime before about 13.7 ka. Deglaciation of major valleys was nearly complete by 12.5 ka, and valley heads within a few kilometers of modern large glaciers were ice free by 11.3 ka (Denton, 1974).

Upper Cook Inlet

The two youngest major glacial advances of Karlstrom (1964), the Knik and the Naptowne glaciations, entered upper Cook Inlet but probably did not entirely fill it (Fig. 11). During Knik time, ice lobes coalesced over the Anchorage area, but part of the Kenai lowland probably remained unglaciated. Moraines are slightly subdued and commonly are capped by as much as 1 m of loess (Reger and Updike, 1983). Subsurface diamict of probable Knik age in the Anchorage area underlies drift of early Naptowne age (Reger and Updike, 1989). Because it underlies peat with a probable interglacial flora (Ager and Brubaker, 1985), drift of the Knik glaciation appears to be older than stage 5e.

Deposits laid down during the Naptowne ice advance cover most lowlands of upper Cook Inlet (Reger and Updike, 1983,

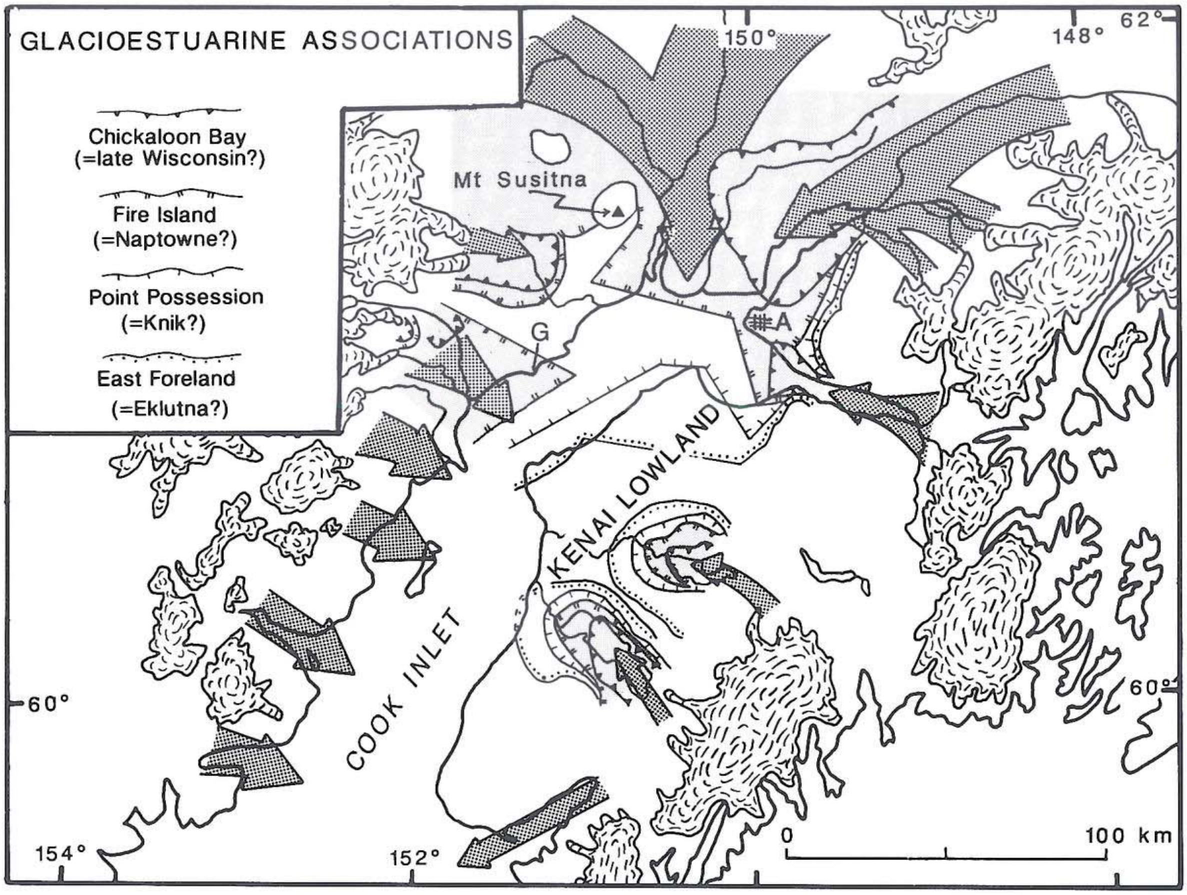

Figure 11. Late Pleistocene and present-day glaciation of Cook Inlet. Modern glaciers shown by concentric dashed patterns. Arrows show flow routes of principal glacier tongues; shaded areas are inferred extents of late Wisconsin glaciers in upper inlet. Principal glacioestuarine associations are shown for upper inlet only (Cairn Point GEA is not shown). A = Anchorage; G = Granite Point. Modified from Schmoll and Yehle (1986).

1989), Drift typically is little modified, with numerous kettles and poorly integrated stream systems; it has about 0.3 m of loess cover. Separate advances of early and late Wisconsin age may have taken place (Reger and Updike, 1989). Radiocarbon ages on early Naptowne drift are contradictory (Reger and Updike, 1983, p. 201), but they imply that the initial advance is older than the effective range of radiocarbon dating. No middle Wisconsin radiocarbon ages have been reported from upper Cook Inlet. Plant growth at that time evidently was severely inhibited by some combination of expanded glaciers and marine submergence (Schmoll and Yehle, 1986). Till and ice-stagnation deposits of early Naptowne age in the Anchorage area are overlain by the glaciomarine Bootlegger Cove Formation (Updike and others, 1984, p. 14–17). Deposition of the Bootlegger Cove Formation probably began sometime before about 18 ka, and radiocarbon ages of 12.2 to 11.5 ka on overlying peat provide a minimum age limit for this unit (Reger and Updike, 1983, 1989).

Most moraines in upper Cook Inlet become indistinct below about 120 m above present sea level, where glaciers probably reached tidewater in the isostatically depressed basin (Schmoll and Yehle, 1986). Schmoll and others (1984) have defined five glacioestuarine associations (GEAs) in upper Cook Inlet (Fig. 11). Each GEA is an assemblage of landforms and deposits that represents the inferred marginal positions of glaciers during a specific interval of time (Schmoll and Yehle, 1986). Some GEAs may have formed during major ice advances, but others probably represent only glacier stillstands or minor readvances. The East Foreland GEA, oldest of the succession, represents a glacier complex that filled all of upper Cook Inlet except for a small part of the Kenai Peninsula. Ice-margin positions are somewhat similar to those of the Eklutna glaciation, and they lie beyond the Knik ice limits mapped by Reger and Updike (1983). The glacier limits of the succeeding Point Possession GEA are close to the outermost Knik moraines. The Fire Island GEA may correspond to the early Naptowne ice limits mapped by Reger and Updike (1983) because reconstructions of both limits show that glacier ice last covered the Anchorage area at that time. The Chickaloon Bay GEA, which most closely coincides with deposition of the

Bootlegger Cove Formation at Anchorage, has a minimum limiting age of 13.8 ka. The Elmendorf moraine, which forms part of the succeeding Cairn Point GEA, was deposited between 13.7 and 11.7 ka (Schmoll and others, 1972).

Submarine diamicts in Shelikof Strait fill deep U-shaped troughs that Hampton (1985) believes formed during Naptowne time. The trough more likely formed during an earlier glacier advance out of upper Cook Inlet, and the thick (>800 m) diamicts may contain a long record of middle and late Pleistocene glaciomarine deposition.

Bristol Bay region

The youngest Pleistocene drift complexes of the Alaska Peninsula are assigned to the Mak Hill and Brooks Lake glaciations. Drift of Mak Hill age formed sometime prior to 40 ka (Detterman, 1986); the succeeding four-fold Brooks Lake drift sequence is assigned to the late Wisconsin.

Drift of Mak Hill age is widely exposed along the northwest side of the Alaska Peninsula and forms much of the Bristol Bay lowland (Fig. 8). Erratics are mainly rock types that originated from volcanoes of the Alaska Peninsula (Detterman, 1986). End moraines are distinct but somewhat subdued, and some segments have been modified by wave action. Drift of Mak Hill age is locally covered by up to 3 m of silt that contains tephra layers. Marine terraces 15 to 18 m above present sea level have been cut into the glacial deposits, and associated marine sediments also overlap Mak Hill drift (Detterman and others, 1987).

During Brooks Lake time, glaciers originating on volcanoes of the Alaska Peninsula coalesced with ice caps that formed south of the Alaska Peninsula (Weber, 1985) and expanded into a nearly continuous glacier complex. Four major glacial advances are termed the Kvichak, Iliamna, Newhalen, and Iliuk advances (Detterman, 1986). Drift of the Kvichak advance locally was deposited in water or was modified by subsequent wave action. The succeeding Iliamna advance was the last major glacial event on the Alaska Peninsula, and its end moraines enclose most of that region's large lakes. Drift morphology resembles that of the Kvichak advance, but moraine crests are a little sharper and flanks are somewhat steeper. The two youngest glacial advances, Newhalen and Iliuk, were minor events largely restricted to mountain valleys. Radiocarbon dates indicate that glaciers retreated from Newhalen moraines before 9 ka and probably prior to 10.6 ka (Detterman, 1986).

On the northwest shore of Bristol Bay, the youngest two Pleistocene glaciations formed distinctive moraines in the Nushagak lowland (Fig. 8). The penultimate glaciation may be represented by discontinuous segments of the Manokotak and Gnarled Mountain moraine belts, which exhibit subdued hummocky microrelief (Lea and others, 1989). The two moraine belts may correlate with the subdued end moraines of Mak Hill age on the Alaska Peninsula (Table 1).

The Okstukuk and Alegnagik moraine belts, which formed during the last glaciation, are within 10 km of range-front lakes and have little-modified hummocky topography (Lea and others, 1989). According to Lea and others (1989), these moraines probably are equivalent to the Kvichak and Iliamna moraines of the Brooks Lake glaciation on the Alaska Peninsula. Wind-deflated sediments from glacial outwash were redeposited as sand sheets and sand-loess intergrades; these deposits contain beds of slightly organic silt that probably formed during interstadial episodes of the last glaciation (Lea, 1989, 1990b). A distinctive silt bed near the base of the eolian sand has radiocarbon ages of 21.6 to about 18 ka (Lea, 1989). The silt bed represents a possible interstade that may correspond to the Hanging Lake thermal event of eastern Beringia, which is dated at about 22 to 18 ka by Matthews and others (1989) and 22 to 19 ka by Hamilton and Fulton (1994). A higher organic horizon marks another possible interstadial at about 14.3 ka. Eolian deposition ceased by about 12.5 ka, probably when glaciers receded into mountain valleys and their sediments became trapped by the range-front lakes (Lea, 1989).

Similar relations of eolian sand to glacial intervals 200 km farther north in the Holitna lowland are reported by Short and others (1990) and by Waythomas (1990). Eolian sand and silt that may correlate with the penultimate glaciation contains the Old Crow tephra near its upper contact, indicating that the tephra was deposited during waning stages of that glaciation. If the stratigraphic placement of the Old Crow tephra is valid, and if the 149 ka age estimate for this unit is correct, a stage 6 age would be indicated for the penultimate glaciation north of Bristol Bay.

The Copper River Basin

During late Pleistocene time, glaciers in the Chugach Range blocked the Copper River and caused a large lake to form in the basin enclosed between the Alaska Range and the Talkeetna, Chugach, and Wrangell Mountains (see Figs. 1 and 2). Large glaciers entered the basin from the Chugach Mountains; smaller glaciers from the other ranges intermittently blocked northern and western spillways at higher altitudes. Lava flows, volcanic debris flows, and tephra also entered the basin from active volcanoes of the Wrangell Mountains (Connor, 1984; Nichols and Yehle, 1985).

River bluffs contain multiple diamicts together with lacustrine, fluvial, and eolian deposits and volcanic detritus (Ferrians and others, 1983). Many of the diamicts are deposits of debris flows, turbidites, and sediment gravity flows that originated by slumping around the margins of the basin, perhaps triggered by earthquake shocks (Eyles, 1987; Ferrians, 1989). Glaciers probably covered much of the basin floor during the penultimate glacial advance (Hamilton and Thorson, 1983; Nichols, 1989); during later episodes, they calved into a deep lake that filled the central part of the basin (Williams and Galloway, 1986).

The highest recorded lake levels in the northeastern part of the basin are at 915 m altitude (Schmoll, 1984), and Williams and Galloway (1986) show lake deposits at altitudes as great as 975 m farther to the west. Discrepancies in shoreline altitudes may be due in part to isostatic tilting, but they also could result from different ages or differential preservation of the deposits.

The youngest major lake (glacial Lake Atna) began to fil the basin sometime before 40 ka (Williams and Galloway, 1986; Nichols, 1989) and perhaps before 50 ka (Ferrians, 1989); associated lacustrine deposits occur to altitudes as high as 775 m in the western part of the basin (Williams, 1989). An interstadial episode of lowered lake levels is dated at about 31 to 28 ka in the eastern sector of the basin (Ager, 1989; Ferrians, 1989; Nichols, 1989) and persisted from at least 31 ka until sometime after 22 ka in the northwestern part of the basin (Thorson and others, 1981). During this episode, lake level dropped below 655 m altitude across much of the basin (Ferrians, 1989).

The highest shorelines of the youngest lake stage occur at about 947 m asl throughout the Copper River Basin (Williams and Galloway, 1986), and indicate little or no local isostatic warping at this time. Moraines of Wisconsin age in southern and western parts of the basin are mantled with lacustrine silt to altitudes as high as 945 m, indicating that the lake persisted until glaciers had withdrawn into the Chugach Range (Williams and Galloway, 1986). Deglaciation was well advanced by 14 to 13 ka (Williams, 1986; Williams and Galloway, 1986; Sirkin and Tuthill 1987), and most of the basin became ice-free by 11.8 to 10.5 ka (Hamilton and Thorson, 1983).

The Aleutian Islands

Glaciers of late Wisconsin age covered nearly all of the inner Aleutian Islands and extended over much of the adjoining Aleutian platform (Black, 1983; Thorson and Hamilton, 1986; see Fig. 1). The axis of ice dispersal lay south of the inner islands, indicating that the Pacific Ocean was the primary moisture source for glaciers through this part of the Aleutian chain. Glaciers flowed north across the Aleutian platform and probably terminated as a floating ice shelf in the southern Bering Sea. Ice caps on Umnak and Unalaska Islands were confluent with the glacier complex on the Alaska Peninsula, but the western limit of completely coalesced ice is uncertain. Minimum limiting ages from postglacial sediments commonly are 8 to 6 ka (Black, 1983), but Thorson and Hamilton (1986) suggest that eustatically rising sea level, together with warming air and ocean temperatures, probably caused disintegration of the glacier complex sometime before 11 ka.

Gulf of Alaska

The Gulf of Alaska continental shelf consists of Tertiary and Pleistocene sedimentary rocks into which glacial troughs have been incised (Molnia, 1986). Overlying sediments include till, outwash, and glaciomarine deposits that thicken to more than 50 m in major troughs. The extent of late Pleistocene ice cover on the continental shelf is uncertain, but glaciers that reached the edge of the shelf may have been confined to the major troughs (see Fig. 1).

Carlson and others (1982) delineated eight major troughs and valleys that are incised into the continental shelf of the northeastern Gulf of Alaska (see Fig. 6). These features generally have U-shaped cross sections, concave longitudinal profiles, and sediment fillings whose seismic signatures are identical to those of known glacial deposits (Molnia, 1986). Their positions correspond to major onshore glacier systems or probable glacial flow routes during late Pleistocene time, and they probably were formed by major ice streams that flowed to the shelf edge (Carlson, 1989; Powell and Molnia, 1989).

The last major episode of ice retreat in the Gulf of Alaska region probably began at 12 to 15 ka (Heusser, 1985; Molnia, 1986). Radiocarbon ages from coastal areas suggest that glacier ice had retreated from the continental shelf by at least 10 ka and that glaciers draining the coastal mountains had reached positions comparable to those of the present day by about 9.6 to 9.3 ka.

Lituya Bay and Glacier Bay

Mann (1986) describes three informally named late Pleistocene drifts near Lituya Bay—the High terrace, Raven House, and Eurhythmic drift units. The High terrace drift unit, oldest of the succession, is preserved only on terraces more than 500 m above sea level (asl). Surface deposits are rare, and bogs blanket all glacial deposits. Weakly oxidized till and outwash in one exposure overlie peat that has a radiocarbon age of >72 ka and a pollen content that indicates interglacial vegetation. Because deep stream incisions are graded to a lower terrace, which probably formed between 50 and 30 ka, Mann (1986) assigned the High terrace drift unit to the early Wisconsin.

The Raven House drift unit includes subparallel moraines armored with wave-washed boulders that are distributed across a marine terrace at 180 to 210 m asl. Maximum limiting ages for Raven House drift were obtained from estimates of the age of formation of its associated terrace (Mann, 1986). Calculated rates of uplift and tilting suggest that the terrace system probably is older than 30 ka and that it formed about 60,000 yr after the last interglacial maximum. The terrace therefore probably formed during middle Wisconsin time, and its drift cover probably is late Wisconsin in age. The distribution of end moraines of Raven House age suggests that late Wisconsin glaciers in the Lituya Bay area formed separate piedmont lobes that extended as much as 16 km across the continental shelf from the present-day coast (Mann, 1986). Wave erosion of the Raven House drift unit probably resulted from isostatic depression due to crustal loading by glaciers, which was superimposed on slow tectonic uplift. Mann (1986) estimates that 60 to 100 m of isostatic depression would have been necessary to submerge and erode drift deposited at the time of the maximum late Wisconsin ice advance.

Till of the succeeding Eurhythmic readvance contains wood dated at 12.4 ka. Glaciers in the Lituya Bay area had receded to positions close to their present limits by that time (Mann, 1986).

Glaciers in Glacier Bay reached altitudes of 1,000 to 1,500 m during late Wisconsin time (McKenzie and Goldthwait, 1971);

they coalesced into a large ice stream that flowed through Glacier Bay and onto or across the continental shelf (Goldthwait, 1987). By about 13 ka, retreating glaciers had reached positions close to their present termini, and deglaciation was accompanied by marine transgression to as much as 60 m asl. The Brady Glacier, which drains into Icy Strait just west of Glacier Bay, reached altitudes of about 600 m asl along the coast during the late Wisconsin (Derksen, 1976, p. 13) and probably extended across the continental shelf at that time (Mann, 1986).

Raised beaches and bedrock terraces south of Glacier Bay record two marine incursions that probably resulted from ice loading (Ackerman and others, 1979). The highest terrace, at 12 to 15 m asl, was cut sometime between 13.4 and 9.2 ka; it records a halt or slowing in isostatic rebound.

Southeastern Alaska

The extent and history of glacial advances in most parts of southeastern Alaska are unknown. Regional morphology and limited field data suggest that the late Pleistocene glacier complex was formed by mountain ice sheets expanding out of icefields on the mainland and coalescing with small mountain ice caps on islands of the Alexander Archipelago (D. H. Mann, written communication, 1986). Outlet glaciers flowed through the fjord system that presently dissects the archipelago. Local valley glaciers and ice caps were generated on higher parts of the outer islands, but some local uplands and parts of the continental shelf probably remained ice free (see Fig. 1). This reconstruction is compatible with Mann's (1986) analysis of late Wisconsin glacier dynamics in the Lituya Bay area and with uplift curves and glacier flow patterns in coastal British Columbia (Warner and others, 1984; Barrie and Bornhold, 1989; Clague, 1989).

Peat beds beneath glaciomarine deposits near Juneau are older than 39 ka and may have been deposited during a middle Wisconsin interstade when glaciers had receded from inner fjords (Miller, 1973). The glaciomarine Gastineau Channel Formation subsequently was deposited during deglaciation and isostatic emergence about 13 to 9 ka. Miller's (1973) data show that emergent marine deposits as high as 230 m asl formed immediately after deglaciation and that subsequent isostatic recovery was interrupted by a stillstand or readvance of glaciers sometime after 13 ka.

The morphology of the Alexander Archipelago reflects intensive glacial erosion. Valleys connected by smoothly abraded passes cross most islands (Mann, 1986), and only a few areas that lie within precipitation shadows on large islands appear to have escaped inundation by regional ice streams or by locally generated alpine glaciers. Swanston (1969) describes two tills on Prince of Wales Island that differ in oxidation and therefore may represent two separate glacial advances. The younger till occurs only below 500 m asl and has a minimum limiting radiocarbon age of about 9.5 ka. Emergent marine deposits occur to heights as much as 150 to 210 m throughout this region (Mann, 1986).

SUMMARY AND DISCUSSION

Tertiary diamicts of probable glacial origin are widespread in Alaska, and they commonly represent several successive episodes of glaciation. The oldest glacial deposits in northern and central Alaska typically have been intensively eroded and occur only as erratic boulders redeposited in younger gravel; glacial deposits of younger Tertiary age more commonly occur as remnants of generally featureless drift. Tertiary glaciomarine deposits are more extensive around the Gulf of Alaska, where thick sections have been measured onshore (Plafker, 1981), and also are evident in seismic reflection profiles offshore (Carlson, 1989).

The general increase in glacial activity after about 2.5 Ma recorded in the Gulf of Alaska has possible counterparts elsewhere in Alaska, where glaciation of probable Miocene or early Pliocene age was followed by a second glacial episode late in the Pliocene. Tertiary glacial deposits in northern Alaska extend far beyond the most extensive drift sheets of Pleistocene age, but farther to the south they are exposed only locally and for short distances beyond the outer limits of younger drift. This distribution pattern strongly suggests that the Arctic Ocean was an important moisture source for glaciers in Tertiary time (Hamilton, 1986a, b), and it supports the arguments of Herman and Hopkins (1980) and Carter and others (1986b) that the Arctic Ocean probably lacked sea-ice cover until near the end of the Pliocene.

Glacial deposits of early Pleistocene age are identified by subdued morainal morphology, by association with ancient landscapes into which modern valley systems have been incised, and by the presence of alpine landforms that developed during a long postglacial interval of weathering and erosion. Few firm ages are available for these deposits. Glaciomarine sediments of the Yakataga Formation on Middleton Island were deposited during the interval from 2.2 to 1.0 Ma, and they record increasing intensity of glaciation upward in the section. Paleomagnetic data from the Kobuk River valley indicate that the Anaktuvuk River glaciation of the Brooks Range is older than 0.73 Ma.

Distribution of Anaktuvuk River drift and of possibly correlative deposits offshore in the Kotzebue Sound region (Decker and others, 1989) indicates that glaciers of early Pleistocene age were much more extensive throughout the Brooks Range than were glaciers of all succeeding advances. Similarly, extensive glacial deposits in the Yukon-Tanana Upland and the Kuskokwim Mountains were formed by ice caps that covered local highlands prior to incision of modern valley networks. The early Pleistocene glaciers and ice caps of northern and central Alaska probably required more abundant snowfall than at present.

The multiple glacial events in the Nenana River valley are difficult to correlate with the less complex early Pleistocene glacial records of tectonically stable areas farther to the north. Continued uplift of the Alaska Range, incision of the Nenana River valley, and segmentation of the valley by active faults evidently inhibited the obliterative overlap (Gibbons and others, 1984) that tends to simplify the glacial record in more stable regions.

Large kettles on broad river terraces that developed within drift of Anaktuvuk River age indicate that buried glacier ice remained for a long time after deglaciation of the foothills north of the Brooks Range. This persistent ice suggests the presence of continuous permafrost, which in turn generally is associated with perennial sea-ice cover. These relations suggest that the perennial ice cover of the Arctic Ocean could have developed during Anaktuvuk River time. However, deep weathering in drift deposits of northern and northwestern Alaska indicates that permafrost either disappeared or was significantly degraded during one or more interglacial intervals that followed the early Pleistocene.

Through much of northern and central Alaska, glaciers of middle Pleistocene age occupied valley floors close to modern levels and extended only short distances beyond the outermost limits of late Pleistocene ice advances. Glacier expansion followed a long interval of reduced glaciation during which mountain valleys were eroded to levels close to those of the present day. Two separate drifts having contrasting morphologies are present in some valleys, suggesting successive glacial advances spaced widely in time. Paleomagnetic determinations in the Kobuk River valley indicate that drift assigned to the middle Pleistocene formed during the Brunhes Normal-Polarity Chron (Hamilton, 1986a). Persistence of relict glacier ice in the Brooks Range and the Seward Peninsula suggests that much of the drift assigned to the middle Pleistocene in those regions could have been deposited as late as stage 6 and that continuous permafrost has persisted from that time until the present.

Unusually extensive glaciers or local ice caps on Seward Peninsula and in Kotzebue Sound are of middle Pleistocene age according to potassium-argon, amino-acid, and paleomagnetic age estimates. Glaciers also flowed across the Bering platform from the mountains of the Chukotsk Peninsula of Asia to St. Lawrence Island at about this time (Hopkins and others, 1972). Part of the Bering platform may have remained submerged and nourished locally large glaciers and ice caps, as advocated by Petrov (1967) and Hopkins (1973).

Little is known of the regional distribution of glaciers or moisture sources of middle Pleistocene age in southern Alaska. Drift around the head of Bristol Bay was derived from sources to the north and northeast (Detterman, 1986), a flow pattern that would be compatible with overflow of ice from upper Cook Inlet during either the Eklutna glaciation or an earlier ice advance that filled the inlet.

Drift of the penultimate glaciation is older than the age range of radiocarbon dating but, at least in northern Alaska, it appears to be younger than the maximum of the last interglaciation (stage 5e) because of its little-altered character, sharply defined outer boundaries, and poorly developed soil and weathering profiles. In addition, breakup of an ice sheet over the Canadian Arctic probably was responsible for deposition of the glaciomarine Flaxman Member of the Gubik Formation on the Arctic Coastal Plain sometime during stage 4 or late stage 5. Farther south, the penultimate glaciation may be older than stage 5e, but in some areas the distinction between late Wisconsin and older glaciations is blurred by the great extent of late Wisconsin ice advances and by persistence of glaciers through all or most of the Wisconsin stage. Two distinct penultimate-age glacial advances have been reported in many parts of Alaska (Table 1).

Alaska was partly deglaciated during the middle Wisconsin interstade, but parts of the state still remained ice covered. These areas incude some mountain valleys within the central Brooks Range (Hamilton, 1986a), parts of the St. Elias Mountains (Denton, 1974), upper Cook Inlet (Schmoll and Yehle, 1986), and parts of the Chugach Mountains near the lower course of the Copper River (Hamilton and Thorson, 1983).

The final major Pleistocene glacial episode in Alaska is correlated with late Wisconsin fluctuations of the Laurentide Ice Sheet, on the basis of numerous radiocarbon ages that generally show onset of glaciation at about 24 ka and deglaciation beginning at about 13.5 ka (Denton, 1974; Porter and others, 1983; Hamilton, 1982a, 1986a, b; Hamilton and Fulton, 1991). Throughout northern and central Alaska, the late Wisconsin glaciers were much less extensive than were those of the penultimate glaciation because of precipitation deficiencies resulting from emergence of the Bering platform, lower water temperatures in the North Pacific Ocean and Gulf of Alaska, and extensive seasonal and perennial sea-ice cover over the southern Bering Sea (Sancetta and others, 1985). Low precipitation and abundant sediment sources on barren glacial outwash and emergent marine shelves caused widespread deposition of eolian sand and silt (Hopkins, 1982; Hamilton and others, 1988; Lea and Waythomas, 1990). As many as four substages of late Wisconsin glaciation are reported from the Brooks Range, the Alaska Range, and the Bristol Bay region, but correlations among them are uncertain.

Glaciers of southern Alaska extended onto the continental shelf throughut the coastal area that stretches from the Aleutian Islands around the Gulf of Alaska into southeastern Alaska. Late Pleistocene ice caps may have covered much of the shelf south of the Alaska Peninsula, but farther to the east, glaciers probably reached the shelf edge only along a few U-shaped troughs. Most of those ice streams issued from major valleys that allowed passage of ice from the interior (Mann, 1986). Coastal areas probably were depressed isostatically 100 to 250 m, and records of deglaciation began at about 13 ka.

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MANUSCRIPT ACCEPTED BY THE SOCIETY OCTOBER 24, 1990

ACKNOWLEDGMENTS

Parts of this chapter have been condensed from the volume *Glaciation in Alaska* (Hamilton and others, 1986a). I am grateful to the Alaska Geological Society for their support of that publication and for permission to adapt several illustrations from it for the present paper. Helpful comments on sections of this chapter were contributed by J. M. Armentrout, J. L. Bischoff, J. Brigham-Grette, L. D. Carter, R. L. Detterman, C. H. Eyles, D. M. Hopkins, D. S. Kaufman, D. H. Mann, B. F. Molnia, G. Plafker, R. D. Reger, H. R. Schmoll, R. M. Thorson, J. A. Westgate, and J. R. Williams. An initial draft was reviewed by O. J. Ferrians, Jr. and R. M. Thorson. Illustrations were drafted by G. A. Lancaster and Inyo Ellersieck.

NOTES ADDED IN PROOF

This chapter was last updated in 1989. Since that date, nearly 200 reports and abstracts have been published on Alaskan glaciation and related late Cenozoic climatic changes. The space allowed for this note permits discussion of only a few of those studies.

Tertiary glaciation

Drill cores from the Arctic Ocean basin and erratics in transgressive deposits around its margins indicate episodes of regional cold climate and glaciation. The oldest ice-rafted material in the Canada Basin is of late Miocene age (Grantz and others, 1990), and sediments on Alpha Ridge indicate a subsequent late Pliocene cold interval about 2.48 to 2.08 Ma (Scott and others, 1989). Erratic clasts in deposits of the Colvillian transgression (sometime between 2.7 and 2.48 Ma) and in basal beds of the Bigbendian transgression (about 2.48 Ma) indicate that glaciers were present somewhere around the margins of the Arctic Basin at those times (Brigham-Grette and Carter, 1992). Erratic boulders in late Tertiary deposits of the central Arctic Coastal Plain are reported by Rawlinson (1993), but ages of those deposits are uncertain.

Drill cores from the Gulf of Alaska and the North Pacific, seismic stratigraphy along the Gulf of Alaska shelf (Carlson, 1989), and outcrop studies of the Yakataga Formation provide additional insights into Tertiary glaciation in the Gulf of Alaska region. Sediment cores show that glaciation extended back into the latest Miocene (von Huene, 1989), and that a later major episode of glaciation began about 2.6 Ma (Morley and Dworetzky, 1991; Krissek and others, 1993). Glacial maxima are recorded at 2.5 to 1.8 Ma and during the last 1.3 m.y. (Morley and Dworetzky, 1991).

Initial widespread appearance of ice-rafted debris in the Yakataga Formation was assigned an early middle Miocene (15 to 16 Ma) age by Marincovich (1990) on the basis of mollusc faunas, but alternatively may be late Miocene (5 to 6 Ma) according to the foraminiferal record (Eyles and others, 1991; Zellers, 1990). Renewed glaciomarine deposition in the late Pliocene (Zellers, 1990; Zellers and others, 1992) followed a middle Pliocene cool-temperate episode (Zellers and others, 1992). Stratigraphy and sedimentology of the Yakataga Formation are reviewed by Eyles and Lagoe (1990), C. H. Eyles and others (1991), and N. Eyles and others (1992).

Early and middle Pleistocene glaciations

A general climatic reconstruction for the Arctic Ocean borderland by Repenning and Brouwers (1992) indicates mild conditions about 2 Ma followed by oscillating cool and warm periods 1.7 to 1.2 Ma and then by more severe cold and glacial conditions after 1.1 Ma. Sediment cores from the Arctic Ocean basin show repeated intense glaciations within the past 0.8 to 1.0 m.y. (Bischof and others, 1992; Phillips and others, 1992). Although an early Pleistocene age has been assumed for extensive, subdued drift sheets in northern Alaska, the age of these deposits is still poorly constrained. Correlative drift on Banks Island is now known to be magnetically reversed, hence older than 780 ka (Vincent, 1990). Moraines assigned to the early Pleistocene Anaktuvuk glaciation in the northeastern Brooks Range have been strongly deformed by tectonism (Rawlinson, 1993).

Cores taken from Alpha Ridge indicate that middle Pleistocene glaciation may have been interrupted by a warm interval 0.6 to 0.4 Ma (Scott and others, 1989), and insolation values suggest that major glaciation in northern Alaska would have been unlikely during 575 to 250 ka (Bartlein and others, 1991). The Anvilian marine transgression of the Nome area, which may date about 410 ka (Kaufman, 1992), probably occurred within this interval.

The middle Pleistocene Nome River glaciation had been dated as between 580 and 280 ka (Kaufman and others, 1991), and a correlative glacial advance built the Baldwin Peninsula moraine (Huston and others, 1990) sometime between 400 and 300 ka (Roof and Brigham-Grette, 1992). A probable correlative glacier from Chukotka flowed eastward across the Bering platform and encroached on St. Lawrence Island sometime prior to the last interglaciation (Brigham-Grette and others, 1992; Heiser and others, 1992).

The Old Crow tephra, which is widespread in Alaska and the Yukon Territory (Waythomas and others, 1993), generally occurs near the base of organic-rich deposits of interglacial character (Hamilton and Brigham-Grette, 1992). The fission-track age of 149 ± 13 ka for this deposit is supported by an age estimate of 135 ± 5 ka on stratigraphic grounds (Hamilton and Brigham-Grette, 1992) and by bracketing thermoluminescence ages of 110 ± 32 and 140 ± 30 ka (Berger and others, 1992). The Old Crow tephra is an important stratigraphic marker for the stage 6/5 glacial-interglacial transition (Begét and others, 1991; Hamilton and Brigham-Grette, 1992).

New studies of glacial sequences that include early and (or) middle Pleistocene as well as late Pleistocene ice advances include Peck and others (1990), Reger and Bundtzen (1990), Reger and Pinney (1993), and Waythomas (1990).

Late Pleistocene glaciation

Glaciation of probable isotope stage 4 or late stage 5 age was extensive in the Bering Strait region, where ice again advanced from Chukotka to St. Lawrence Island (Brigham-Grette and others, 1992; Hopkins and others, 1992). This advance may correlate with the glaciomarine Flaxman Formation of northern Alaska, which probably was deposited about 85 to 80 ka during isotope stage 5a (Carter and Whelan, 1991), and with extensive glaciation of the western Canadian Arctic (Vincent, 1992).

Multiple stades of Late Wisconsin glaciation are recognized and partly dated in the Cook Inlet region (Reger and Pinney, 1993) and on the Alaska Peninsula (Pinney and Begét, 1991). The most extensive advances ended by about 14 ka, when pollen records show a marked change from full-glacial to late-glacial environments (e.g., Anderson, 1991). Subsequent glacial readvances include the 12 to 10 ka Ukah Stade on the Alaska Peninsula (Pinney and Begét, 1991), a readvance about 12 ka at Valdez (Reger, 1991), and readvances about 13.4, 12 to 11.7, and 11 to 10 ka on Kodiak Island (Mann, 1992). Some of these readvances could coincide with Younger Dryas cooling, for which pollen evidence has been reported by Engstrom and others (1990).

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Printed in U.S.A.