Summary of Quaternary geology of the Municipality of Anchorage, Alaska

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Abstract

Quaternary geology of the Upper Cook Inlet region is dominated by deposits of glacier retreats that followed repeated advances from both adjacent and more distant mountains. At several levels high on the mountains, there are remnant glacial deposits and other features of middle or older Pleistocene age. Late Pleistocene lateral moraines along the Chugach Mountain front represent successively younger positions of ice retreat from the last glacial maximum. As the trunk glacier retreated northeastward up the Anchorage lowland, Cook Inlet transgressed the area, depositing the Bootlegger Cove Formation and Tudor Road deposits. The glacier then readvanced to form the latest Pleistocene Elmendorf Moraine, a prominent feature that trends across the Anchorage lowland. Extensive alluvium was deposited both concurrently and somewhat later as Cook Inlet regressed. Mountain valleys contain (1) locally preserved moraines possibly of early Holocene age; (2) poorly preserved moraine remnants of older late Holocene age; and (3) well-preserved moraines formed mainly during the Little Ice Age. Glaciers still occupy large parts of the mountains, the upper ends of some mountain valleys, and small cirques. Holocene landslide deposits, including those formed during the great Alaska earthquake of 1964, occur throughout the area, especially along bluffs containing the Bootlegger Cove Formation. © 1999 Elsevier Science Ltd and INQUA. All rights reserved.

1. Introduction

The Municipality of Anchorage, Alaska, is located at the head of Cook Inlet where it bifurcates into Knik and Turnagain Arms, about 275 km up the inlet from the Pacific Ocean (Fig. 1). With an area of about 5000 km², it is one of the largest cities in the United States (larger than the State of Rhode Island); its population is only about 250,000, but comprises nearly half the population of Alaska. The Municipality encompasses both the Anchorage lowland where most of its urbanized part lies and a much larger area within the Chugach Mountains. Much of the mountainous part of the Municipality is occupied by Chugach State Park, and most of the remainder lies within the Chugach National Forest. This report summarizes geologic mapping, stratigraphic studies, and landslide investigations by Ernest Dobrovolny and Schmoll, 1965–1973, continued intermittently by Schmoll and Yehle through 1997, and by Updike, 1976–1987.

1.1. Physiography

The Anchorage lowland is an informal subdivision of the Cook Inlet-Susitna Lowland (Wahrhaftig, 1965) which corresponds roughly with Cook Inlet basin. This basin is a major intermontane structural and physiographic basin about 325 km long and 95 km wide, occupying about 31,000 km², of which about 70% is covered by Cook Inlet. It is one of a series of basins that lie inland from the coastal mountains bordering southern Alaska. Unlike either the entirely landlocked Copper River basin to the northeast, or the present marine basin beneath Shelikof Strait to the southwest, Cook Inlet basin has fluctuated between terrestrial and estuarine environments during the Quaternary. The interplay of these two environments has given rise to a long series of shoreline changes as the configuration of the inlet changed through time, complicating the glacial history of the region.

The present Cook Inlet is a brackish, tide-dominated estuary (Pritchard, 1967; Gatto, 1976) that ranges in depth from less than 18 m in near-shore areas and in isolated shoals, to about 45 m in troughs. Salinity in the upper inlet, influenced by a large influx of fresh water, ranges from 29 parts per thousand (ppt) near Kalgin

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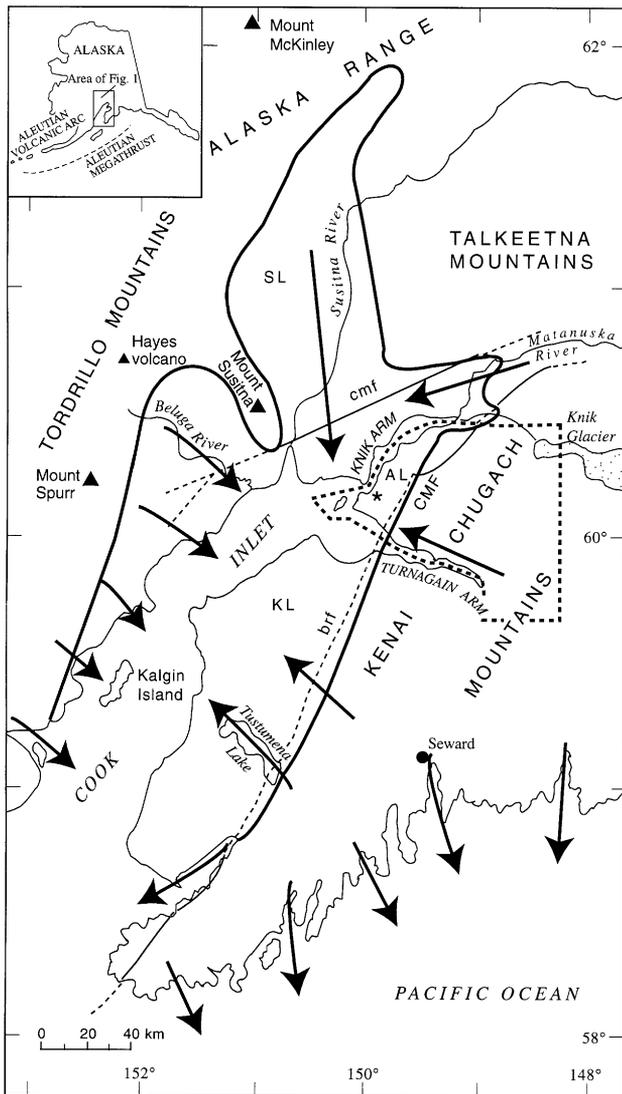* Corresponding author. Tel.: + 1-303-273-8634.

Fig. 1. Index map showing location of the Municipality of Anchorage (dashed line) within the Cook Inlet region, Alaska; downtown Anchorage is centered at the asterisk. Heavy line indicates margin of Cook Inlet-Susitna Lowland including three informal subdivisions: AL, Anchorage lowland; KL, Kenai lowland; SL, Susitna lowland. CMF, Chugach Mountain front; cmf, Castle Mountain fault; brf, Border Ranges fault. Arrows indicate principal glacier lobes that entered the Cook-Inlet-Susitna Lowland and the Pacific Ocean during the Pleistocene.

Island to 8 ppt at the mouth of the Susitna River (Kinney et al., 1970). Diurnal tides are among the highest in the world, with a normal range of about 9 m, but as much as 11.6 m during maximum spring tides. Low tide in Turnagain and Knik Arms is marked by a tidal bore as much as 1.5 m high that advances up each arm at about 4.5 m/s.

The upper part of Cook Inlet is bounded by filled embayments, lowlands, and plateaus. The embayments contain Holocene alluvial and tidal deposits as much as a few hundred meters thick. Here, the ground surface is

graded to a level just above that of present Cook Inlet and rises to less than 60 m in altitude. The surfaces of the old embayments are nearly flat, except where tidal channels rarely more than 1 m deep are incised. The filling of the embayments caused a nearly continuous, slow retreat of the shoreline throughout the Holocene.

The lowlands range from about 25 to 300 m in altitude; they are separated from the inlet by bluffs 25–60 m high that contain the best exposures of Pleistocene deposits in the region. These deposits are as much as 300 m thick (Freethey and Scully, 1980), mainly glacial and glacio-estuarine, and include the geologically significant Boot-legger Cove Formation. Slopes are gentle, but some low hills have relief of as much as 30 m.

The bluffs bordering plateaus are not much higher than those bordering the lowlands, but the plateaus rise gently to as much as 1000 m in altitude. Pleistocene deposits are thin and the underlying Tertiary continental rocks crop out locally.

The Chugach and Kenai Mountains, with high peaks in the range of 2000–2500 m in altitude, rim the Cook Inlet basin on its southeast side. Within the Municipality of Anchorage mountains rise abruptly above the Anchorage lowland along the Chugach Mountain front. The northwest side of the basin is marked by a more heterogeneous group of mountains, most prominently the Tordrillo Mountains. On clear days three peaks in excess of 3350 m form a spectacular backdrop along the west side of the basin. The Alaska Range lies to the north where Mt. McKinley, the highest peak in North America, reaches an elevation of 6194 m.

1.2. Bedrock and structure

Bedrock of the Chugach Mountains within the Municipality of Anchorage consists of structurally complex and variably metamorphosed sedimentary and igneous rocks (Capps, 1940; Clark, 1972; Tysdal and Case, 1979; Winkler, 1992; Plafker et al., 1994). These rocks, like bedrock throughout the Cordilleran region of western North America, are divided primarily into tectonostratigraphic terranes, each of which has a distinct suite of geologic formations with a common tectonic history. In this area two such allochthonous terranes accreted to the North American tectonic plate in late Mesozoic to early Tertiary time (Coney and Jones, 1985; Jones et al., 1987). Most of the rocks within the area are part of the Chugach terrane; they include the highly variable McHugh complex (Clark, 1973), and farther east, the Valdez Group, a deep-water marine turbidite sequence (Clark, 1972; Budnik, 1974a,b; Nilsen and Bouma, 1977). Rocks of the Peninsular terrane occupy a narrow belt along the Chugach Mountain front and extend as well to the northwest beneath the Anchorage-Susitna Lowland. The exposed rocks include principally greenstone, greenschist, and gneiss, as well as a narrow, discontinuous belt

of ultramafic rocks (Burns, 1985; Newberry, 1986). Felsic to intermediate hypabyssal intrusive rocks of Tertiary age occur in a few places.

Continental sedimentary rocks of Tertiary age, the Kenai Group (Calderwood and Fackler, 1972), overlie the metamorphic rocks beneath much of the Anchorage lowland. They occur at relatively shallow depth along most of the its eastern margin but crop out only in the vicinity of Eagle River (Fig. 2A). They aggregate as much as several thousand meters in thickness in the central part of the basin (Ehm, 1983) and consist mainly of sandstone, siltstone, claystone, and conglomerate. The Tertiary rocks contain economically significant oil, gas, and coal in some formations. Locally they contain probable glacial deposits including diamicton beds exposed on the west side of Cook Inlet (Schmoll et al., 1984) and diamicton interpreted from drill-hole data on the Kenai lowland (Boss et al., 1976).

Major faults in the vicinity of Anchorage lie both between and within tectonostratigraphic terranes. The Border Ranges fault (Fig. 1) separates Chugach and Peninsular terranes (MacKevett and Plafker, 1974); it roughly parallels the Chugach Mountain front within exposed bedrock northeast of Eagle River, but it is concealed beneath lowland deposits southwest of there. Within the Chugach terrane, the Knik fault marks the boundary between the Valdez Group and the older McHugh Complex which is thrust over it. The Castle Mountain fault (Fig. 1) lies northwest of Anchorage within the Peninsular terrane. It is the only fault in the region known to have been active in Holocene time (Detterman et al., 1974; Haeussler, 1994). It is a present locus of seismicity, and epicenters of some earthquakes felt in Anchorage are associated with it (Lahr et al., 1985). Principal seismic activity, however, is related to the Aleutian megathrust (Fig. 1, inset) that lies seaward of the southern Alaska coast and marks the present boundary between the North American and Pacific tectonic plates. From this boundary, the present subduction zone descends beneath southern Alaska, including the Cook Inlet basin (Plafker et al., 1982). The great Alaskan earthquake of 1964 with moment magnitude 9.2 (Kanamori, 1977) struck southern Alaska on March 27 of that year and the Anchorage area suffered great damage; intensive geologic and other investigations that soon followed were summarized by Eckel (1970).

1.3. Quaternary deposits

Overlying the Tertiary rocks in upper Cook Inlet basin are complexly interbedded glacial, glacioestuarine, and related alluvial deposits of Quaternary age that thin toward the margins of the basin (Schmoll et al., 1986). These deposits, together with deposits in glacial landforms on the margins of the basin and within numerous small valleys in the mountains, record a series of gla-

ciations during the Quaternary. Even though central basin stratigraphy is locally well exposed, it is difficult to correlate those deposits with basin-margin and mountain-valley landforms. Each data set seems to yield its own interpretations that are not necessarily compatible with those of the other. Thus, varying interpretations of glacial history and shoreline reconstruction have evolved over the years (Miller and Dobrovolny, 1959; Karlstrom, 1964; Cederstrom et al., 1964; Schmoll and Dobrovolny, 1972; Schmoll and Yehle, 1986; Reger and Updike, 1983, 1989; Reger et al., 1995, 1996). The consensus, however, is that glaciers advanced into Cook Inlet basin, and into ancestral Cook Inlet itself, from a variety of mountain sources each situated differently with respect to the basin (Fig. 1). Consequently, each source area provided a glacier lobe that behaved somewhat differently with respect to advance, retreat, and contact with inlet water and with neighboring ice lobes.

Glacial deposits preserved in stratigraphy beneath the Anchorage lowland are the products of several trunk glaciers that advanced from mountains to the north and east, as well as of smaller glaciers that advanced from local mountain valleys. In and along the Chugach Mountain front, however, evidence for older glaciations is based largely on relict landforms rather and the deposits within them than on stratigraphy. In general, the landforms resulting from the most recent glacier to occupy a given area are the most likely to be preserved. Consequently, the oldest glacier advance for which there is evidence is probably the most extensive one, and successively younger advances from which landforms and their deposits still survive were successively less extensive. Landform evidence is lacking for intervening advances that were less extensive than following advances. Such intervening, but lesser ice advances have been recorded in deep-sea drill holes, for example, and they probably occurred here as well. In the Anchorage lowland, stratigraphic sequences known from subsurface investigations, mainly water-well logs, indicate a series of five to least seven glacial advances separated by deposits indicative of ice withdrawal (Trainer and Waller, 1965; Schmoll and Barnwell, 1984). These sequences cannot, however, be correlated with confidence to glacial landforms.

Glacial, glacioestuarine, and glacioalluvial deposits that resulted from the long sequence of varied glacier, estuarine, and subaerial positions are shown in a generalized way in Fig. 2A and B. The wider, southern part of the Anchorage lowland is shown in more detail in Fig. 3. Some deposits that have been mapped separately at 1 : 25,000 scale are grouped into more broadly defined units. The deposits portrayed in Figs. 2 and 3 can be divided into eight categories: (1) Remnant glacial deposits and features of glacial erosion of poorly defined Pleistocene age that occur relatively high on the Chugach Mountain front and on interfluve ridges and some mountain tops back from the front; (2) late Pleistocene lateral

moraines that occur along the Chugach Mountain front and on the sides of larger mountain valleys; (3) late Pleistocene glacioestuarine and glacioalluvial deposits of the Anchorage lowland; (4) the Elmendorf Moraine of latest Pleistocene age (the only prominent end moraine in the area) together with an extensive area of ground moraine; (5) Holocene-age end and lateral moraines that are restricted to mountain valleys; (6) colluvial deposits that occur extensively in the mountains and locally in the lowland; (7) peat, pond, eolian and other mainly Holocene deposits including major earthquake-induced landslides in the lowland; and (8) estuarine deposits that border Knik and Turnagain Arms.

Most of our understanding of the Quaternary geology summarized above was deduced during geologic mapping of the Anchorage lowland and adjacent areas. Much
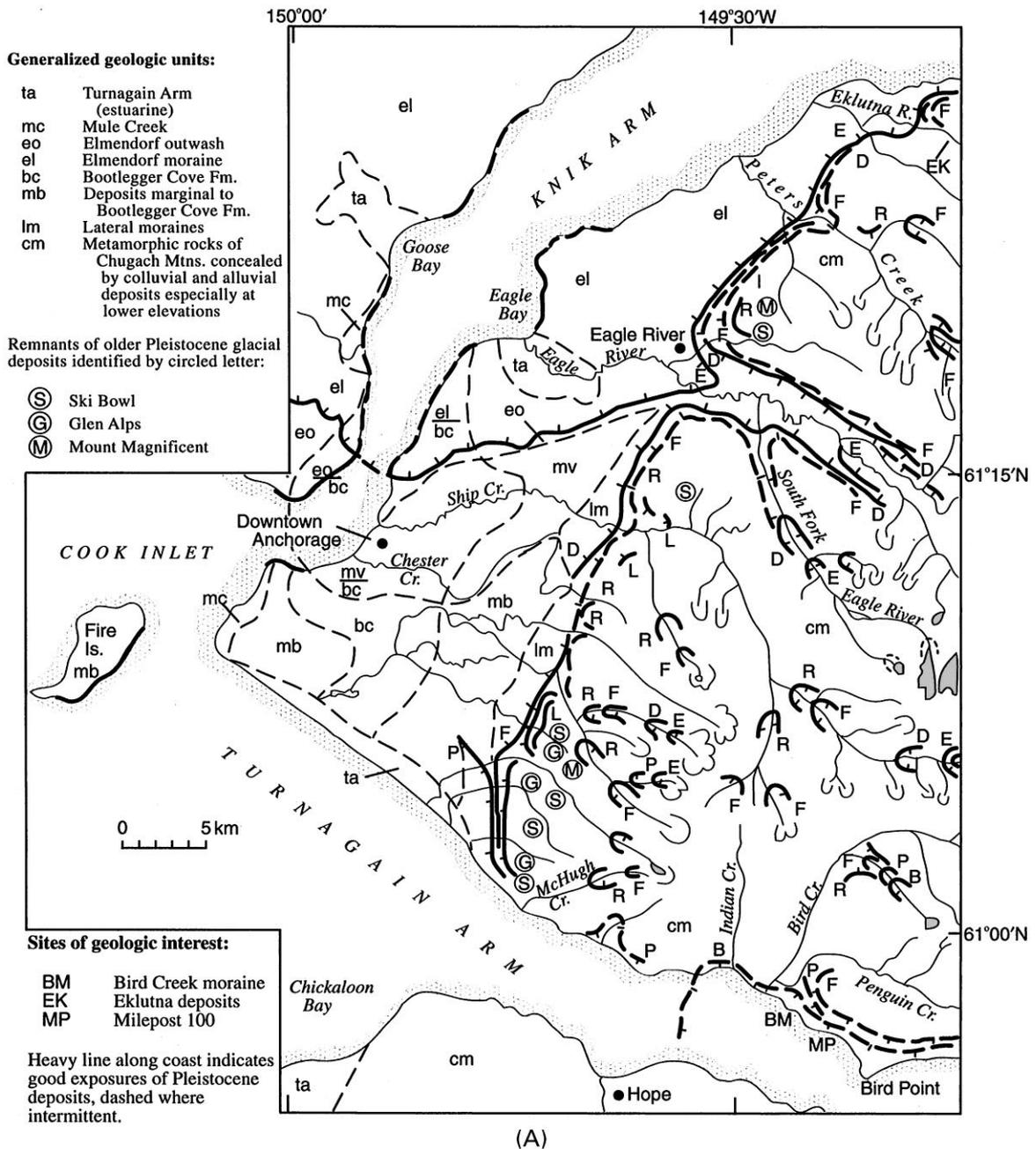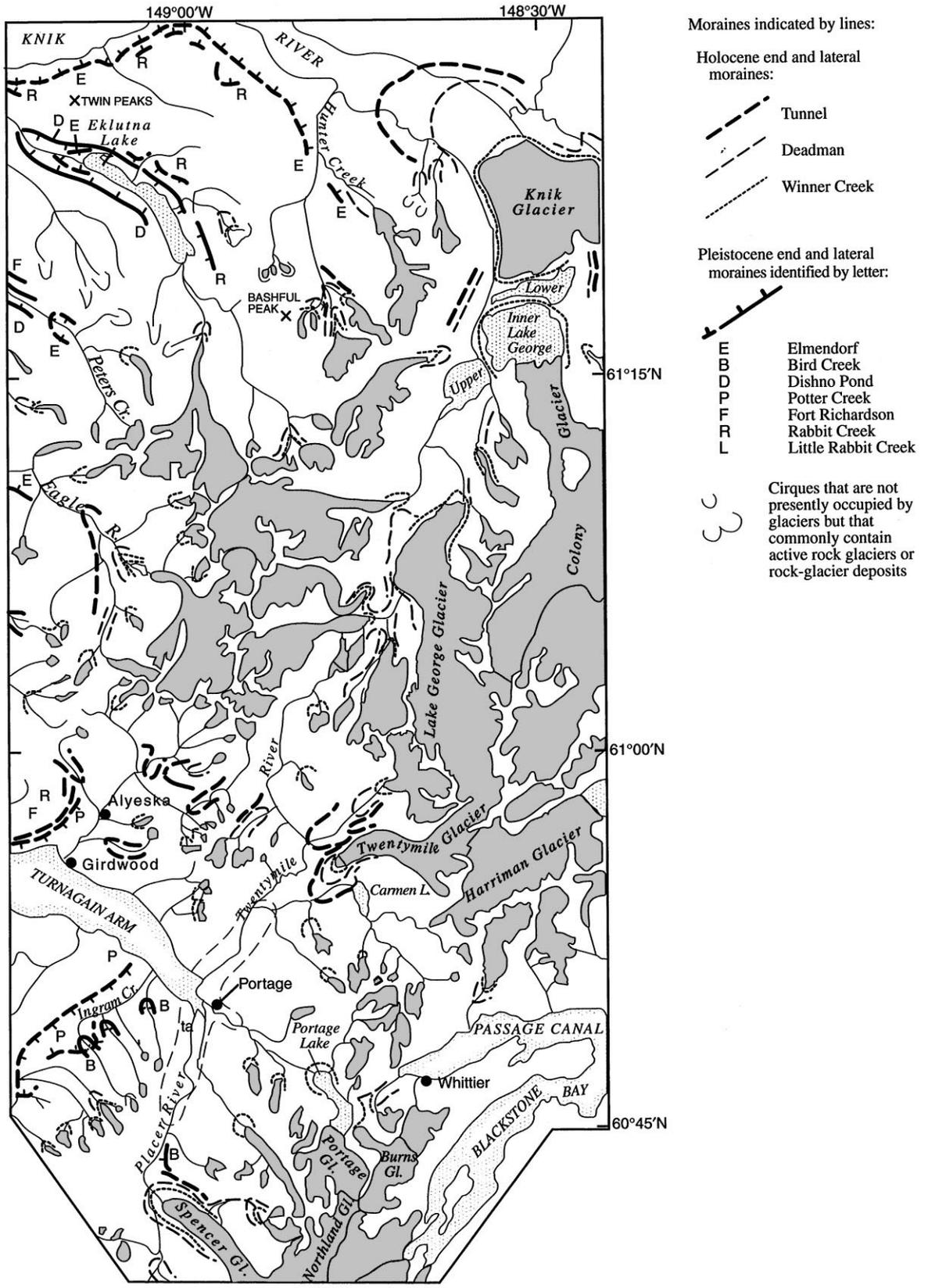

Fig. 2. (A). Map of the western part of the Municipality of Anchorage and vicinity showing highly generalized geologic units in the Anchorage lowland and principal end and lateral moraines and present-day glaciers in the Chugach Mountains. The Chugach Mountain front lies along and just southeast of the prominent line of lateral moraines that trends diagonally across the map. (B). Map of the eastern part of the Municipality of Anchorage and vicinity showing principal end and lateral moraines and present-day glaciers. Metamorphic rocks partly concealed by colluvium underlie the entire area.

(B)

Fig. 2. (*Continued*).

Geologic Sites:
ia   International Airport silt
AR   Abbott Road
KV   Klatt Valley Road
BR   Tudor Road
CD   Fire Island Drive
CPR   Bulk Products Road
ERF   Eagle River Flats
EV   Early View
FM   Flattop Mountain
FR   Fort Richardson kames
GA   Glen Alps
GV   Golden View
HR   Huffman Road
LK   Lower Klatt delta
MR   Muldoon Road
NC   Nunaka channel
PH   Potter Hill
PW   Point Woronzof
RC   Rabbit Creek
RJ   Russian Jack diamicton
SH   Spring Hill channel
SL   Sand Lake delta
UK   Upper Klatt delta
WS   Winchester Street

Old landslide deposits:
OLD   Dishno Pond
OLI   Inlet View School
OLT   Third Avenue

Principal 1964 earthquake-related landslides:
QLRP Roosevelt Road
QLBP Bridge Point
QLFA Fourth Avenue
QLLS L Street
QLTH Turnagain Heights

pc   Potter Creek
fr   Fort Richardson
rc   Rabbit Creek
gl   glaciolacustrine deposits
      related to above moraines

Pleistocene moraine deposits:
lr   Little Rabbit Creek
sb   Ski Bowl
ga   Glen Alps
mm   Mount Magnificent

Late Holocene estuarine deposits:
ta   Turnagain Arm
Holocene and latest Pleistocene alluvium:
er   Eagle River
cc   Clunie Creek
sc   Ship Creek
mv   Mountain View
nc   Nunaka
ns   North Fork and South Fork
      Campbell Creek
Late Pleistocene glacioestuarine and deltaic deposits:
mc   Mule Creek
bc   Bootlegger Cove Fm. partly
      overlain by Hood Lake sand

Latest Pleistocene glacioalluvium:
sx   Sixmile Lake

Late Pleistocene lateral-moraine and glaciolacustrine deposits:
dp   Dishno Pond

Latest Pleistocene moraine deposits:
em   Elmendorf (end) moraine
el   Elmendorf ground moraine

149°50' W   149°40' W

61° 20' N
61° 15' N
61° 10' N
61° 05' N

Fig. 3. Map of the Anchorage lowland south of the Eagle River showing generalized Quaternary geology and location of geologic sites.

of this mapping is based on interpretation of air photos, the inferred contacts between geologic units commonly coinciding with a break in slope bounding a landform. Correlation of landforms with geologic deposits depends on verification of mappable geologic units in stratigraphic exposures on the ground. There are few sites in the Anchorage lowland where such relationships can be demonstrated. Their location is dependent on the vagaries of natural erosion and of anthropogenic exposure of geologic material. The exposures tend to be ephemeral; many sites that provided important information no longer exist because of urbanization, slumping, or continued erosion. In the discussions that follow, generally accessible sites that have been important to the geologic interpretations are described briefly and their locations are shown by letter symbol in Figs. 2 and 3; the quadrant of each figure is also given to aid in location.

2. Pleistocene glacial deposits

2.1. Older Pleistocene glacial features

Evidence that glaciers once covered areas high on the Chugach Mountain front include (1) glacially smoothed ridge crests and shoulders at and near the basinward ends of mountain interfluve ridges, (2) small areas of ground moraine, and (3) well-developed, small saddles on some ridges, some of which are underlain by gravel. There are also a few discontinuous lateral-moraine remnants. We assign these features to three named groups termed, from highest to lowest (oldest to youngest), the Mount Magnificent, Glen Alps, and Ski Bowl moraines and related deposits (Schmoll and Yehle, 1986; Yehle et al., 1992). Each group lies on a southwest-descending gradient along the length of the mountain front that is subparallel to gradients of the better preserved lateral moraines lower on the slope. Because these gradients are separated from each other, and from the Little Rabbit Creek moraines downslope, by about 100 to 150 m, we regard deposits along each gradient as representing a different major episode of glaciation. At altitudes higher than any of the named features, a few glacial erratics have been found on planed surfaces such as that of Flattop Mountain (altitude 1070 m; site FM, Fig. 3-se) which is higher than any other such surface in this part of the Chugach Mountains. Only a few high-lying saddles on some nearby mountain ridges are thought to be of the same age. Such features are similar in altitude and mode of occurrence to those assigned to the Mount Susitna glaciation by Karlstrom (1964) because of their location on top of that mountain (Fig. 1) on the west side of the Susitna lowland, but we have not used that term for these few features in the Chugach Mountains. As we have no direct evidence for the age of any of these deposits, we consider them undifferentiated Pleistocene. Future stud-

ies using cosmogenic nuclides might shed light on the age of these older glacial features.

The Mount Magnificent ground moraine is best developed on a glacially planed surface high on the Chugach Mountain front (altitude about 975 m) northeast of Eagle River (M, Fig. 2A-ne; Schmoll et al., 1971; Yehle and Schmoll, 1989). The diamicton comprising the till on this surface is somewhat compact and more clayey than most tills in the area and granitic boulders are commonly scattered about; bedrock rubble is a locally important constituent. Lateral-moraine remnants correlated with these deposits have been recognized in only a few widely scattered places.

A good site to observe some of these older features is the typical locality of the Glen Alps deposits (GA, Fig. 3-se). It is one of a series of glacially smoothed bedrock knobs whose summits are covered by thin patches of glacial drift. Bedrock is exposed locally on the numerous small knolls, and glacial deposits may be restricted to areas between the knolls. The drift is commonly quite rubbly and locally may be mostly rubble derived from adjacent bedrock, with only a few erratics. Glen Alps lateral or ground moraines can be identified only at scattered localities elsewhere. Glacially planed areas similar to this one extend along the Chugach Mountain front roughly parallel to the gradient of similar Ski Bowl features that lie about 100 m downslope.

The typical locality of Ski Bowl moraines and associated features is a ski area north of Ship Creek (S, Fig. 2A-ne) located within an outlying part Fort Richardson. The area is now called Arctic Valley, but we have retained its former name because of our early use of it for the moraines there. Fairly well-formed lateral moraines at an altitude of about 700 m are now within a restricted area off the public road. At the ski area are mainly kame-fan deposits that were graded to the glacier margin. Gravel and sand are well exposed near the end of the access road; they are more oxidized than similar materials downslope.

A discontinuous series of moderately subdued lateral-moraine remnants lies about 100 m lower on the slope than the Ski Bowl features. We term these Little Rabbit Creek moraines, from the creek of that name. These moraine remnants might be significantly older than the lower lying, better preserved Rabbit Creek moraines. Because there is little evidence that early Wisconsin glaciers advanced beyond those of the late Wisconsin, a pre-Wisconsin (Illinoian [greater than 130 ka]) age is preferred, and these deposits are included here with the older, higher-lying features.

Older glacial deposits in stratigraphic sequence are found only along the Eklutna River where they crop out over a distance of about 6 km (site EK, Figs. 2A-ne). They are termed informally Eklutna deposits (Yehle and Schmoll, 1987) but were previously mapped as Knik glacial drift by Updike and Ulery (1983). They include

moderately compacted and slightly indurated gravel, silty sandy gravel, and gravelly diamicton that are in part well bedded and that have a distinctive yellowish–gray color. The beds dip moderately downvalley, suggesting a source up valley and possibly a deltaic origin for some of the beds. These are the deposits on which the concept of the Eklutna glaciation of Karlstrom (1964), p. 33) was based, and which he thought were of Illinoian age. We find no evidence for such a definitive correlation, nor can the age relationship to named lateral moraines along the Chugach Mountain front be demonstrated.

2.2. Interglacial deposits

There are few places within the Municipality of Anchorage where deposits interpreted as interglacial are recognized, and unlike the glacial features discussed above, none have geomorphic representation. Exposures at two sites along Knik Arm reveal probable older interglacial deposits whereas the last interglacial may be represented at three sites near Eagle River.

Exposed only in the lower beach at Point Woronzof (Fig. 3-sw) and at very low tides along Eagle Bay (Fig. 2A-n) are strongly oxidized and somewhat indurated sand and gravel, silty clay, and silty fine sand; peat is present at the Eagle Bay site as well. The lack of poorly sorted material such as diamicton and the occurrence of peat, in particular, suggest that these deposits are interglacial rather than glacial. Similar appearing deposits were first reported by Karlstrom (1964) from the west side of Knik Arm just northeast of Goose Bay (Fig. 2A-nw), also exposed only at very low tides. There, the deposits include finely comminuted shell fragments and twigs within silty clay as well. Karlstrom regarded these fine-grained deposits as equivalent to the glacioestuarine Bootlegger Cove Formation, but the nature of the fine-grained deposits with scattered organic debris suggests rather that they probably are interglacial pond deposits. An $^{40}Ar/^{39}Ar$ age of 378 ± 0.67 ka for volcanic ash within the Goose Bay peat (Reger et al., 1995, 1996) that overlies these deposits indicates that the pond deposits are probably middle Pleistocene in age. We infer that the Eagle Bay deposits, and possibly those at Point Woronzof as well, could be that old also.

Near the community of Eagle River (Fig. 2A-ne) there are two outcrops of interglacial deposits along the Eagle River, one just south of the community and the other about 4 km downstream (Miller and Dobrovolny, 1959, pp. 16, 18). The deposits include silt and clay (with some fine sand beds) and intermixed or interbedded organic material, mainly peat and twigs. A roadcut about 3 km north of the community includes similar material (Yehle and Schmoll, 1989). The organic material at all three sites has an age beyond the range of the radiocarbon method. The deposits underlie those mapped at the surface as Elmendorf ground moraine, although older till might be present at depth. None of these deposits exhibit the indurated and oxidized character of the material exposed locally along Knik Arm, and we infer that the deposits near Eagle River probably represent the last interglacial period.

2.3. Late Pleistocene moraines of the Chugach Mountain front

Four morainal systems consist of locally well preserved lateral moraines on the slopes along the Chugach Mountain front; they are separately named on the basis of their position along the front. From oldest to youngest, these are the Rabbit Creek, Fort Richardson, Dishno Pond, and Elmendorf systems. They are arranged somewhat en echelon, with successively older moraines dominant toward the southwest. Each system consists of deposits in (1) two or more individual moraine ridges, (2) kame fields and meltwater channels within the confines of the lateral moraines, and (3) kame terraces and kame fans deposited outbound from the moraine ridges. Correlations with moraines in mountain valleys are made on the basis of tracing lateral moraines from the mountain front into the valleys, or on the basis of lake and (or) alluvial deposits in the valleys that seem to extend between them.

Rabbit Creek moraines, well developed where they cross that creek, merge with moraines from Turnagain Arm valley; moraines curve southwestward out of the south side of the valley and extend onto the northern part of the Kenai lowland. There, they curve gently away from the mountain front, but do not form an end moraine across the lowland; instead, the landforms become increasingly more rounded and subdued with distance from the front until their identity as morainal features is lost. We refer to this type of landform modification, found at several localities around the margin of Cook Inlet basin, as "feathering out". The glacier that impinged on the slope of the mountain front farther northeast is interpreted as having terminated in glacioestuarine water. Lateral moraines can be traced discontinuously northeastward from Rabbit Creek to about Eagle River. They are also tentatively identified in the vicinity of Hunter Creek (Fig. 2-nw).

The Rabbit Creek moraine can be observed east of where Rabbit Creek Road climbs past a narrow Fort Richardson moraine, curves north, and runs along the crest of the moraine (site RC, Fig. 3-se). The relatively level terrain upvalley from the moraine is underlain by glaciolacustrine deposits that accumulated when Little Rabbit Creek valley was blocked by the lowland glacier. A coeval valley glacier (too small to show in Fig. 3) terminated about 4 km farther up Little Rabbit Creek valley. Good exposures of glaciolacustrine deposits are located along the original alignment of Rabbit Creek Road north of Little Rabbit Creek, now a side road; the

material consists of fine sand with some interbedded silt and a few stony horizons.

Next downslope are the Fort Richardson lateral moraines, so named because they are prominently developed in the formerly more extensive southern part of that military reservation; they follow a similar pattern and most of them feather out between Chester Creek and Rabbit Creek. Only the highest-lying Fort Richardson lateral moraine seems to join a moraine that emanates from Turnagain Arm valley, implying that at this time lowland and valley glaciers joined; the combined glacier also terminated in ancestral Cook Inlet, probably not far to the southwest. To the northeast Fort Richardson lateral moraines can be traced well up Eagle River valley, and discontinuously farther northeast along the Chugach Mountain front to the vicinity of Eklutna River (Fig. 2A-ne).

The lower-lying parts of the Fort Richardson lateral-moraine system are dominated by kames that trend parallel to the lateral-moraine ridges just upslope. The kame field at site FR (Fig. 3-se) is one of several along the Chugach Mountain front and can be observed along several roads in this vicinity. Each kame field is best developed southwest (down ice) of a major mountain valley, and presumably represents drainage from the partially blocked valley that penetrated the margin of the lowland glacier. This kame field probably derives largely from the valleys of both North and South Forks of Campbell Creek. Kame deposits have been exposed when exploited locally as sources of gravel. Many kames, however, contain much irregularly interbedded sand and silt and cores of diamicton, making quantity of these potential resources unpredictable. Residential and parkland developments now preclude most further gravel extraction in this area.

The Dishno Pond lateral moraines are named after a pond just north of Ship Creek (Fig. 3-ne). They are the next features downslope from the Fort Richardson lateral moraines along the mountain front; they extend about 15 km southwest from Eagle River valley to near South Fork Chester Creek where they also feather out. They also extend into Eagle River valley and northeast about as far as Peters Creek. Beyond there, distinction between the various lateral moraines becomes uncertain both because moraines are not continuously preserved and because they converge in altitude northeastward.

Dishno Pond moraines are well formed where they cross the mouth of Ship Creek valley; Arctic Valley Road provides a good access as it climbs the Chugach Mountain front along the north side of Ship Creek (site AV, Fig. 3-ne). Like most of the mountain valleys, Ship Creek valley is hanging with respect to the lowland and graded to lateral moraines along the mountain front. The lake that formed in the valley when it was blocked by both Dishno Pond and Fort Richardson glaciers is here named glacial Lake Ship. South of Ship Creek, the

Dishno Pond moraines descend and feather out where there is a kame field that probably developed when Ship Creek water entered the glacier during its waning stages. The glacier formed no end moraine in this vicinity because it terminated in Cook Inlet.

The prominent Elmendorf Moraine (em, Fig. 3-ne,nw) is a formal geographic feature named after the Air Force Base on which it is best developed (Miller and Dobrovolny, 1959). It extends across the Anchorage lowland and continues on the other side of Knik Arm, marking the terminal position of an ice lobe from the Knik and Matanuska valleys. Where the Elmendorf Moraine intersects Knik Arm, deposits of the moraine may be seen to overlie the glacioestuarine Bootlegger Cove Formation. In the vicinity of Eagle River valley, the massive end moraine merges with lateral moraines on the lower slope of the Chugach Mountain front. They can be traced as far northeast as Hunter Creek valley (Fig. 2B-n), but in that vicinity lateral moraines mapped as Elmendorf may represent the Dishno Pond and (or) Fort Richardson glaciers as well, as indicated above.

North of the end moraine on the Anchorage lowland there is an extensive area of ground moraine (el, Fig. 3-ne,nw) characterized by drumlinoid hills, kames fields, and meltwater channels. Some channels formed or were enlarged well after the glacier withdrew and contain Clunie Creek alluvium (Fig. 3-ne; Yehle and Schmoll, 1989) that was probably deposited by streams from the Peters Creek and Eklutna River valleys.

In the vicinity of Eagle River, the glacier apparently was able to block drainage in that valley and form a lake which we here name glacial Lake Eagle. Deposits of the lake can be traced upvalley to lateral moraines thought to be equivalent to the Elmendorf Moraine. When Dishno Pond and earlier glaciers in Eagle River valley joined lowland glaciers, no lake could form; there may have been other intervening times, however, when valley and lowland glaciers separated and earlier incarnations of glacial Lake Eagle did form. Subsurface deposits in the valley include as much as 140 m of soft, fine-grained sediments that probably are lacustrine (Dearborn, 1977; Johnson, 1979; Deeter and George, 1982; Munter, 1984); deposits that thick may represent more than those of a relatively short-lived lake of Elmendorf age. The breakout of the Elmendorf-age Lake Eagle played an important part in subsequent erosion and deposition on the Anchorage lowland to the southwest as discussed later.

All four of the moraine systems are similar in degree of preservation at similar positions on the slope of the Chugach Mountain front. In the few places where we have observed good exposures of the deposits of these moraines, they all appear similarly unoxidized and uncompacted and are probably close in age. The age of the Bootlegger Cove Formation, or at least of the shell horizon in its upper part, is about 14,000 uncalibrated

radiocarbon years (as discussed below). Since the youngest moraine of this series, the Elmendorf Moraine, has to be somewhat younger than that, we infer that the entire series of moraines discussed here records the retreat of glacier ice from a late Pleistocene maximum advance of about 22,000 to 17,000 yr BP (Mann and Hamilton, 1995), perhaps with minor to considerable readvance. We have no direct evidence, however, that this is the case, and stratigraphic evidence at the Potter Hill site (discussed below) suggests that the sequence of events represented by these moraines could have occupied a longer period of time within the late Pleistocene.

The chronology just outlined differs substantially from that proposed by Karlstrom (1957, 1964) and followed, albeit with varying differences in correlation, by most other workers and compilers (Miller and Dobrovolny, 1959; Cederstrom et al., 1964; Reger and Updike, 1983, 1989; Hamilton, 1986). Following Schmoll and Yehle (1986) and Yehle et al. (1990), we discuss briefly here reasons for our differing with Karlstrom's interpretations.

In a cut along Arctic Valley Road (site AV, Fig. 3-ne), Karlstrom observed oxidized diamicton and gravel probably within moraines mapped by us as Fort Richardson or Dishno Pond moraines. These oxidized deposits served as the basis for his correlation of lateral moraines in this vicinity, and all of the well-developed lateral moraines along the Chugach Mountain front, with the oxidized Eklutna deposits discussed above. He thus considered all of these to be Eklutna moraines, and thought they were of Illinoian (middle Pleistocene) age. His exposure has not been relocated, and most of the deposits we have seen here are generally not oxidized. The exposure Karlstrom observed was probably of localized oxidation within younger deposits; alternatively, he might have seen deposits that are older than the dominant landforms preserved along this hillside. In either case, our observations that most materials in these moraines are only slightly oxidized indicate a younger age for them.

Near where Dishno Pond moraines feather out, Karlstrom recognized moraines that he thought were modified by glacial Lake Cook. He termed them Knik moraines on the basis of a correlation with diamicton that crops out underlying the Bootlegger Cove Formation along Knik Arm, and considered that the deposits and moraines were of early Wisconsin age (pre-Wisconsin in his terminology). Correlation of these Dishno Pond moraines with the "Knik diamicton" may be reasonable, but because other correlations are possible we have preferred to name moraines separately. For example, Karlstrom correlated the main body of Knik diamicton south of Goose Bay with thin diamicton that underlies the middle Pleistocene Goose Bay peat bed northeast of Goose Bay. We are skeptical of Karlstrom's correlation because our observations south of Goose Bay suggest that the Knik diamicton there is close in age to the

Bootleger Cove Formation and therefore of late Pleistocene age as well. On the assumption that Karlstrom's trans-Goose Bay correlation is correct, however, Reger et al. (1996) conclude that the Knik diamicton (south of Goose Bay) is also middle Pleistocene in age. This might be so, but then we would strongly doubt the correlation between the Knik diamicton and the Dishno Pond (formerly Knik) moraines.

The Elmendorf Moraine was correlated by Karlstrom with the suites of Naptowne moraines at Tustumena and Skilak Lakes on the Kenai lowland (represented in Fig. 1 by two arrows). These moraines served as type landforms and deposits of the Naptowne glaciation, thought to be equivalent to the Wisconsin glaciation of mid-continental North America. Although correlating the boldest moraine forms on the Kenai lowland with a similar feature on the Anchorage lowland was not unreasonable, several factors argue against the correlation. The Naptowne moraines include four separately named moraine ridges distributed over a distance of more than 20 km, whereas the Elmendorf Moraine is mainly a single albeit comparably bold ridge. Neither mapping by Miller and Dobrovolny (1959) nor larger-scaled mapping by Yehle et al. (1990) recognize any of the three end moraines upvalley from the Elmendorf Moraine that Karlstrom (1965, Fig. 9–47) mapped there at small scale and thought equivalent to the three inner moraines of the Naptowne complex. The 12,000–14,000 yr age of the Elmendorf Moraine suggests a correlation with only the youngest of the Naptowne moraines, especially since the Skilak (next older) moraine probably has an estimated age somewhat older than 14,500 yr (Rymer and Sims, 1982).

We conclude that the four Naptowne moraines of the Kenai lowland more likely correlate at least in a general way with the four late Pleistocene moraines of the Anchorage lowland we have discussed. Except on the basis of sequence, however, there is little evidence to support specific correlations of other than the youngest moraines. It might not be unreasonable to regard the late Pleistocene moraines of the Anchorage area as representing the Naptowne glaciation, as do Reger et al. (1996); we have not done so, however, in part because deposits of the Naptowne glaciation on the Anchorage lowland would then include respectively the type deposit and principal moraines of the preceding Knik and Eklutna glaciations of Karlstrom's chronology. Instead, we believe it preferable to restrict the term Naptowne to moraines and deposits on the Kenai lowland. In any case, it is not appropriate to use the term Naptowne for the Elmendorf Moraine and its related deposits.

2.4. Late Pleistocene moraines of Turnagain Arm valley

Potter Creek lateral moraines emanate from Turnagain Arm valley and occupy the area between Rabbit

Creek and Potter Creek just outslope from the Chugach Mountain front. They lie downslope from and truncate the Rabbit Creek and uppermost Fort Richardson moraines, but not the middle to lower Fort Richardson moraines. We infer that when the lower Fort Richardson glacier retreated far enough that it no longer reached Turnagain Arm valley, the Potter Creek glacier was able to spread onto the margin of the lowland. The Potter Creek moraines, however, feather out in this vicinity. The Potter Creek glacier apparently terminated in Cook Inlet, and it is unlikely that a closed end moraine ever developed. Instead, both ice front and water level fluctuated, the latter as high as a present-day altitude of about 140 m, below which apparent morainal landforms are quite subdued. Both the Potter Creek and Dishno Pond moraines are younger than Fort Richardson moraines, but they might not be precisely coeval; the Dishno Pond moraines could be somewhat younger than some of the Potter Creek moraines.

The crest of the Potter Creek moraines may be observed along Golden View Drive; streets to the west descend across several lateral-moraine ridges and intervening channels to near sea level (site GV, Fig. 3-se). No single site in this general vicinity, however, provides definitive exposures of these or the Fort Richardson and Rabbit Creek moraines, but good exposures appear as roads are cut through lateral-moraine ridges or house sites are excavated. One such site observed by Dobrovolny and Schmoll in the late 1960s revealed highly oxidized diamicton and gravel, a characteristic that elsewhere convinced Karlstrom (1964) that these were Eklutna moraines. Most of the other exposures observed nearby by Dobrovolny and Schmoll, however, were of gray, unoxidized, and not very compact diamicton similar to material in moraines as young as the Elmendorf. Dobrovolny and Schmoll thought that the oxidation was only of local occurrence and probably not an indicator of great antiquity.

The Bird Creek lateral moraine forms a narrow, relatively low ridge that extends along the north side of Turnagain Arm in the vicinity of Bird Creek (BM, Fig. 2A-se). It continues westward down the arm as far as Indian Creek (about 3 km) between the highway and the railroad. Although poor, exposures along the railroad indicate that the till is interbedded with silt and underlain by stony silt and clay that contain fragmentary mollusk shells not dated here. Shells at Milepost 100 a few kilometers to the east, however, are of the same age as the Bootlegger Cove shells (as discussed below) and the stratigraphic relationship is similar to that of the Elmendorf Moraine overlying shell-bearing Bootlegger Cove Formation along Knik Arm. Thus, a reasonable correlation can be made between this moraine and the Elmendorf Moraine. This is one of the few radiocarbon-dated moraine correlations in the entire region, and it further serves to indicate that relying only on moraine

morphology for correlation can be misleading, as this low moraine bears little resemblance to the massive Elmendorf Moraine.

The moraine forms a low reëntrant less than 1 km up the valley of Bird Creek which may well have been the site of a glacier-dammed lake. Exposures upvalley from the moraine reveal mostly gravel and diamicton; this is not atypical of ice-margin lakes receiving much coarse debris, as can be observed at some present-day ice-margin lakes. A moraine front across Turnagain Arm west of Hope valley on the south side of the arm (farther west than we show in Fig. 2A) is shown on maps by Kachadoorian et al. (1977) and Reger et al. (1996). We conclude that the ice front probably fluctuated over a distance of about 5 km during the thousand years or so that the glaciers terminated in this part of Turnagain Arm.

3. Late Pleistocene Glacioestuarine deposits

Many of the deposits within the southern part of the Anchorage lowland are related to ancestral Cook Inlet. Among these are (1) the glacioestuarine Bootlegger Cove Formation, (2) the younger Mule Creek deposits, and (3) a number of separately named deposits interpreted as transitional between inlet and glacial deposits; these include mainly Tudor Road deposits on the east side and Fire Island deposits on the west side of the Bootlegger Cove depocenter. Although our interpretations differ from those of Karlstrom in many ways, based largely on new data, we acknowledge the importance of his concept of the major role played by a large body of water in Cook Inlet basin during glacial times, a concept that took several years to evolve (T.N.V. Karlstrom, oral communication, 1957).

3.1. Bootlegger Cove Formation

The only clearly glacioestuarine deposit within the Anchorage lowland is the Bootlegger Cove Formation. It consists of silty clay and clayey silt with minor interbedded silt, fine sand, fine to medium sand, and thin beds of diamicton; scattered pebbles and cobbles are present in widely varying concentrations. Some of the silty clay beds include zones of sensitive clay thought responsible for massive landslides that have occurred during large-magnitude earthquakes, as discussed later. The formation originally was named Bootlegger Cove Clay (Miller and Dobrovolny, 1959) after the small embayment at the mouth of Fish Creek (just east of site QLTH, Fig. 3-sw); most of the exposures lie along the shore west of the creek. In the early 1950's the only access to this area was by Clay Products Road, the west end of which was a poor corduroy road across a peat-covered surface, and we have retained the name Clay Products Road (Schmoll et al., 1972) for the site of the best present-day exposures

(CPR, Fig. 3-sw). The Bootlegger Cove Clay was subsequently redesignated as Bootlegger Cove Formation (Updike et al., 1982) because clay is only one of the textures characteristic of the unit.

Mollusk shells of brackish-marine affinity were found along the beach in the vicinity of the Bootlegger Cove exposures, but because they could not be determined to have weathered out of the formation Miller and Dobrovolny (1959) considered it to be glaciolacustrine in origin. Later, both Cederstrom et al. (1964) and Karlstrom (1964) reported shells in situ within the formation. Because accuracy of radiocarbon dates of shell material was questioned in the 1950s, and because Karlstrom thought the shells should date beyond the range of the radiocarbon method, he decided to use the still experimental ionium-thorium age-dating method. He received an age of 33,000 to 48,000 yr for a sample of shell material collected by F.W. Trainer (Sackett, 1958). In Karlstrom's interpretation only the shell horizon was regarded as estuarine ["marine"] and deposited during an interglacial interval of high sea level that he postulated to have occurred at about 45,000 yr BP. The remainder of the formation was considered evidence for glacial Lake Cook, a regional lake in Cook Inlet basin postulated to have beeen dammed by coastal glaciers during both the earlier Knik glaciation and the later Naptowne glaciation.

Micropaleontological investigations by Schmidt (1963) of a channel sample through the exposed stratigraphic thickness of the Bootlegger Cove Formation, undertaken subsequent to the completion of Karlstrom's report, determined that both ostracods and foraminifera are present in all but the uppermost and lowermost parts of the section and indicate a brackish marine environment. The microfossils were most abundant in the shell horizon. This work established that most of the formation was of marine rather than lacustrine origin. In conjunction with investigations following the great 1964 Alaska earthquake, Smith (1964) found a similar microfauna in borehole samples that extended stratigraphically lower and covered a geographically larger area. Table 1 lists the species reported by Schmidt (1963).

The Turnagain Heights landslide caused by the 1964 earthquake (discussed later) destroyed most of Clay Products Road and about two-thirds of the nearby bluffs with their exposures of the Bootlegger Cove Formation. Following the earthquake, however, the remaining bluffs became better exposed, possibly because of renewed erosion following earthquake-induced tectonic subsidence of about 0.6 m (Plafker, 1965), and better roads provided good access to the nearby bluffs. More shells were found in situ, not in a well-defined bed, but in a zone about a meter thick that descends from about 11.6 m above the high water line at the Clay Products Road site to about 7.6 m nearly 500 m to the east. It is not clear whether this gradient of about one percent has any depositional or tectonic significance. New collections from this zone yielded a substantially younger age for the mollusk shells using both radiocarbon (about 14,000 yr, uncalibrated) and uranium-series (about 15,000 yr) methods (Schmoll et al., 1972). The new dates mandated a revision of the upper Cook Inlet glacial chronology that places the Bootlegger Cove Formation in a late glacial rather than an interglacial or interstadial context. The shells on which the new dates were obtained include the species, shown in Table 2, identified by W.O. Addicott, U.S. Geological Survey.

Overlying the Bootlegger Cove Formation in the bluffs are thin beds of well-sorted medium to fine sand termed Hood Lake sand (Yehle et al., 1991). Although less than

Table 2

Gastropods	Pelecypods
Buccinum cf. *B. Physematum* (Dall)	*Clinocardium ciliatum* (Fabricius)
Cryptonatica clausea (Broderip and Sowerby)	*Clincardium* sp.
Cryptonatica sp.	*Hiatella arctica* (Linné)
Trichotropis borealis (Broderip and Sowerby)	*Macoma* cf. *M. calcarea* (Gmelin)
Barnacle	*Mya truncata* (Linné)
Balanus sp.	

Table 1

Ostracods	Foraminifera
Normanicythere sp.	*Quinqueloculina seminula* (Linné)
N. cf. *N. macropora* (Bosquet)	*Guttulina lactea* (Walker and Jacob)
N. leioderma (Norman)	*G.* sp.
Loxoconcha sp.	*Globulina* cf. *G. glacialis* (Cushman and Ozawa)
Trachyleberis cf. *T. rastromarginata* (Brady)	*Elphidium incertum* (Williamson)
Palmanella sp.	*E. incertum* (Williamson) var. *clavatum* (Cushman)
Cytheropteron sp.	*E.* cf. *E. bartletti* (Cushman)
	Elphidiella groenlandica (Cushman)
	Protelphidium orbiculare (Brady)

a meter is exposed in the bluffs sand, a meter or more thick is spread to the south as the mapped surface deposit; it is named from Hood Lake, a major float-plane base just south of the Clay Products Road site. This sand deposit is mapped separately from the Bootlegger Cove by Schmoll and Dobrovolny (1972, and unpublished mapping) but is included within that formation by Updike and Ulery (1986). As it seems to represent a shoaling facies of Bootlegger Cove deposition, we believe that designating the Hood Lake sand a member of that formation would be appropriate.

In the area now occupied largely by International Airport (Fig. 3-sw), but extending somewhat to the south as well, are deposits that Miller and Dobrovolny (1959) called "silt near International Airport". They regarded the deposits as mainly eolian although they recognized that finer and coarser beds within them might have other origins. In unpublished mapping by Schmoll and Dobrovolny the unit is termed International Airport silt and regarded as estuarine because interbedding of clay and silt beds like those of the Bootlegger Cove Formation has since been observed. It is possible, however, that some silt in the upper part of the deposits might be eolian. Updike and Ulery (1986) consider these deposits to be a facies of the Bootlegger Cove Formation.

Although the shell horizon within the Bootlegger Cove Formation was first noted only along Knik Arm near Anchorage, subsequent investigations located similar shells of about the same age both in Turnagain Arm valley and on the west side of Cook Inlet along the Beluga River (Fig. 1) (Schmoll and Yehle, 1986). Somewhat older shells, almost entirely barnacles about 16,000 yr old (uncalibrated radiocarbon age) have been found in silt and clay deposits on the Kenai lowland to the south (Schmoll and Yehle, 1986; Reger et al., 1995). These findings clearly establish a wider geographical distribution of the formation than was originally known.

A rare occurrence of badly crushed *Inoceramus* specimens of Late Cretaceous (Maastrichtian) age from the Valdez Group (Jones and Clark, 1973) is found at a site along the highway in Turnagain Arm valley designated as Milepost 100 (MP, Fig. 2A-se). On a lazy Sunday afternoon in 1970, Dobrovolny and Schmoll intended to look for one of these fossils. Instead, amid bedrock rubble they found mollusk shell fragments similar to those collected from the shell horizon of the Bootlegger Cove Formation in the Anchorage lowland. A radiocarbon age of 13,900 ± 400 yr BP (uncalibrated; Schmoll and Yehle, 1986) from this material established for the first time that the Bootlegger Cove shell horizon extended beyond the Anchorage lowland and at least this far up Turnagain Arm. With widening of the highway, the exposure has improved somewhat, and some of the stony silt-clay that contained the shells has become exposed, but most of the material visible in the cut is colluvium and bedrock rubble.

Other evidence indicates that the Bootlegger Cove estuary extended even farther up Turnagain Arm. Just east of Girdwood (Fig. 2B-sw) microfossils similar to those in the Bootlegger Cove Formation along Knik Arm were found in an exposure of silty to fine sandy diamicton. Silt and clay deposits in the lowest part of a 300-m deep water well at Portage at the head of the arm have an extrapolated age of about 14,000 yr (Bartsch-Winkler et al., 1983); although this age is only an approximation, it suggests that these deposits are Bootlegger Cove Formation. Coarse material that is indicated somewhat higher in the well log could represent till deposited by the Bird Creek glacier. It is thus thought that the Bootlegger Cove estuary extended not only this far inland but perhaps up the valleys of Twentymile and Placer Rivers as well.

3.2. Mule Creek deposits

Mule Creek deposits (Yehle et al., 1991) are restricted to altitudes lower than the Bootlegger Cove Formation but above the level of modern Cook Inlet. They are found on the west side of Knik Arm within the area of the Elmendorf Moraine and presumably overlying it (Figs. 2A-nw, 3-nw). They are also present south of Point Woronzof (Fig. 3-sw) where the bluff splits into an upper and a lower bluff and a terrace-like surface between them slopes gently to the south. This surface is lower than the level of the Mountain View fan and the Bootlegger Cove Formation; it may have formed after relative sea level dropped considerably from Bootlegger Cove levels. Gravel present at depth in the terrace may represent an early post-Elmendorf drainage channel; further development of that channel resulted in separation of Fire Island from the mainland. The surface of the terrace, however, is underlain by a few meters of clayey silt, implying perhaps that sea level rose briefly and deposited the silt after the subaerial interval represented by the underlying gravel. We correlate the silt and clay here with Mule Creek deposits on the west side of Knik Arm because deposits in both areas seem to represent a short-lived estuary at this intermediate level.

3.3. Deposits along the eastern margin of the Bootlegger Cove Formation

Deposits east of the glacioestuarine Bootlegger Cove Formation were originally mapped (Miller and Dobrovolny, 1959) as pitted outwash or ground moraine, implying a subaerial environment of deposition and a relatively low water level for the Bootlegger Cove estuary. However, generalized mapping by Schmoll and Dobrovolny (1972) suggested that these deposits instead might be glacioestuarine ["marine"], and that the Bootlegger Cove water reached somewhat higher levels. More recently, Yehle et al. (1992) and Schmoll et al. (1996)

regarded these deposits as of a marginal glacioestuarine, near-ice environment and subdivided them into several map units; some of these units include glacial deposits thought to have been reworked or their landforms reshaped in a subaqueous environment, whereas the deposits of other units are more clearly of a primary glacioestuarine origin. Each of the subdivided deposits differs somewhat in landform, lithology, relationship with moraines, and altitudinal range. Landforms seem to grade through a continuum that extends from drumlinoid and kame-like hills to apron-like areas surrounding the hills and finally to lower-lying plains. Lithologies include variously interbedded diamicton, gravel, sand, and silt. No shorelines have been observed, but the upper limits of these deposits provide clues to possible shoreline positions. Most of the separately mapped deposits are here collectively termed Tudor Road deposits; Birch Road deposits remain separate because they might be glaciolacustrine, as discussed below.

The principal characteristics of the Birch Road and Tudor Road deposits are given in Table 3. They are listed approximately from highest to lowest in altitude, and in a more general way from oldest to youngest (some deposits may be coeval with others above and below them on the list).

In the vicinity of Birch Road (site BR, Fig. 3-se), the hummocky Fort Richardson kame field fades into a smoothly sloping apron-like area that ranges in altitude from about 140 to 220 m. This area is underlain by silt and fine sand termed the Birch Road deposits, named from good exposures along Birch Road when the road was constructed; in recent years the deposits have become increasingly concealed as roadcuts slumped and became covered with vegetation. The deposits probably formed in water that was in contact with the Fort Richardson glacier. In the higher-lying parts of the area, gravel and diamicton are prevalent, probably in part Fort Richardson-age material modified by the action of water. Because these deposits lie higher than other gla-

cioestuarine deposits, Yehle et al. (1992) considered that they may have been deposited in a lake blocked by lower-level Fort Richardson ice on the northwest side and by Potter Creek ice on the southwest side. Another possibility is that they are the product of a high-level regional lake, as proposed by Karlstrom (1964), who regarded such deposits as evidence for a 750-ft (230 m) level of glacial Lake Cook. No microfossils that might aid in determining their origin have been found in these deposits.

Abbott Road deposits, named from their occurrence along part of that road (site AR, Fig. 3-se), generally lie subjacent to Fort Richardson moraines and kames. In shallow exposures elsewhere, the deposits consist mainly of poorly bedded diamicton, stony silt, and silt; here they are characterized additionally by rubbly material that may have been the result of a landslide (earthquake-induced?) emplaced on a Fort Richardson glacier, transported to about this position, and subsequently reworked by inlet water (Schmoll et al., 1996, following Dobrovolny and Schmoll, written commun., 1970).

In the vicinity of Huffman Road and Bragaw Street (site HR, Fig. 3-se), the ground surface descends below about 140 m, slopes are more gentle, and four types of deposits converge: (1) Birch Road deposits to the northeast; (2) Abbott Road deposits, to the north; (3) Huffman Road deposits to the southwest; and (4) the lowest-lying vestige of a Potter Creek lateral moraine to the southeast. Here, Huffman Road deposits appear to lap upon the Potter Creek moraine, but just to the south, the Huffman Road deposits are interpreted as grading into DeArmoun Road deposits. The latter occupy elongate hills similar to (but more subdued than) the Potter Creek lateral moraines and are thought to be inlet-modified morainal landforms.

Deposits interpreted as equivalent to those mapped at the surface as Huffman Road are exposed at a site called Potter Hill (PH, Fig. 3-sw). The name refers to the grade on the Alaska Railroad as it ascends from near sea level

Table 3
Birch Road and Tudor Road Deposits

Name	Landform	Principal lithologies	Associated moraines	Altitude range (m)
Birch Road	Apron	Fine sand and silt; diamicton	Fort Richardson	140–220
Tudor Road:				
Abbott Road	Apron	Diamicton, stony silt, silt, rubble locally	Fort Richardson	110–140
DeArmoun Road	Hill	Diamicton, stony silt, silt	Potter Creek	110–140
Huffman Road	Apron	Silt, stony silt, fine sand, diamicton	Potter Creek	< 110
Russian Jack	Hill	Diamicton, minor interbedded silt	Fort Richardson	< 110
Boniface Parkway	Hill	Gravel and sand	Fort Richardson	< 110
Muldoon Road	Apron	Silt, stony silt, fine sand, diamicton	Dishno Pond	< 110
Early View	Plain	Silt and silty clay, sand, minor gravel	Dishno Pond	< 95
O'Malley Road	Hill	Gravel and sand	—	< 75
Geneva Woods	Hill	Silt and silty clay, sand, minor diamicton	—	50–75
Winchester Street	Plain	Sand; gravel locally	—	50–75

up to the level of the Anchorage lowland. Prior to the 1964 earthquake the railroad track was farther seaward than it is now; the track was built partly on fill that was probably derived from the prior cut and emplaced on intertidal deposits. During that earthquake, a complex landslide about 1300 m long developed along the bluff and destroyed about 475 m of track. It is not clear how much of the destruction can be attributed to failure of natural ground and how much to failure of the fill, but evidently both types occurred (McCulloch and Bonilla, 1970).

Stratigraphy along the new railroad cut was well exposed in 1965 (Fig. 4) and engineering stations surveyed during the reconstruction work were still marked on the ground. Subsequent erosion of the cut has built talus cones that now cover about the lower half of the geologic section exposed in 1965. However, it is now easier to reach the upper part of the cut and differential erosion has made bedding more visible. The stratigraphic section includes three thin, laterally discontinuous beds of compressed peat, wood, and other organic matter. All three organic horizons have radiocarbon ages greater than 40,000 yr. These horizons mark subaerial conditions at this site, and, by inference, throughout much of the Anchorage lowland. The highest organic horizon has the greatest lateral extent and divides the Potter Hill section into upper and lower parts.

The upper part of the section consists of interbedded diamicton, silt, sand, and gravel. Between engineering stations 2875 and 2900, it is dominated by diamicton that has a matrix of fine sand and silt and contains many distinct interbeds of fine sand and silt. The diamicton becomes finer grained to the northwest, where it grades laterally into bedded silt and fine sand; to the southwest,

diamicton coarsens and grades into medium sand and gravel. A sparse estuarine microfauna and a tiny shell fragment were found in diamicton in this upper part of the section. The organic horizon at the base of the upper part contains reworked waterlaid material and is overlain by beds of fine sand. The upper part is interpreted to indicate the following concommitant conditions: (1) a shallow embayment or estuary, similar to the present-day Cook Inlet; (2) a nearby glacier from Turnagain Arm; (3) a basinward lessening of the influence of that glacier; and (4) an ancestral Rabbit Creek providing coarser and better sorted sand and gravel at the southeast end of the exposure. The thick, well sorted gravel at the top of the southeast end of the section is a later subaerial alluvial fan of Rabbit Creek.

The exposures at Potter Hill illustrate the difficulty in correlating deposits in stratigraphic sections with those found in moraine and other landforms. All of the deposits of the upper part of the section might be considered Huffman Road deposits because there is no apparent discontinuity in the stratigraphy above the highest organic zone. If so, Huffman Road deposits extend back in time to more than 40,000 yr BP. Geomorphic interpretations presented above, however, suggest that they be substantially younger than that, and that glacier ice probably advanced into and receded from this area about 20,000 yr ago. Either the geomorphically derived age inferences are incorrect, or the Huffman Road deposits should be limited to the uppermost part of the stratigraphic section here. In the latter case, some of the discontinuous diamicton beds in the upper part of the section may represent Potter Creek, Fort Richardson, Rabbit Creek, or even older till and (or) waterlaid equivalents; this does not seem reasonable, however, because of
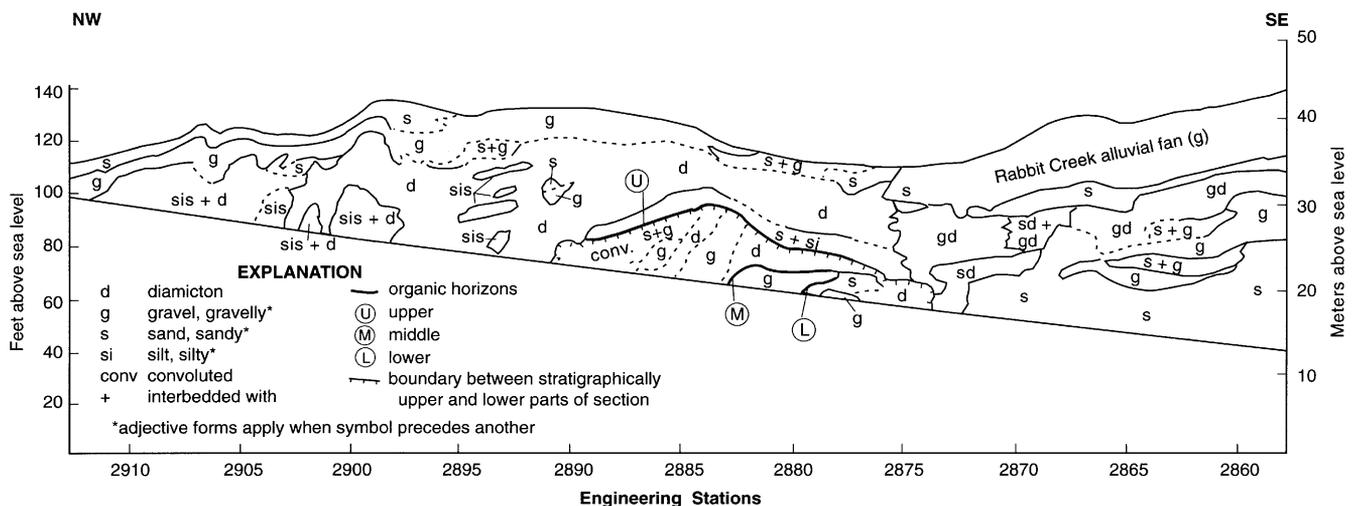

Fig. 4. Generalized stratigraphy in the Potter Hill railroad cut. Derived from a larger-scaled portrayal based on detailed measurements made in 1965 by Dobrovolny and Schmoll. Horizontal control provided by engineering stations measured in hundred-foot (30.5 m) units that had an origin to the southeast. (From Bartsch-Winkler and Schmoll, 1984b).

the overall interbedded nature of these deposits. Perhaps the glacial advances so prominently represented by lateral moraines were relatively short lived, and therefore did not leave a substantially thick or continuous record in the stratigraphy; what is preserved stratigraphically might be only some small part of the geomorphic record.

The lower part of the Potter Hill stratigraphic section consists mainly of gravel, sand, and diamicton characterized by highly contorted bedding; individual beds are not easily traceable. The two laterally discontinuous organic horizons in this lower part may be the same bed. Subglacial deposition during one or two episodes of glaciation may be represented by the lower section; the deposits could correlate with one or more of the older, more extensive glaciers evidenced by remnant moraines high along the Chugach Mountain front.

Included in the Tudor Road deposits are diamicton in small drumlin-like hills and gravel and sand in kame-like hills some of which trend northeast-southwest in alignment. Many hills are surrounded by a belt of lower, smoother Muldoon Road deposits (discussed below). The deposits in the drumlin-like hills are called Russian Jack deposits, named from Russian Jack Springs Park and shown in the subsurface cross section of Schmoll and Barnwell (1984). The best exposure, however, is at a site about 750 m east of the park (RJ, Fig. 3-ne), where about 16 m of mainly fine-sandy diamicton is locally interbedded with silt and fine sand in thin beds. In some hills the upper part of the diamicton may have been reworked by glacioestuarine water. Many hills seem to have been partly eroded and reshaped, perhaps with the eroded material deposited in the surrounding belt of Muldoon Road deposits. Whether the hills owe their present shape to drumlinization or later subaqueous modification is not clear. Some hills, however, definitely have been eroded by streams flowing in the adjacent Nunaka Valley and Mountain View channels; these hills are narrower and more elongate than most of the others.

The kame-like hills are more sharply defined than the Russian Jack hills, and more variable in shape and size. They are composed of gravel and sand named Boniface Parkway deposits from a small gravel pit along that street about 1 km southwest of the Russian Jack site (Schmoll et al., 1996). The Boniface Parkway site has since been urbanized, but when the pit was active, Muldoon Road silt was observed there lapping onto the gravel. O'Malley Road deposits (Schmoll et al., 1996) are similar to the Boniface Parkway deposits but occur at lower altitudes and generally are not aligned. They partly adjoin flatter-lying Klatt Road delta deposits (near site LK, Fig. 3-sw). Many of the hills have been destroyed by gravel-pit operations.

Muldoon Road deposits (Schmoll and Barnwell, 1984) consist chiefly of fine sand and silt with minor interbedded diamicton. They were named from exposures in a small hill where that road now joins Tudor Road (site

MR, Fig. 3-se); subsequent highway reconstruction demolished the hill. The Muldoon Road deposits occupy a broad northeast-trending belt of smoother, lower-lying terrain adjacent to (and in places completely surrounding) hills of Russian Jack deposits. Elsewhere, they occur in low hills surrounded by broad glacioalluvial channels; these hills are probably remnants of previously more extensive Muldoon Road deposits, much of which have been eroded in the process of channel formation (Schmoll et al., 1996).

Winchester Street (site WS, Fig. 3-se) gives its name to poorly exposed sand deposits that lie between about 50 and 75 m, lower than Muldoon Road deposits but higher than Hood Lake sand to the west (Schmoll et al., 1996). They may represent a marginal facies of the Bootlegger Cove Formation; the lower part of that formation may underlie the Winchester Street deposits.

Early View deposits (Schmoll et al., 1996) occupy channel-like areas in the vicinity of South Fork Chester Creek (site EV, Fig. 3-se). They consist predominantly of silt and clay but include coarser deposits marginally. It is not clear whether they accumulated in an arm of the Bootlegger Cove inlet, or whether they are younger and possibly were deposited in a local lake. No microfossils have yet been found in these deposits.

Geneva Woods deposits occur in low hills in an area near Campbell Creek (Fig. 3-sw). They are shown in unpublished mapping by Schmoll and Dobrovolny and seem to be part of a landform and lithology continuum extending from higher to lower altitudes and from Russian Jack (well-formed hills of diamicton) through Muldoon Road (lower-lying but slightly hummocky areas of silt and diamicton) and Geneva Woods (irregular hills of sand, silt, and some interbedded diamicton) to broad low hills of sand farther west. Although the Russian Jack hills probably represent buried and (or) modified glacial deposits, it is not clear whether the Geneva Woods hills owe their origin to similar but more deeply buried and modified glacial deposits, or whether the deposits represent facies of inlet deposition unrelated to glacial landforms. In either case these deposits could be considered facies of the Bootlegger Cove Formation. Exposures of definitive stratigraphic relationships between these deposits and the more characteristic silt-clay facies of the Bootlegger Cove have not been available, however, and Schmoll and Dobrovolny (1972) map them separately. Here we have provisionally included them with the Tudor Road deposits.

3.4. Fire Island deposits

The principal deposits that border the Bootlegger Cove Formation on the west side of the Anchorage lowland are termed Fire Island deposits in unpublished mapping by Schmoll and Dobrovolny. They are regarded as formerly coextensive with deposits on Fire

Island (Fig. 2A-sw) which lies about 6 km west of Point Campbell, the broad western tip of the Anchorage mainland. The island rises to altitudes of about 100 m, with a surface of sharply defined low hills and intervening flat-floored channels. Bluffs as much as 60 m high border the island except at the southwestern end where there is a low platform of beaches and lagoons and a developing bar.

Fire Island deposits may be part of a glaciodeltaic complex deposited subaqueously near the glacier-inlet interface (Schmoll and Gardner, 1982; Schmoll et al., 1984) and contemporaneous with some of the Tudor Road deposits discussed above. A preliminary geologic map (Schmoll et al., 1981) subdivides the deposits on the island mainly on the basis of surface morphology. They were previously mapped as end moraine by Cederstrom et al. (1964), who regarded the deposits as possibly correlative with the Elmendorf Moraine, whereas Karlstrom (1964) considered them an extension of the Potter Creek moraine. The Fire Island deposits on the mainland were regarded as delta deposits closely related to deposits on the island by Miller and Dobrovolny (1959), whereas they are included within the Bootlegger Cove Formation by Updike and Ulery (1986).

Fire Island deposits are well exposed nearly continuously for about 6 km in the steep bluffs of the southeast side of the island. A diagrammatic portrayal of the full length of the exposures is presented in Schmoll et al. (1984), and forms the basis for the following description. The deposits include beds of diamicton, sand, and gravel, with only minor beds of material finer than fine sand. The matrix of the diamicton in most places, however, is silty fine sand. These varied textures are complexly interbedded, and few beds, except for some diamictons, are very continuous. Beds thicken and thin within short distances, in places only a few tens of meters; this characteristic is especially true of some diamicton beds.

The Fire Island diamicton beds have been subdivided into five types: (1) massive, thick diamicton units that grade laterally over their full thickness into better sorted materials; (2) locally thick masses of diamicton that thin drastically in both directions; (3) lenses of diamicton less than a meter thick and less than 50 m long occurring mainly within gravel beds; (4) relatively discontinuous diamicton beds that grade into and out of gravel, sand, and silt, and appear mainly at the top of the bluff (where they have not been observed closely); and (5) a 3- to 20-m thick and 3-km long diamicton unit that dominates the northern part of the bluff and is similar to type 4 except in dimension.

The better sorted materials, mainly gravel and sand, are more difficult to classify because of their lateral gradation over short distances. Nevertheless, seven distinct deposit types can be recognized throughout the exposures: (a) well sorted silty fine sand that occurs in evenly bedded, relatively thin units that can be traced as marker beds over several hundred meters; (b) similar material that occurs in contorted lenses interbedded in coarser sand and gravel and that may constitute a horizon that originally consisted of a continuous noncontorted bed; (c) interbedded and commonly contorted sand and gravel of widely varying clast sizes; (d) relatively continuous even-bedded units of sand; (e) similar units of gravel; (f) pebble gravel with a fine sandy matrix; and (g) "dirty gravel", less well sorted than the pebble gravel and with coarser phenoclasts and a high proportion of silty matrix; this material could almost be considered a diamicton, but has a higher phenoclast/matrix ratio than most diamictons and is generally clast supported.

Although hilly, the ground surface does not have the typically hummocky, pond-dotted character of most relatively young moraines; the hills are smooth, well-rounded, and separated by through-going channels. The deposits exposed in the bluffs could be interpreted as ground moraine, with intervening kames and other products of stagnant ice and associated running water. We are not the first, however, to consider that Cook Inlet (or, as earlier thought, glacial Lake Cook) might be involved in the process of deposition of the Fire Island deposits; Cederstrom et al. (1964) describe the diamicton as "water-laid till". We concur and view the interbedded nature of the well sorted deposits as the key to the depositional environment. The thinner diamicton lenses are a more poorly sorted facies of a bedded, water-deposited sequence. We believe that the thicker diamicton beds are of similar origin but larger in scale.

We believe that the Fire Island deposits formed in a glaciodeltaic environment. In our model, a glacier was located to the north, and it rapidly dumped large amounts of debris into standing water, resulting in the build-up of a body of unstable material of all size ranges from clay to boulders. Much of the finest component, however, did not settle out, whereas locally well sorted coarser materials were deposited here. Sediment probably slid periodically (possibly, but not necessarily, triggered by earthquakes), and formed subaqueous flows that in part were deposited as beds of diamicton.

There seems to be no obvious glacier-terminus position marked by a moraine to serve as source for the deposits. We ascribe the source to a glacier that came down the Susitna valley from the north, joined by a glacier in the Knik-Matanuska valley from the northeast. On the basis of the position of feathering out of lateral moraines on both sides of the basin, these were probably Fort Richardson glaciers. The Rabbit Creek glacier extended farther south than Fire Island, possibly to near Point Possession on the northern tip of the Kenai lowland (Fig. 1) where landforms and stratigraphy appear similar to those of Fire Island.

An alternate interpretation of the Fire Island deposits considered but not favored by Schmoll et al. (1984) is that they are older than any of the moraines and glacier

positions mentioned above and thus not related to them. Deep subsurface investigations in conjunction with oil and gas exploration indicate that Fire Island lies on the trend of anticlines mapped to the northeast and southwest that are developed on subsurface Tertiary rocks (Magoon et al., 1976). If such structures continued to develop throughout the Quaternary, Fire Island deposits might have been uplifted to their present position. The bluffs are lower at both their southwest and northeast ends, giving the impression that the stratigraphy rises up out of the inlet to a high elevation on the island and then descends again. Although these deposits do not appear to have the antiquity of those previously mentioned as exposed at low tide at Point Woronzof, the presence of older deposits there suggests that these deposits, too, might have an age greater than those of landforms at the ground surface.

Sand and gravel beds of the Fire Island deposits are well exposed at Point Woronzof (PW, Fig. 3-sw) where sand dominates the upper part of the bluff and gravel beds are present in the middle. The gravel beds dip gently eastward and thin downdip. The lower part of the bluff contains sand beds that are locally rich in coal fragments; the number of coal phenoclasts varies from year to year as the bluff erodes. There are early reports of in-situ coal beds at Point Campbell (possibly not distinguished from Point Woronzof; Martin, 1906, p. 25); coal is reported as unconfirmed, but is mapped between Ship Creek and Point Woronzof, by Capps (1916), p. 154), and reported, but not mapped there, in Capps (1940), p. 62). No coal beds have been found here by more recent workers and it is likely that detrital coal was mistaken for in situ material. Karlstrom (1964) regarded the sand beds low in the bluff as Eklutna (pre-Wisconsin) deposits on the basis of their yellowish color; Updike and Ulery (1986) map the same beds as slightly to moderately indurated older deposits. Dobrovolny and Schmoll, and Schmoll and Yehle, however, found no discontinuities in the stratigraphy, but only subtle changes in color, texture, and cohesiveness at numerous horizons. The yellowish color of these beds might be caused by ground-water oxidation or by the color of the coal-bearing source beds. Still, the possibility remains that the lower sand beds are significantly older than the overlying gravel and sand.

Stratigraphy in the bluff is exposed almost continuously from Point Woronzof east to the Clay Products Road site. The Bootlegger Cove Formation dominates the bluffs for about the first 400 m west of that site; it includes sand beds as much as 3 m thick. At a small gully, a 50-m stretch of bluffs is largely concealed; to the west, the bluffs rise to about 40 m in height and include only Fire Island deposits, dominantly sand. Not far west of the gully, diamicton has been observed at one place high in the section. In the vicinity of the gully, we have observed some interfingering between silt-clay and sand beds, and attribute the transition from silt-clay to sand and gravel

to lateral gradation; by this interpretation the thick sands within the Bootlegger Cove are thus stringers of Fire Island sand. Because stratigraphic relationships are not entirely clear, it is possible that upper and middle parts of the Bootlegger Cove might lap onto rather than interfinger with the Fire Island.

An alternate interpretation of the Bootlegger Cove-Fire Island transition is that the two units are in fault contact (Shannon and Wilson, Inc., written commun., 1969). This idea is not supported by field observations of Dobrovolny and Schmoll or subsurface investigations of Updike and Ulery (1986); all of these workers interpret their data as indicating interfingering relations. Still, there is some speculative support for the possible presence of a fault here. Normal faulting in the subsurface (relative movement down on the east side) has been interpreted from a seismic refraction line about 5 km southsouthwest of here (discussed later). If the Fire Island deposits lie on a growing anticlinal structure, as suggested above, normal faulting on the east limb of the anticline could occur here as well. Very small steeply dipping faults with the same sense of movement have been observed near the top of the section where the bluff rim rises, but they most likely are evidence not of tectonic activity but of penecontemporaneous slumping, common in such a glaciodeltaic environment of deposition as envisioned for the Fire Island deposits. Further shallow subsurface investigation, perhaps by the Mini-Sosie method, is required to confirm or deny the presence of a fault at this site.

Point Woronzof provided the name of the Woronzofian transgression of Hopkins (1965, 1967) for an interglacial or interstadial marine transgression represented by the Bootlegger Cove Formation as exposed at the Clay Products Road site. No deposits present at the point were pertinent to this concept. With the Bootlegger Cove Formation now clearly a late glacial deposit, the term Woronzofian is no longer useful as originally proposed.

4. Late Pleistocene glacioalluvial and deltaic deposits

Much of the Anchorage lowland is underlain by well sorted gravel and sand in glacioalluvial deposits and their downstream deltaic equivalents. Streams coursed from both the lowland trunk glacier farther northeast and glaciers and glacial lakes in local mountain valleys to the inlet farther southwest. The resulting deposits occur in (1) narrow channels within the areas of lateral and ground moraine, (2) deltas adjacent to high levels of ancestral Cook Inlet, and (3) broader channels and fans formed subsequent to the withdrawal of both glacier ice and Cook Inlet. Altitudes of delta fronts correspond approximately with the upper limits of some Tudor Road deposits. The glacioalluvial and deltaic deposits are subdivided on the basis of source and (or) base level to which they are graded and as shown in Table 4.

Table 4
Glaciofluvial deposits and features

Klatt Road	Channels along the Chugach Mountain front graded to deltas on the east side of the Bootlegger Cove estuary
Spring Hill	Similar to but graded to levels lower than Klatt Road deposits
Sand Lake	Deltas on the west side of the Bootlegger Cove estuary
South Fork and North Fork Campbell Creek	Alluvial-fan and channel deposits of local mountain-valley source
Nunaka Valley	Channel deposits of Eagle River source
Mountain View	Broad alluvial-fan deposits of Eagle River source
Fossil Creek, Roosevelt Road, Gwenn Lake, and Sixmile Lake	Eagle River alluvium graded lower than the Mountain View deposits
Ship Creek	Alluvium of Ship Creek source graded to levels higher than, equal to, and lower than the Mountain View deposits

Klatt Road deposits (Yehle et al., 1992) occur in a series of channels that parallel the grain of the Fort Richardson moraines and were formed before the streams could cut across the newly emerged, higher-level Tudor Road deposits to reach the inlet more directly. They lead to deltas at two or more major levels. Large gravel pits were developed in several of these deltas on the eastern side of the Bootlegger Cove estuary. The gravel in the pit at site UK (Fig. 3-sw) is graded to about the 110-m level of glacioestuarine water, corresponding to part of the Tudor Road deposits as well as to the Bootlegger Cove Formation. When the pit was in active use, good exposures exhibited westward-dipping deltaic gravel beds; more steeply dipping foreset beds could be seen in a few places.

The Spring Hill channel (site SH, Fig. 3-se) is the lowest of several channels that extend between the vicinity of South Fork Campbell Creek and the deltas that border the east side of the Bootlegger Cove Formation. These channels probably carried water from both forks of Campbell Creek and may have received outwash from a glacier farther northeast as well. The 150-m width and 25-m depth of this channel attests to the quantity of water that must have coursed through it. Perhaps large channels such as this one formed when water was temporarily blocked in a tributary valley, only to be released in a surge. Most of the channels were occupied by streams more than once, as terrace remnants along their sides attest, before being abandoned in favor of a different, lower channel. The thickness of the gravel that is presumed to underlie this channel is unknown; in places where there are sizable bogs on the channel floor, gravel may be very thin or lacking. The Spring Hill channel leads to a lower-lying Klatt Road delta (site LK, Fig. 3-sw). Gravel extraction there has ceased, but when the pit was active exposures also suggested that the gravel and sand are of deltaic origin. The deposits lie at about 60 m, and terminate abruptly at the margin of Bootlegger Cove Formation to the west. Much of the gravel in all the Klatt Road deltas probably derived from erosion of the lateral moraines and associated kame fields through which the channels feeding them were cut.

Gravel and sand in deltas on the west side of the Bootlegger Cove estuary are termed Sand Lake deposits by Schmoll and Dobrovolny (unpublished mapping); they are included within the Bootlegger Cove Formation by Updike and Ulery (1986). The extensive Sand Lake gravel pits (site SL, Fig. 3-sw) produced much of the aggregate necessary for construction in the southern part of Anchorage. Gravel and sand seen when the pit was active were well stratified, with gentle to locally more steep (foreset?) dips to the east; these deltas are similar in altitude to the lower deltas on the east side of the Bootlegger Cove depocenter. The source for the Sand Lake deltas is less clearcut, however. The original surface of the deltas does not seem to be an eastward extension of the surface of the Fire Island deposits. Instead, the surface leads upslope into channels cut into Fire Island deposits; the channels appear headless. These relations suggest that the streams leading to the low Sand Lake deltas came from north of the present areas of the Fire Island deposits, after the ice had withdrawn some distance and also after the subaqueous Fire Island deposits were no longer submerged. Some of these streams may have occupied channels on Fire Island. Such a scenario, of course, argues against an alternate idea that the Fire Island deposits were uplifted tectonically.

Following regression of the ancestral inlet from Bootlegger Cove levels to levels near the modern inlet, streams from the Chugach Mountain valleys were able to make their way across the lateral moraines and emerge more directly onto the Anchorage lowland rather than flowing subparallel to the mountain front. These streams had sources in small glaciers farther up the valleys and deposited alluvial fans and related channels in the margin of the lowland especially along the North and South Forks of Campbell Creek (Fig. 3B-ne). These deposits have been subdivided on the basis of several levels to which they appear graded. Some may be coeval with the southwestward advance of a glacier moving down the lowland to form the Elmendorf Moraine. At least one channel is substantially younger, possibly Holocene in age.

As the Elmendorf glacier waned, it was no longer able to provide a dam for glacial Lake Eagle on a steady basis, and the lake drained, perhaps only partially, but probably repeatedly. The damming and drainage of this lake is comparable to that of modern Lake George (Fig. 2B-ne), which was dammed by Knik Glacier but broke out on a yearly basis for much of the middle part of the 20th century and perhaps earlier as well (Stone, 1963). The first streams to flow southwest into the Anchorage lowland from glacial Lake Eagle probably coursed through the broad Nunaka Valley channel, named from the small community of Nunaka Valley (Schmoll et al., 1996), eroding both ground moraine and softer glacioestuarine deposits and depositing substantial gravel and sand. University Lake was developed in an area of large former gravel pits excavated mainly into the Nunaka Valley channel deposits (site NC, Fig. 3-se). As much as 10 m of well-bedded, well-sorted, pebble- to cobble-gravel has been observed in the east wall of the largest pit. In that part of the Nunaka Valley channel in the vicinity of Checkmate Drive (site CD, Fig. 3-se), the ground surface was characterized by numerous large granitic boulders, many containing large plagioclase phenocrysts. During urbanization, most of these boulders were placed decoratively on house lots. Little gravel is present in this part of the channel; instead, it is underlain by finer-grained and more poorly sorted material. Schmoll et al. (1996) speculate that erosion, rather than deposition, was dominant here, and that the boulders represent a lag concentrate from erosion of glacioestuarine Tudor Road and (or) earlier ground-moraine deposits.

Repeated drainage of glacial Lake Eagle spread alluvium more extensively over the Anchorage lowland at levels slightly lower than the Nunaka Valley channel. A final series of outbursts produced the Mountain View alluvial fan (mv, Fig. 3-ne,sw), the largest body of alluvium in the area. It is named from the community of Mountain View east of downtown Anchorage (Bartsch-Winkler and Schmoll, 1984b; Schmoll et al., 1996). Material in the fan grades in texture throughout its length from bouldery gravel near its head to well sorted pebble and cobble gravel in mid-reach to finer gravel and sand at downtown Anchorage, and finally to medium sand in the distal parts farthest southwest (Miller and Dobrovolny, 1959). From the vicinity of Elmendorf Air Force Base to its distal edge, the fan overlies the Bootlegger Cove Formation, as indicated in Fig. 3. Although parallel to the Elmendorf Moraine on its outbound side, the Mountain View fan does not represent outwash directly from a glacier terminating at the moraine, originating as it does from Eagle River valley adjacent to the glacier. Local outwash fans along the moraine are truncated at the northwest side of the Mountain View fan, indicating that it developed sometime after the maximum extent of the Elmendorf glacier. Alluvial deposits that may be distal remnants of the

outwash occur in channels higher than the Mountain View fan that extend south from it and that are termed Patterson Street and Cheney Lake deposits (Schmoll et al., 1996).

In addition to supplying copious quantities of gravel aggregate especially from pits developed in its middle reach, the Mountain View fan has provided a good foundation for easy development of both the city and the military installations. The Mountain View fan was originally termed, by inference at least, Naptowne outwash (Miller and Dobrovolny, 1959). The deposits of the fan are still referred to that way particularly in many engineering geology reports. Because the Elmendorf Moraine is now correlated with only a small part of the Naptowne morainal complex, and the Mountain View fan is not outwash in the strict sense, this terminology is no longer appropriate.

As the Elmendorf glacier continued to wane, drainage from Eagle River valley found successively lower routes to the inlet (Miller and Dobrovolny, 1959). First, ancestral Eagle River developed the multi-level channel now occupied by underfit Fossil Creek (fc, Fig. 3-ne), which cuts deeply into the Elmendorf Moraine. Ice was still present in the area, however, because upper levels of the channel lead to Roosevelt Road, Gwenn Lake, and Sixmile Lake kames and kame terraces, all of which lie adjacent to or within the channel now containing Otter Lake and Sixmile Lake, north of the moraine (rr, gw, and sx, Fig. 3-ne). Later, the river deepened the Fossil Creek channel and flowed on the floor of the Otter-Sixmile channel. Finally, the Eagle River took its present more northerly course to reach Knik Arm through Eagle River Flats, the large filled embayment that cuts across Elmendorf ground moraine (ERF, Fig, 3-ne).

When lowland glaciers formed the several lateral moraines along the Chugach Mountain front, Ship Creek valley was blocked, forming glacial Lake Ship as discussed above. When this lake first drained, the water flowed through marginal parts of the glacier and developed prominent kame fields. Later, drainage from the valley fed streams that deposited Klatt Road alluvium. Still later, alluvium more clearly identifiable as related to ancestral Ship Creek was deposited at lower levels corresponding to a variety of environments that obtained through time, as the creek emerged onto the lowland. The two highest levels of Ship Creek alluvium are deltaic to the Bootlegger Cove estuary. Subsequently an alluvial fan was graded first to the Nunaka Valley channel and later to the Mountain View fan. When Eagle River first formed the Fossil Creek channel, Ship Creek began to incise the Mountain View fan, flowing for some time through the valley of what is now the lower reach of Chester Creek (Fig. 3-sw) to join Knik Arm south of downtown Anchorage. Finally it incised its present channel north of downtown, perhaps within the Holocene.

5. Holocene deposits within the Chugach Mountains

5.1. Glacial deposits

Holocene glacial deposits within the Municipality of Anchorage are found only in valleys within the Chugach Mountains. They consist mainly of (1) small end- and lateral-moraine complexes most of which have been considerably eroded by modern streams emanating from present-day glaciers, and (2) relatively restricted outwash deposits both in terraces bordering the streams and as aprons in front of glaciers. They have been divided into three groups based on distance from modern glaciers, degree of preservation, and relative age. From oldest to youngest these are called Winner Creek, Deadman, and Tunnel deposits (Fig. 2B). None of them except the youngest Tunnel moraines is dated with any degree of certainty.

Especially in the upper parts of the Girdwood valley system, there are several well preserved, massive end moraines which are named Winner Creek, after a tributary valley (Fig. 2B-sw; Schmoll and Dobrovolny, unpublished mapping; Schmoll and Yehle, 1983). It is not clear why these moraines are well developed in several places in this relatively small area but occur as remnants at only widely scattered places elsewhere in the Chugach Mountains. Few exposures are present and we have no direct evidence for moraine ages. Their position well upvalley from the dated late Pleistocene Bird Creek moraine in Turnagain Arm valley, but downvalley from both Deadman and Tunnel moraines, suggests an early Holocene age. On the west side of Cook Inlet, the Chichantna moraines mapped by Yehle et al. (1983) are dated at about 8000 yr BP; evidence elsewhere in the world also suggests glacial advances at about this time (Röthlisberger et al., 1980; Alley et al., 1997).

We have found a few well-preserved moraines upvalley from Winner Creek moraines where they exist, and a few kilometers distal to the Tunnel moraines. In other valleys, however, there are only remnants of lateral moraines and some erosional evidence (such as valley wall configurations and areas of ice scour) for glacier terminal positions; such remnants are particularly well marked near Twentymile Glacier (Fig. 2B-sw). Karlstrom (1964) mapped Tustumena moraines, named from a major glacier on the Kenai Peninsula, at comparable positions in some valleys there and also south of Portage. On the basis of limited radiocarbon data, he regarded Tustumena moraines as mid-Holocene and formed by a series of post-altithermal (to use his terminology) ice advances. Having little evidence to correlate features in Anchorage valleys with moraines near Tustumena Glacier, we prefer to name them Deadman moraines and associated features, after a small glacier south of Portage (Schmoll and Yehle, 1983). Their relative position in each valley is permissive of an age of about 3500 yr as pro-

posed by Karlstrom, but we have no new radiocarbon data.

Glaciers are still present within the Chugach Mountains in the upper parts of most larger valleys and locally prominent end moraines, and related features such as outwash aprons and terraces are present within a few kilometers of most of them. In some valleys there is a set of two to four such moraines, and in others a single moraine or only outwash deposits. Although subdivided into Bartlett Glacier (older) and Portage Glacier (younger) by Schmoll and Dobrovolny (unpublished mapping) and Schmoll and Yehle (1983), we here retain the name Tunnel for all of these young moraines, following the usage of Karlstrom (1964). They are named from a station on the Alaska Railroad just south of Spencer Glacier (Fig. 2B-sw). He considered that the moraines ranged in age up to about 1500 yr based on three radiocarbon dates, the validity of which we have not been able to confirm. Such moraines generally have been regarded as products of the Little Ice Age of a few hundred years ago, and the moraines in this area are thought to be of this age by Crossen (1992, 1997a, b) based on a synthesis of radiocarbon dates and lichenometric and tree-ring data.

Most glaciers have been retreating for at least the last 100 yr, probably with some intervening minor readvances. Many glacier terminal positions of the 1990s differ somewhat from those mapped by Dobrovolny and Schmoll in the 1960s using mainly 1957 air photos, from which the positions shown in Fig. 2 are derived. At the scale of Fig. 2, however, the differences would hardly show; only at Portage Glacier do we attempt to show the estimated 1997 glacier position. This glacier which until very recently terminated in Portage Lake has retreated drastically during the 20th century (Schmidt, 1961; Mayo et al., 1977; Crossen, 1997a, b).

5.2. Rock glaciers and rock-glacier deposits

Active rock glaciers and the deposits of formerly active rock glaciers may be considered transitional features that have characteristics of both true glaciers and colluvium. Active rock glaciers are thought to contain clear ice at depth (Moore and Friedman, 1991) but are fed copious supplies of coarse debris from talus at their heads. This material is still in the transport mode, albeit moving very slowly and having the appearance of a deposit. Rock-glacier deposits result when an active rock glacier ceases its forward movement presumably because there is not sufficient ice present to provide a lubricating mechanism. Although the distinction between the two forms is not always readily apparent, rock-glacier deposits lack the active rock glacier's steep front from which clasts are falling and which commonly exhibits a prominent moisture zone. Older rock-glacier deposits might be regarded as a type of ground moraine, but generally the surface morphology is more massive and uniformly hummocky

in appearance, and well-developed flow fronts character-
istic of younger landforms are still recognizable.

Rock glaciers and (or) younger rock-glacier deposits
occupy many cirques in the higher parts of the Chugach
Mountains that are not occupied by present-day glaciers
(Fig. 2); some cirques contain both small glaciers and
rock glaciers. Although we have no dates for them, the
younger rock-glacier deposits are almost certainly of late
Holocene age. Older rock-glacier deposits occupy rela-
tively long areas farther down some valleys that lead
from cirques, especially those leading generally north-
ward. The age of the older rock-glacier deposits is poorly
known; some might be of early Holocene age, but others
could be of late Pleistocene age, especially those in val-
leys that lack evidence of late Pleistocene lateral or end
moraines.

5.3. Colluvial deposits

A large part of the Chugach Mountains within the
Municipality of Anchorage is covered by variably thick
colluvial deposits that form a nearly continuous cover on
all but the uppermost and steepest slopes. These deposits
have been subdivided for mapping purposes into several
units, each of which is somewhat different in origin and in
engineering geologic characteristics. On smoothly slop-
ing surfaces the colluvium is derived mainly from well-
jointed and fractured bedrock; locally it is admixed with
minor alluvium in rills too small to map separately. On
many somewhat irregular slopes, where remnants of lat-
eral and ground moraine are too small to map separately
or are thought to be present at shallow depth, colluvium
is derived from both bedrock and glacial deposits. On
slopes below the proximal side of some lateral moraines,
colluvium is derived largely from morainal deposits. Es-
pecially in areas where the underlying bedrock, mainly
the Valdez Group, is fine grained and easily frost shat-
tered, the surface materials appear to be creeping slowly
downslope, forming an irregular surface commonly
having many small lobes with relief of a few tens of
centimeters. We interpret these as solifluction deposits.

The highest part of the Chugach Mountains within the
Municipality lies in a belt that extends from the head-
waters of South Fork Eagle River (Fig. 2A-se) to the
vicinity of Bashful Peak (Fig. 2B-nw), at 2440 m, the
highest point in this belt. Outcrops of slightly metamor-
phosed graywacke mapped as part of the McHugh
Complex are common, and colluvium consists largely of
talus accumulations, many of them the products of rock-
falls that contain clasts in excess of a meter in major
dimension.

Landslides are a form of colluvium; they are quite
common within the mountains and are present at some
lowland localities as well. The latter are discussed later in
relation to seismic activity, but some landslides within
the mountains were probably triggered by earthquakes

as well. Several landslides in the mountains cover more
than one square kilometer, and one dams the lake shown
along South Fork Eagle River (Fig. 2A-se). Major land-
slides are especially prevalent in a belt that extends
northeastward from there to Hunter Creek (Fig. 2B-ne)
(Schmoll and Dobrovolny, unpublished mapping).

A special type of slope process, albeit one that com-
monly does not give rise directly to major colluvial de-
posits, is inferred from small, narrow, bedrock-flanked
trenches that are interpreted as sackung features. The
trenches occur along or just downslope from and sub-
parallel to the crests of a few high mountain ridges.
Sackung features are thought to have formed through
gravitational spreading of a ridge by gradual displace-
ment along a series of disconnected planes or by deep-
seated plastic deformation of the rock mass without
formation of a through-going discrete slide surface
(Zischinsky, 1966; Radbruch-Hall, 1978; Savage and
Varnes, 1987; Varnes et al., 1989). Conditions especially
conducive to sackung formation are thought to include
oversteepened valley walls left gravitationally unstable
after retreat of a glacier. Sackung trenches are present in
several areas and are especially well developed on the
ridge northwest of Ingram Creek (Fig. 2B-sw) and along
the Chugach Mountain front adjacent to Knik River
north of Eklutna Lake (Fig. 2B-nw).

6. Holocene deposits of the Anchorage lowland

In the Anchorage lowland, Holocene deposits are
widespread but significantly thick only locally. They in-
clude (1) pond and peat deposits; (2) eolian deposits that
include minor tephra beds and that locally form dunes;
(3) minor alluvium along streams; (4) colluvium on bluffs
that border both stream valleys and the inlet; and (5)
landslides caused mainly by earthquakes (discussed
separately later).

Ponds formed mainly in the many shallow depressions
on the surfaces mainly of glacioestuarine deposits and
Elmendorf ground moraine. Organic silt and clay, and in
places marl, accumulated in the ponds. Some of the larger
ponds are still extant, but many have been filled in by
accumulations of mostly organic debris that have formed
peat; the lowest peat beds may be of latest Pleistocene
age. In some places peat is as much as 10 m thick.
Elsewhere, especially in channels no longer occupied by
major streams, and overlying the Hood Lake sand as
well, peat that is as much as 2 m thick has accumulated
on relatively large areas that originally may not have
contained ponds.

Peat deposits at the top of the bluff at the Clay Prod-
ucts Road site (CPR, Fig. 3-sw) are about 0.5 m thick;
within about 200 m to the west, they thicken to about
4 m, and then thin farther to the west. Radiocarbon ages
of about 12,000 yr have been obtained from the base of

this unit. Within it, a prominent volcanic ash bed has a radiocarbon age of about 3300 yr, comparable in age to a set of ashfalls from Hayes volcano about 150 km northwest of there (Fig. 1). A sample from this site indicates that this informally named volcano is the source for the bed (Riehle, 1985) and that the bed is part of Hayes tephra set H (Riehle et al., 1990; Riehle, 1993). Ashfalls from other volcanoes of the Aleutian volcanic arc (Fig. 1, inset) have also reached the Anchorage lowland, most recently in 1992 from Crater Peak on Mount Spurr (Fig. 1; Keith, 1995).

A ubiquitous mantle of organic and eolian deposits, commonly less than a meter thick, overlies most other deposits but is not mapped separately from them. The mantle includes mainly silt, fine sand, and some thin tephra beds overlain by a variably thick surface vegetation mat; it grades laterally into peat deposits thick enough to map separately. Eolian deposits are mapped separately only in the two areas where there are well-developed cliffhead dunes. Both areas are near the mouth of Turnagain Arm, a locus of frequent and strong winds that blow northwestward down the arm. One area is along the southeast side of Fire Island (Fig. 2A-sw) and the other along the bluffs that extend southeast from Point Campbell (Fig. 3-sw). Some dunes are stabilized, but others are actively developing and in places encroaching onto adjacent terrain.

Bluffs bordering Knik and Turnagain Arms and the steep walls of valleys incised into surficial deposits were formed originally by erosion and most are subject to renewed erosion. These bluffs are veneered by a down-slope-thickening wedge of colluvium derived mainly from material into which the bluff was cut during the last episode of erosion. Where the Bootlegger Cove Formation or other fine-grained deposits are present beneath the colluvium, potential for instability is high and landsliding is common. One such area is east of the Port of Anchorage (Fig. 3-nw) where landsliding on the bluffs has been analyzed in some detail (Varnes, 1969; Updike, 1986; Updike and Carpenter, 1986). Bluffs along abandoned glacial meltwater channels probably formed in the late Pleistocene, whereas those along modern stream channels probably formed relatively early in the Holocene; neither of these have changed much in position since then. Bluffs along Knik and Turnagain Arms and around Fire Island, however, have migrated significantly landward through time as erosion has progressed and the present-day bluffs are probably late Holocene in age; good exposures attest to continuing erosive activity (Fig. 2A). Bluff erosion is not active everywhere, however. Along the southwest side of the Anchorage lowland, the fairly low bluffs seem stable, perhaps protected by a broad accumulation of Turnagain Arm deposits (Fig. 3-sw).

Point Woronzof (Fig. 3-sw) is of interest in this regard because of the amount of bluff erosion and shift in the shoreline that has occurred within the last 100 yr or so. Miller and Dobrovolny (1959) used a 103-ft (31-m) altitude triangulation station lying on its side on the beach to deduce that the bluff line near the point had receded about 28 m between 1909 when the first station was installed and 1953, yielding a recession rate of about 0.6 m per year. More recent measurements suggest a slower rate: J.R. Riehle, then Alaska Division of Geological and Geophysical Surveys (unpublished data), compared 1969 and 1978 air photos and estimated a rate of about 0.3 m per year. Measurements by A.D. Pasch, University of Alaska, Anchorage, and by us indicate a rate of about 0.4 m per year (Pasch, 1987) that continues to the present. Bluff recession in this vicinity probably is faster than at most other sites in the region; tidal currents may be stronger here as they course past the point.

7. Holocene estuarine deposits

Estuarine deposits form a belt around much of the margin of the Municipality where it borders Knik Arm and Turnagain Arm, but there is considerable varation in the width of the belt and the thickness of the deposits. They were shown as Turnagain Arm deposits by Schmoll et al. (1997) following unpublished data by Schmoll, Bartsch-Winkler, and Yehle, and this term is used here as well. These deposits are best developed in filled embayments such as those in the lower Twentymile and Placer River valleys at the head of Turnagain Arm (Fig. 2B-sw) and the Eagle River Flats (ERF, Fig. 3-ne); the latter is paired with a similar embayment at Goose Bay on the opposite side of Knik Arm (Fig. 2A-nw), and Yehle et al. (1991) term these two features the Goose-Eagle alignment. Elsewhere the Turnagain Arm deposits form a long and relatively narrow border, such as along the south side of the Anchorage lowland (Figs. 2A-sw and 3-sw); they also occupy the area between Fire Island and the mainland. Where high bluffs border the arms, there are only narrow beaches; at low tides, however, the beaches broaden and are relatively steep, extending downward to levels as much as 10 m lower than the base of the bluff. The Turnagain Arm deposits consist mainly of a mixture of silt and fine to medium sand, except that the beaches are generally sandy to gravelly. Many of these deposits formed in the intertidal zone, but those in the filled embayments are older deposits no longer being actively reworked. Elsewhere, however, deposits are reworked daily by the tides; especially in Turnagain Arm and in the area between Fire Island and the mainland, large areas are alternately submerged and subaerially exposed, although not as very dry land.

Turnagain Arm deposits locally contain thin to relatively thick organic beds, including peat and buried tree stumps, overlain by intertidal deposits. Karlstrom (1964)

related the burial of these organic deposits to higher stands of sea level during Holocene interstadial intervals when glaciers were less extensive than at present. During the 1964 earthquake, however, the area around the upper end of Turnagain Arm subsided as much as 2.4 m (McCulloch and Bonilla, 1970) and a large area was immediately inundated. One result of this event was deposition of a new geologic unit, the Placer River Silt (Ovenshine and others, 1976), which buried the pre-earthquake ground surface with its extensive vegetation cover. By analogy to this seismically caused burial, an alternate interpretation of the older buried organic beds evolved, namely that their burial also might have been induced seismically. Efforts to infer recurrence intervals of great earthquakes such as that of 1964 by dating the older buried organic horizons have been undertaken especially in the vicinity of Girdwood (Fig. 2B-sw) by Combellick (1986, 1991, 1993, 1994, 1997); Combellick and Reger (1994) and by Bartsch-Winkler and Schmoll (1987, 1992). The studies have not yielded precise recurrence intervals, but suggest that they are on the order of several hundred years, although much shorter intervals are deemed possible. Recurrence intervals of the same order of magnitude were also suggested by repeated horizons of contorted bedding in Turnagain Arm deposits southwest of Point Woronzof (Fig. 3-sw) and at Goose Bay (Fig. 2A-nw) (Bartsch-Winkler and Schmoll, 1984a).

Because of the interest in dating past earthquake events, an unusually large number of radiocarbon dates are available from the Turnagain Arm deposits in the reports cited above. In general, ages of about 3000 yr have been obtained at depths of about 10 m on material both in bore holes and subaerially exposed at low tides and of about 4000 yr from a few meters deeper in some bore holes. In a test hole cored to 93 m at Portage (Bartsch-Winkler et al., 1983), an extrapolated age of about 8000 yr is given for the base of the core; an adjacent water well extends to a depth of about 300 m. Turnagain Arm deposits can be considered to occupy the upper 200 m of the deposits beneath Portage; deposits interpreted as glacial (possibly equivalent to the Bird Creek moraine) occur below that depth. The core and water-well log together could serve as the type locality, if the Turnagain Arm deposits become formalized as a formation.

8. Permafrost

The Anchorage lowland is included within a region of Alaska that is generally free of permafrost (Ferrians, 1965; Brown et al., 1997). However, permafrost has been reported or is suspected at a few sites. One of these is at locality MR (Fig. 3-se) where road reconstruction revealed several oval-shaped, tree-covered palsas (peat-cored frost mounds) about 150 m × 60 m in plan and

3.5 m in height. One palsa contained an irregularly shaped core of perennially frozen peat and silt up to 7.5 m thick that contained numerous lenses of segregated ice (J.R. Williams, written commun., 1979, cited in O.J. Ferrians, Jr., written commun., 1993). At another site about 3 km to the west (CD, Fig. 3-se), a residential building sustained extensive damage attributed to the melting of permafrost a few years after construction. The house was removed and its lot and a few adjacent ones have become a small park. At two construction sites within a few kilometers of these places, permafrost was encountered during preconstruction investigations, and amelioration measures were undertaken before construction proceeded.

Within the Chugach Mountains, isolated masses of permafrost are likely to exist in selected areas (Ferrians, 1965). Their location is hard to predict because of the variability in geologic material, orientation of the ground slope, vegetation cover, and microclimate factors responsible for mean annual air and ground termperatures at any given site. The presence of what appear to be solifluction deposits suggests that permafrost is present also, but no firm conclusions can be drawn without detailed investigations to determine whether permafrost, melting of seasonal frost, or some other lubricating mechanism is responsible for apparent downslope movement of the material (Schmoll et al., 1996).

9. Eustacy, isostasy, and tectonics

The discussion of glacial and glacioestuarine geology contains reference to changes in relative level of the land and sea, but without considering absolute levels. Worldwide sea level, of course, was eustatically much lower during glacial times than at present. Glaciers that advanced into Cook Inlet basin must have lowered the land surface isostatically. The net result was that both sea level and the local land surface were lower than at present during glacial maxima and also when the late-glacial Bootlegger Cove Formation was deposited. Inlet water at that time was at levels that are now more than 20 m to perhaps as much as 140 m above present sea level. We have not sufficiently evaluated the relationship between eustacy and isostasy to determine the extent to which tectonic effects might have played an additional role in determining the present altitude of the Bootlegger Cove and related deposits. Karlstrom (1964) regarded the upper Cook Inlet area as "relatively stable", but a few months after publication of that report, the 1964 earthquake seemingly demonstrated otherwise. Although the same earthquake also produced areas of uplift (Plafker, 1965), the net effect of a series of presumably similar earthquakes over the last several thousand years has been subsidence at least in Turnagain Arm, as discussed above. Cook Inlet basin is also an area of

long-term subsidence, since the subsurface beneath the Anchorage lowland, for example, preserves deposits of past glaciations that probably formed above sea level but are now 20–300 m below it. Such relative subsidence is contrary to the relative emergence of the Bootlegger Cove Formation, suggesting that, at present, long-term subsidence has yet to overcome short-term (14,000-yr) postglacial emergence.

Some other neotectonic and structural features in and around the Anchorage lowland may bear on glacial and glacioestuarine history. Growth anticlines aligned similarly to anticlines shown by Magoon et al. (1976) have been proposed as a mechanism to account for the splaying out of raised beach ridges at several embayment margins around Cook Inlet (Kelly, 1961; Tysdal, 1976), notably at Chickaloon Bay (Fig. 2A-sw) and north of Fire Island on the north side of Cook Inlet (Fig. 2A-nw). The position of the Fire Island deposits at the west end of the Anchorage lowland also has been attributed possibly to growth of an anticline, although as discussed above not as a preferred interpretation (Schmoll et al., 1984). Haeussler and Bruhn (1996) summarize evidence for young folding in the region. Releveling along the Alaska Railroad in Turnagain Arm valley several years after sudden subsidence caused by the 1964 earthquake (Brown et al., 1977) revealed a pattern of uplift attributed to crustal relaxation following the 1964 earthquake; the results alternatively could be ascribed to local nonseismic vertical movement. Bird Point (Fig. 2A-se), a bedrock high projecting into Turnagain Arm valley, is not readily explained as a glacier terminal position; it lies within the zone of maximum post-earthquake uplift and could be explained as a product of such uplift if it persisted over a long period of time.

A feature that Yehle et al. (1990) speculate might be structurally controlled is the Goose-Eagle alignment of paired embayments that transects Knik Arm (Fig. 2A-nw). This feature is transverse rather than parallel to mapped anticlines, but it does parallel a second-order structural grain in the region expressed by the orientation of valleys in the Chugach Mountains and readily apparent on satellite imagery (Riehle et al., 1997). The embayments seem to have formed subsequent to the Elmendorf ice advance because Elmendorf deposits are truncated by the embayments. The alignment might be related to local anticlinal growth, as has been proposed for other embayments in the region (Kelly, 1961), or some other structural mechanism. Perhaps, however, the alignment may have originated as part of the Eagle River valley drainage sequence. If so, the impounded water had sufficient head to carve a subglacial passage across what is now Knik Arm, to emerge near the head of the Goose Bay embayment. In support of this idea, there is a large alluvial fan west of the Goose Bay embayment, somewhat comparable in size to the Mountain View fan. That alluvial fan heads at the northeast end of the

embayment where it appears to lack any other source. Even in this scenario, however, structure might have guided development of the drainage sequence.

The prominent Chugach Mountain front is spectacular when viewed from the lowland, and even more striking when seen on satellite images (Riehle et al., 1997). Yehle and Schmoll (1989) were probably not the first to suggest that the abruptness of this front might have been caused by normal faulting even though they found no faulted deposits. They were not convinced by the local evidence for recent normal faulting (basin side down) along the mountain front mapped by Updike and Ulery (1983) and ascribed by them to late Holocene activity within the Border Ranges fault zone in the vicinity of Twin Peaks (Fig. 2B-nw). Recent interpretations by Haeussler and Anderson (1997) dispute the evidence for active faulting at Twin Peaks. Such faulting would have a sense of motion opposite to that ascribed to the Border Ranges fault (basin side up, Winkler, 1992) which is only approximately coincident with the mountain front. Riehle et al. (1997), on the other hand, suggest that the Border Ranges fault may have been reactivated along the Chugach Mountain front segment, with a change in sense of motion, in response to sedimentary infilling of Cook Inlet basin. However, they make no claim for active faulting.

An east-west seismic refraction profile in the southern part of the Anchorage lowland indicates the presence of a normal fault (west side down), the Abbott Loop fault, about 5 km northwest of the Chugach Mountain front and within the broad Border Ranges fault zone (Schmoll et al., 1996). The profile indicates that the fault does not reach the surface, and no surface evidence of faulting has been found. However, a nearby alignment of low-lying glacial meltwater channels and ponds is parallel to the projected trace of the fault. The seismic profile also indicates a matching normal fault in the subsurface with opposite sense of motion (east side down) in the vicinity of the Fire Island deposits. As discussed previously, such a fault lends credence to the idea that Fire Island deposits could have been tectonically uplifted. The Bootlegger Cove Formation and similar older deposits would then have accumulated in a graben that served as their depocenter.

If the present height of the Chugach Mountains is attributable to uplift during the Quaternary, glacial features high on the mountains might have been formed at substantially lower altitudes than they now occur. Instead of indicating much more extensive glaciers, these features might have been the products of glaciers whose extent was similar to those that produced the lateral moraines lower on the slopes. This idea conflicts with the long-held notion of old glaciations in Alaska during which glaciers extended far beyond the limits indicated by the well-formed moraines of younger glaciations (for example, Péwé et al., 1953; Péwé, 1975). That concept has

held up well in many parts of Alaska, however (T.D. Hamilton, U.S. Geological Survey, oral commun, 1997), and we continue to subscribe to it here as well.

10. Earthquake-induced landslides

In addition to the subsidence that inundated the area at the upper end of Turnagain Arm, liquefaction effects caused extensive damage to both the Alaska Railroad and the highway along the arm (McCulloch and Bonilla, 1970; Walsh et al., 1995). However, the major damage at Anchorage in 1964 was caused by landslides, especially those in areas where sensitive clay of the Bootlegger Cove Formation is exposed or thinly concealed in bluffs that provided a free face along which the landslides could develop. Geologic mapping in these areas by Miller and Dobrovolny (1959) indicated the presence of older large-scale landslides that they thought could have been earthquake induced; they inferred that future large earthquakes might produce new, similarly large, landslides. Within five years of the publication of their report the validity of their inference was confirmed.

10.1. Pre-1964 landslide deposits

From its mouth to about 10 km upstream, both Ship Creek and Chester Creek are sufficiently entrenched into the surface of the Mountain View fan that silt and clay of the underlying Bootlegger Cove Formation are present beneath colluvium in the lower part of bluffs bordering these valleys. In the vicinity of a broad north-facing embayment in the Ship Creek valley wall, as much as 15 m would be revealed if the colluvium were removed. This area was identified as one such older landslide and is here called the Third Avenue landslide (OLT, Fig. 3-nw). Although the area has been urbanized, the sloping ground surface retains sufficient slight irregularities to suggest landslide morphology. If, instead, the embayment were related to erosion by Ship Creek, a smoother, flatter surface would be expected. The embayment is similar in shape and dimension (about 500 m wide and 350 m from head scarp to toe) to the nearby 1964 Fourth Avenue landslide. We conclude that this area is most likely the site of an older landslide that was triggered by a previous large earthquake. There were no reported effects of the 1964 earthquake within this slide area, and it appears to be an example of an old slide that has stabilized naturally.

Mapping by Miller and Dobrovolny (1959) in the low area near Inlet View School (OLI, Fig. 3-nw) does not indicate an old landslide, but with hindsight of the earthquake-caused landslides of 1964, this area is so regarded by Hansen (1965) and so mapped by Schmoll and Dobrovolny (1972). The shape of the putative head scarp, the proximity to documented slides, and the similar geologic and topographic setting all suggest a slide. This area also remained stabilized in 1964.

Two large landslides are tentatively recognized elsewhere in the Anchorage lowland in mapping by Schmoll et al. (1996) at sites where the Bootlegger Cove Formation is not present. One is adjacent to the Dishno Pond lateral moraine (OLD, Fig. 3-ne), and it is interpreted that a segment of the lateral moraine about 300 m wide may have slid into glacioestuarine water that encroached upon the area shortly after the glacier withdrew, spreading distally about 1.5 km. Landforms near the toe of the slide are quite subdued, as if they were modified subaqueously. The other landslide developed in glacioestuarine deposits a few kilometers to the southwest near South Fork Chester Creek. Although there is not much relief here, it appears that the side of a low hill about 200 m wide slid out over a distance of about 500 m into lower-lying terrain, possibly subaqueously. We speculate that landslides of this magnitide, similar in size to some of the 1964 slides, might have been earthquake induced.

10.2. Landslides caused by the 1964 earthquake

The major destructive landslides of 1964 were located in suburban Turnagain Heights (QLTH, Fig. 3-sw) and in downtown Anchorage (Fig. 3-nw, showing only the largest slides); a large landslide occurred on the opposite side of Knik Arm in an undeveloped area (QLBP, Fig. 3-nw; the informal term "Bridge Point" refers to a bridge not built). Most of the slides moved by a combination of translation and rotation of blocks of land, in places sliding on nearly horizontal slip surfaces. Most structural damage to buildings occurred in grabens at the head of each slide and within relatively low pressure ridges at the toe (Updike et al., 1984). It is likely that slides of this magnitude developed only because of the 3- to 4-min duration of strong ground motion; as much as 7 min of shaking was reported in or near some of the major landslides (Hansen, 1965). Three of these slides are discussed briefly and shown in Fig. 5.

The Turnagain Heights landslide was the largest, most complex, and physiographically most devastating of the 1964 Anchorage landslides. The description of this landslide is adapted from Hansen (1965) in which there is a large-scale (1 : 2400) topographic map of the landslide area. The Turnagain Heights housing subdivision (from which this slide is named) is misleadingly named, since it faces Knik Arm rather than Turnagain Arm.

As seen in plan, this slide consisted of two main lobes: an East Turnagain lobe and a West Turnagain lobe, each with its own head scarp. Each lobe probably started as a separate landslide, but the two lobes merged laterally at a northward-projecting salient of terrain centered along Hood Creek that was spared of sliding. The two lobes together extend for about 2.6 km along the bluff, with a maximum headward retrogression from the old bluff
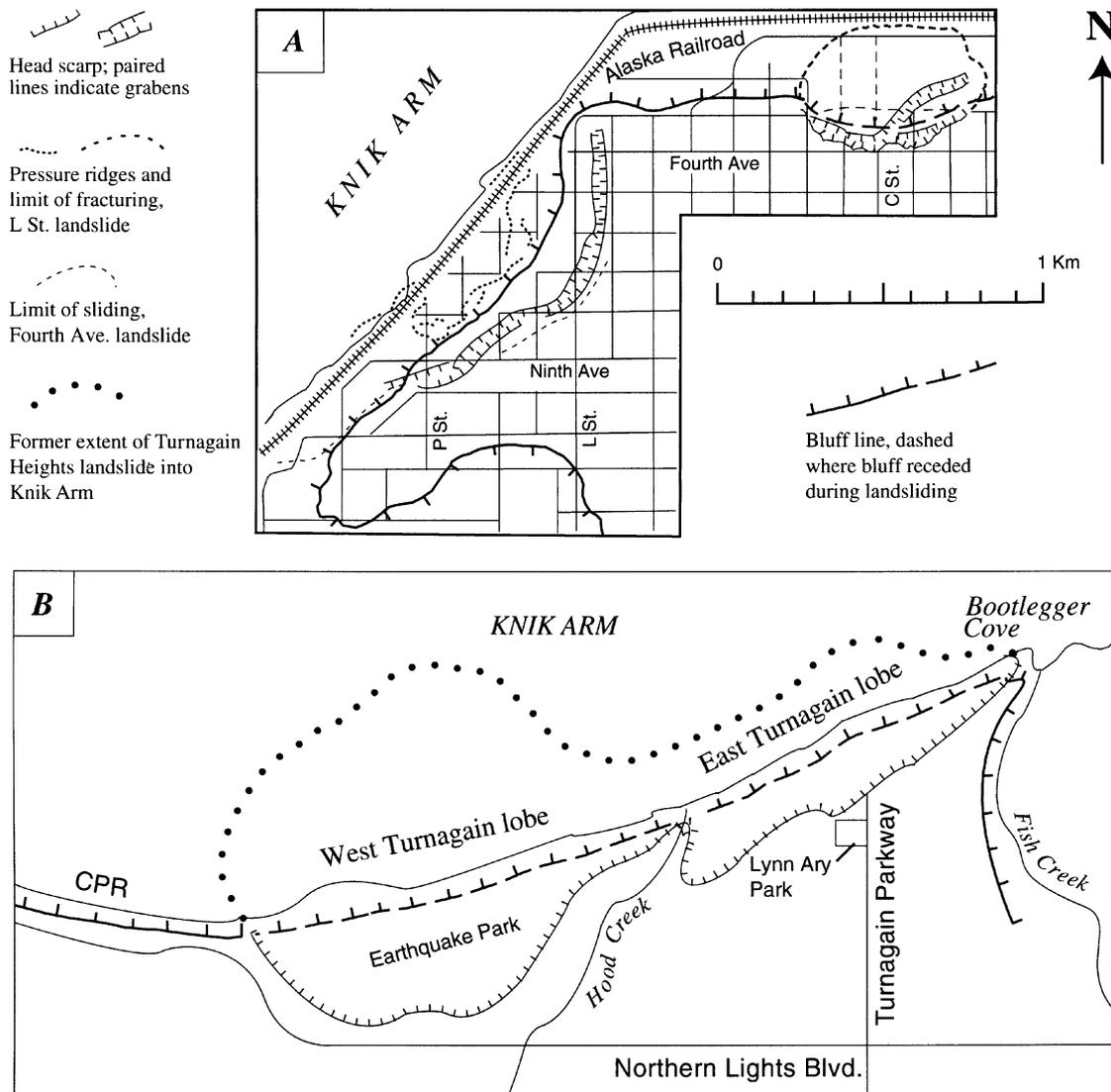
Fig. 5. Outline maps of three major landslides in the Anchorage area that occurred during the 1964 Alaska earthquake: (A), L Street and Fourth Avenue landslides; older Inlet View landslide occupies the area between P and L Streets near the south edge of the map. (B), Turnagain Heights landslide; CPR, Clay Poducts Road site; only major streets are shown. See Fig. 3 for position of these landslides in relation to the Anchorage lowland.

line of about 365 m in the West Turnagain slide. In the densely residential East Turnagain slide, retrogression was only about 150 m, but about 75 homes were destroyed and four persons lost their lives. Behind the two lobes, hundreds of visible tension fractures opened up over a distance of as much as 670 m behind the East Turnagain lobe (Engineering Geology Evaluation Group, 1964). Besides causing significant structural damage to housing in this area, these fractures totally disrupted all underground utilities and seriously damaged streets and curbing. The distance between Northern Lights Boulevard and the head of the slide as measured along Turnagain Parkway increased by about 1 m, this amount distributed among the numerous fractures. The ground surface within the slide area was lowered an

average of about 10 m, and the volume of material moved was about 12 million cubic meters. Lateral spreading extended the slide seaward, as the leading edge glided onto the tidal flats and into Knik Arm beyond the low tide line. The leading edge in fact extended almost twice as far beyond the old bluff line as the head of the slide regressed behind, as shown by the dotted line in Fig. 5. The new shoreline thus created was quite ephemeral, however, and mean high water now reaches to about its former position; at low tide levels, however, some of the lumpy slide debris is still visible.

Urbanization had not reached west to the area of the West Turnagain lobe by 1964, because the area is less well drained and would have been more expensive to develop. In that area the cover of Mountain View

deposits that barely reaches westward to the East Turnagain lobe is lacking, and the Bootlegger Cove Formation is overlain only by thin Hood Lake sand deposits commonly with a veneer of peat. Fortunately, the opportunity was taken to preserve the West Turnagain lobe in its natural state as Earthquake Park, a memento of the catastrophic event and a popular tourist attraction. In the years soon after 1964, the freshly squared off or jagged landslide blocks in jumbled array were an impressive sight and looked very much as they did when slide movement ended; they were free of vegetation, as the pre-slide vegetation cover was torn asunder. By now, however, erosion and natural revegetation have taken their toll, and pathways merely provide a stroll through woods over lumpy ground. The present scene, however, is more typical of what one observes in suspected landslide terrain rather than the more spectacular morphology of the immediate post-slide period.

The Fourth Avenue landslide (QLFA, Fig. 3-nw) receded headward to the center of the most northerly east-west thoroughfare that lies entirely on the Mountain View alluvial surface within downtown Anchorage. The bluff line north of Fourth Avenue is irregular in both plan and height; much of this irregularity is undoubtedly due to former landsliding (Miller and Dobrovolny, 1959; Schmoll and Dobrovolny, 1972). Between A Street and E Street the pre-1964 bluff line was as far south as mid-block between Third and Fourth Avenues. Buildings on the north side of Fourth Avenue in this four-block stretch thus were situated on the Mountain View surface, but behind (north of) them the land surface dropped off irregularly. It is this stretch that failed spectacularly during the 1964 earthquake, destroying nearly all buildings on that side of the avenue.

It appears that the Fourth Avenue landslide formed in an area of previous, probably similar, sliding, causing the head scarp to retreat about another 60 m. The 1964 slide involved all or parts of 14 city blocks and about two million cubic meters of material. Vertical displacement was as much as 3 m, mostly within a graben that formed at the head of the slide. Horizontal movement was as much as 5.5 m; there were additional minor displacements (in fractures parallel to the head scarp) as far as 180 m to the south (Hansen, 1965).

To stabilize this economically important slide area, the US Army Corps of Engineers constructed an earthen buttress (mainly gravel and sand borrowed from the Mountain View deposits near Mountain View), and the present land surface is the engineered surface of this mass of new material. Details of the design of the buttress are discussed in Long and George (1966). A number of specifically designed metal slope-indicator casings were installed around the slide area. These casings can be surveyed by means of an electronic probe inserted down the casing. Updike (1983) examined these casings and concluded that a few centimeters of movement had

occurred since 1964. Low pressure ridges have continued to form just beyond the base of the buttress (Schmoll and Kachadoorian, unpublished mapping), also suggesting that there has been some on-going movement.

The L Street landslide (QLLS, Fig. 3-nw), like the Fourth Avenue landslide, is also located at the site of a previous landslide. Excavation for some of the (post-1964) buildings outbound from the 1964 slide block have revealed chaotic mixtures of silt and clay and buried trees typical of landslide material. The 1964 slide here was somewhat different from the other major 1964 landslides in that almost no vertical displacement occurred; movement was nearly all horizontal as a huge translational gliding block moved intact, producing a graben behind it. This slide involved all or parts of 30 city blocks and was about 1200 m wide and about 300 m from head scarp to toe. The graben at the head of the slide was about 45–60 m wide and 3 m deep and most of the building damage occurred there. The translational block carried overlying structures, at least one of them four stories high, as much as 4 m horizontally, with surprisingly little damage. Low pressure ridges formed beyond the toe of the slide, disrupting railroad tracks along the shore. The L Street slide may be an example of a slide that would require even more time, severe shaking, momentum, and (or) a different direction of seismic pulsing to progress to the block-failure mode typical of the other Anchorage slides. A reminder of the movement that did occur is the present slight jog in L Street, which was originally straight.

The question of future stability in this area has not been resolved. Soon after the earthquake, the graben was back-filled with sand and gravel, the damaged structures were removed, and new construction followed. One cannot help but be dismayed or impressed by the relatively large buildings built since 1964 along the bluff in front of the L street slide block. Will these buildings tumble in the next slide, their added weight aiding slide movement, or will that same weight instead enhance the buttressing effect of existing slide material and decrease the extent of future sliding? Another question concerns the nearby bluff-point area at L Street and Third Avenue: there was no failure here in 1964. Was this because the sensitive zone within the Bootlegger Cove Formation has been better drained here and desensitized, or is this the next locality to slide?

In an effort to better understand the mechanism of landslide failures, extensive investigations were undertaken following the 1964 earthquake (Shannon and Wilson, 1964), and numerous workers analyzed samples and offered views. Two schools of thought emerged, one favoring liquefaction of thin sand beds within the Bootlegger Cove Formation as the chief culprit (Seed and Wilson, 1967), and the other that sensitive clay horizons were chiefly responsible (Long and George, 1966; Kerr and Drew, 1968).

To further evaluate the situation, two test holes were drilled and cored in 1978 in Lynn Ary Park behind the East Turnagain lobe (Updike et al., 1988; Olsen, 1989). Other engineering studies near the L Street and Fourth Avenue slides (Woodward-Clyde, 1982; Idriss, 1985) were undertaken within the next few years. This work concluded that a middle zone of the Bootlegger Cove Formation, previously identified as including the zones of weakness along which failure occurred, contains (1) sands that are denser, and hence less susceptible to liquefaction than was previously recognized, and (2) extremely sensitive clays that appear to be more abundant and more widely distributed than previously known. The sensitive clays therefore now appear more critical than the sands in governing failure of the Bootlegger Cove Formation during large-magnitude earthquakes. Research on the mechanisms by which the sensitive zone induces failure during large-magnitude earthquakes is continuing (Stark and Contreras, 1998). The Lynn Ary Park core also served as the basis for identifying engineering geologic facies within the formation that reflect subtle variations of the glacioestuarine environment (Updike et al., 1988). The drill holes at Lynn Ary Park, as represented by the logs of the cores, would be a better type locality for the formation than the slumped and stratigraphically less complete exposures in the bluffs near Bootlegger Cove.

11. Conclusions

Within the Municipality of Anchorage, evidence of all major glaciations but the recessional phase of the last one (late Pleistocene) is fragmentary. It is clear, however, that there were older (middle to possibly early Pleistocene, and even Miocene) glacial events, and during those for which evidence has been preserved, glaciers probably extended farther and were thicker than the late Pleistocene glaciers. The number and timing of these older events are not well known, but at least four, each with successively less extensive glaciers, probably occurred. It is also likely that there were other, intervening glacial events during which glaciers were about as extensive, or even less extensive, than those of the late Pleistocene but for which there is little or no evidence. A history of glacier retreat, inlet transgression and regression, and landscape formation, such as described below for the late Pleistocene, probably recurred numerous times.

Most of the glacial, glacioestuarine, and related deposits in the Anchorage lowland record events of the late Pleistocene when glaciers from the northeast, north, and probably northwest coalesced in the area southwest of Anchorage. They flowed some distance southwest down Cook Inlet but probably did not entirely fill it. Whether an abbreviated inlet was open to the sea or that area was instead the site of a lake dammed by coastal glaciers

farther south is not clear; if such a lake did form, however, it probably did not extend into the Anchorage area. During the time that glaciers receded from the late Pleistocene maximum, a series of lateral moraines suggests a pulsating retreat with minor readvances. The lack of terminal moraines within the southern Anchorage lowland suggests that as the glacier fronts retreated, inlet water was in contact with them and transgressed concommitantly. Even though worldwide sea level was substantially lower at the time of glacial maximum than at present, Cook Inlet basin must have been isostatically depressed sufficiently for the transgression to occur. There is no real age control for these events, but by analogy to events elsewhere they probably occurred over a time interval of several thousand years centered about 20,000 yr ago.

As glaciers receded, Cook Inlet transgressed northeastward and the Bootlegger Cove Formation accumulated in the center of what might have been an irregularly shaped upper inlet with two or more arms, similar to but configured somewhat differently from the present Cook Inlet. In the margins of the estuary and bordering it, other, generally coarser, deposits accumulated, some reworked or reshaped from deposits formed by the preceding glaciers. Minimal glacial influence and maximal contact with the open sea occurred about 14,000 to 15,000 yr ago as indicated by the shell-bearing horizon within the upper part of the Bootlegger Cove Formation. The estuary advanced at least as far inland as the present head of Turnagain Arm, probably a similar distance up Knik Arm, and at least 70 km farther north up the Susitna lowland (Fig. 1) than the present Cook Inlet. Subsequently, a last glacial readvance deposited the only major end moraine in the area, the Elmendorf Moraine, and the smaller Bird Point moraine formed in Turnagain Arm. The estuary soon regressed, however, to the confines of present-day Cook Inlet and its arms. By about 12,000 yr ago the glaciers retreated and have not occupied the Anchorage lowland since.

By the beginning of the Holocene the landscape began to take a form similar to that of the present. There was probably more land area at first, but coastal erosion gradually reduced the area to its present dimensions. At irregularly spaced intervals, commonly but not necessarily measured in hundreds of years, great earthquakes produced major landslides in select areas — mostly associated with the Bootlegger Cove Formation — that tended to hasten coastal retreat. Elsewhere land area was gained, as embayments and ponds became filled. Following great earthquakes, however, some relatively new embayment land was suddenly but perhaps only temporarily lost because of subsidence and compaction. In mountain valleys, glaciers had retreated to or beyond their present positions, but probably advanced and retreated periodically in response to minor climate change or internal glacier dynamics. Weathering reshaped

mountain slopes and thickened the colluvial fill on valley walls. At irregular intervals, landslides produced marked but local changes in topography.

The late Pleistocene geologic history we have presented differs in some substantial ways from that interpreted by most previous workers. In general, we regard many of the deposits of the Anchorage lowland as younger than they did. We consider the lateral moraines along the Chugach Mountain front to date from the late Wisconsin glacial maximum, whereas Karlstrom (1964) thought them to be of Illinoian age and Miller and Dobrovolny (1959) mapped them as early Wisconsin. Our evidence indicates that the Bootlegger Cove Formation represents a late glacial estuarine transgression; Miller and Dobrovolny considered a marine alternative but opted for a lacustrine origin for the formation during the early Wisconsin; to Karlstrom the formation was evidence for a glacial lake in both the early ["pre-"] and late Wisconsin with an intervening marine incursion that he considered interglacial, whereas most other workers regarded the same interval as interstadial. Schmoll et al. (1996) consider that the Bootlegger Cove estuary extended beyond the limits of the Bootlegger Cove Formation as originally conceived and map adjacent deposits as estuarine as well, although Updike and Ulery (1983) include some of these deposits within a more broadly defined Bootlegger Cove Formation. Karlstrom apparently regarded glacial Lake Cook as encompassing much the same area (although the marine incursion was regarded as more restricted in area), whereas Miller and Dobrovolny apparently restricted lacustrine deposition more nearly to the limits of the Bootlegger Cove sensu stricto. We regard the prominent Elmendorf Moraine as a readvance of latest Wisconsin age and the large body of alluvium adjacent to it as an alluvial fan of Eagle River source. Miller and Dobrovolny concur with the latter interpretation, but they and Karlstrom both correlate the Elmendorf Moraine with the entire complex of Naptowne moraines on the Kenai lowland and apply the term Naptowne to both the moraine and related deposits, a usage that we believe is no longer appropriate.

It can be inferred that the shoreline surrounding the Municipality of Anchorage on two of its three principal sides has undergone nearly continual changes throughout the Quaternary; indeed it also changes every hour of every day under the influence of the large tidal range of upper Cook Inlet. The longer-period changes were not only drastic but quite variable, as the inlet vanished from the scene, probably for several thousand years, to return again with a somewhat different configuration. The land surface, too, underwent minor change over thousands of years, then changed drastically as glaciers overrode it, only to be reformed again to something like its previous state, each time with unique details within a similar overall design. One can surmise with considerable certainty that the present shoreline and landscape will change further in relatively minor ways, usually slowly but occasionally very abruptly, until finally a glacier will come once again to scrape the scene clean and move the shore farther down inlet. Perhaps one of the future glaciers will even be the longest and thickest of all and destroy all high-level evidence of glaciation that had not already succumbed to the ravages of time. Some bits and pieces from our present scene, however, and perhaps a few more sizable remnants of deposits as well (maybe even some of the Bootlegger Cove shell horizon), could survive to become a somewhat more permanent part of a future subsurface.

Acknowledgements

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