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Chapter 4

Geology of Seward Peninsula and Saint Lawrence Island

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SEWARD PENINSULA: THE SEWARD AND YORK TERRANES

Seward Peninsula (Fig. 1) may be divided into two geologic terranes (Fig. 2) on the basis of stratigraphy, structure, and metamorphic history. The Seward terrane, an area 150 by 150 km in the central and eastern peninsula, is dominated by Precambrian(?) and early Paleozoic blueschist-, greenschist-, and amphibolite-facies schist and marble, and intruded by three suites of granitic rocks. The York terrane, roughly 100 by 75 km, occupies western Seward Peninsula and the Bering Straits region; it is composed of Ordovician, Silurian, Devonian, Mississippian, and possibly older limestone, argillaceous limestone, dolostone, and phyllite, which are cut by a suite of Late Cretaceous tin-bearing granites. The boundary between the Seward and York terranes is poorly exposed but is thought to be a major thrust fault because of its sinuous map trace, a discontinuity in metamorphic grade, and differences in stratigraphy across the boundary (Travis Hudson, oral communication, 1984). The boundary between the Seward terrane and the Yukon-Koyukuk province to the east is complicated by vertical faults (the Kugruk fault zone of Sainsbury, 1974) and obscured by Cretaceous and Tertiary cover.

The Seward Peninsula heretofore was thought to consist largely of rocks of Precambrian age (Sainsbury, 1972, 1974, 1975; Hudson, 1977). Microfossil data, however, indicate that many of the rocks considered to be Precambrian are early Paleozoic in age (Till and others, 1986; Dumoulin and Harris, 1984; Dumoulin and Till, 1985; Till and others, 1983; Vandervoort, 1985). It is likely that Precambrian rocks are a minor part of the stratigraphy of the Seward Peninsula.

Previous work

Regional mapping on Seward Peninsula began at the time of the Nome gold rush, with the work of Brooks, Richardson, Collier, and Mendenhall (1901), followed by the more detailed work of Knopf (1908), Smith (1910), Moffit (1913), and Steidtmann and Cathcart (1922). More recent mapping was initiated in the 1950s by Sainsbury (1969, 1972), who focused first on the stratigraphy and economic geology of the York Mountains in western Seward Peninsula and later on the reconnaissance-scale (1:250,000) mapping of the metamorphic rocks of the central peninsula (Sainsbury and others, 1972a, b; Sainsbury, 1974). Miller and others (1972) produced a preliminary geologic map of the southeastern peninsula at the same scale. Regional compilations have been made at 1:1,100,000 by Hudson (1977) and 1:500,000 by Robinson and Stevens (1984).

Detailed studies of the regionally extensive metamorphic rocks of the Seward terrane began with the description of blueschist-facies mineral assemblages of the Nome Group by Sainsbury and others (1970). Till and others (1986) formulated the stratigraphy of the metamorphic rocks of the Nome Group; and Thurston (1985), Forbes and others (1984), Evans and Patrick (1987), Patrick (1988), and Patrick and Evans (1989) conducted detailed mapping and petrogenetic studies of the blueschists. Detailed studies of the high-grade metamorphic rocks of the Kigluaik Mountains were undertaken by Till (1980), Sturnick (1984), and Patrick and Lieberman (1987).

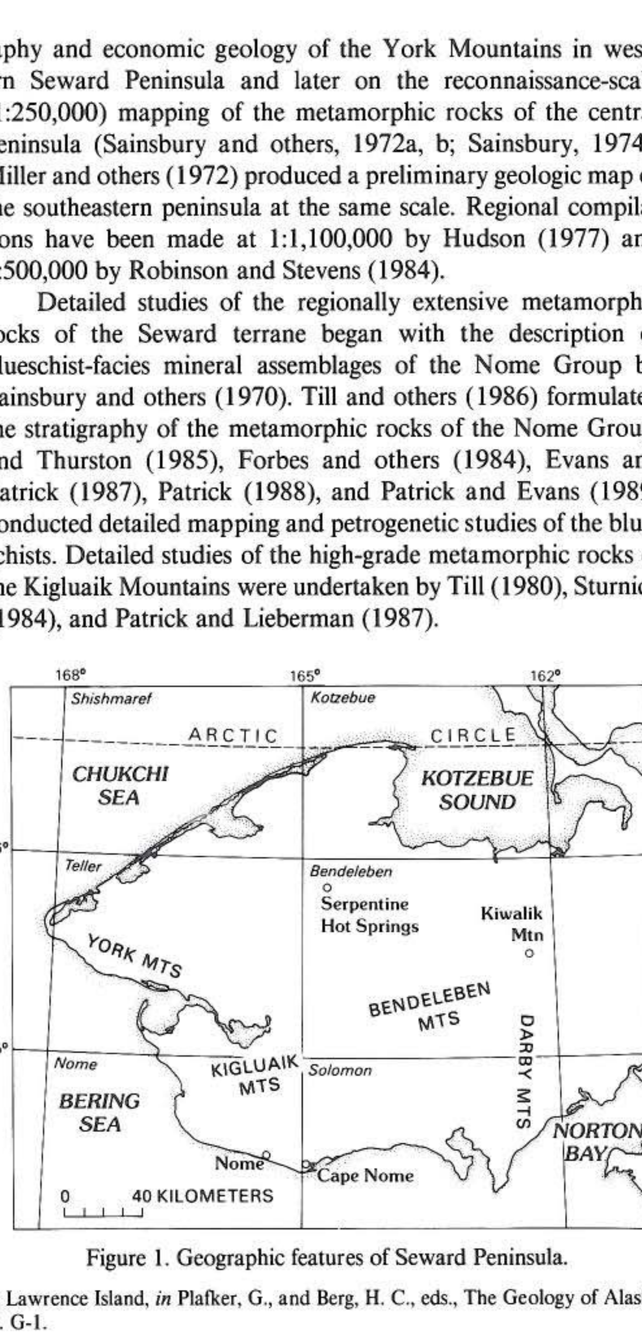

Figure 1. Geographic features of Seward Peninsula.

Till, A. B., and Dumoulin, J. A., 1994, Geology of Seward Peninsula and Saint Lawrence Island, *in* Plafker, G., and Berg, H. C., eds., The Geology of Alaska: Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-1.

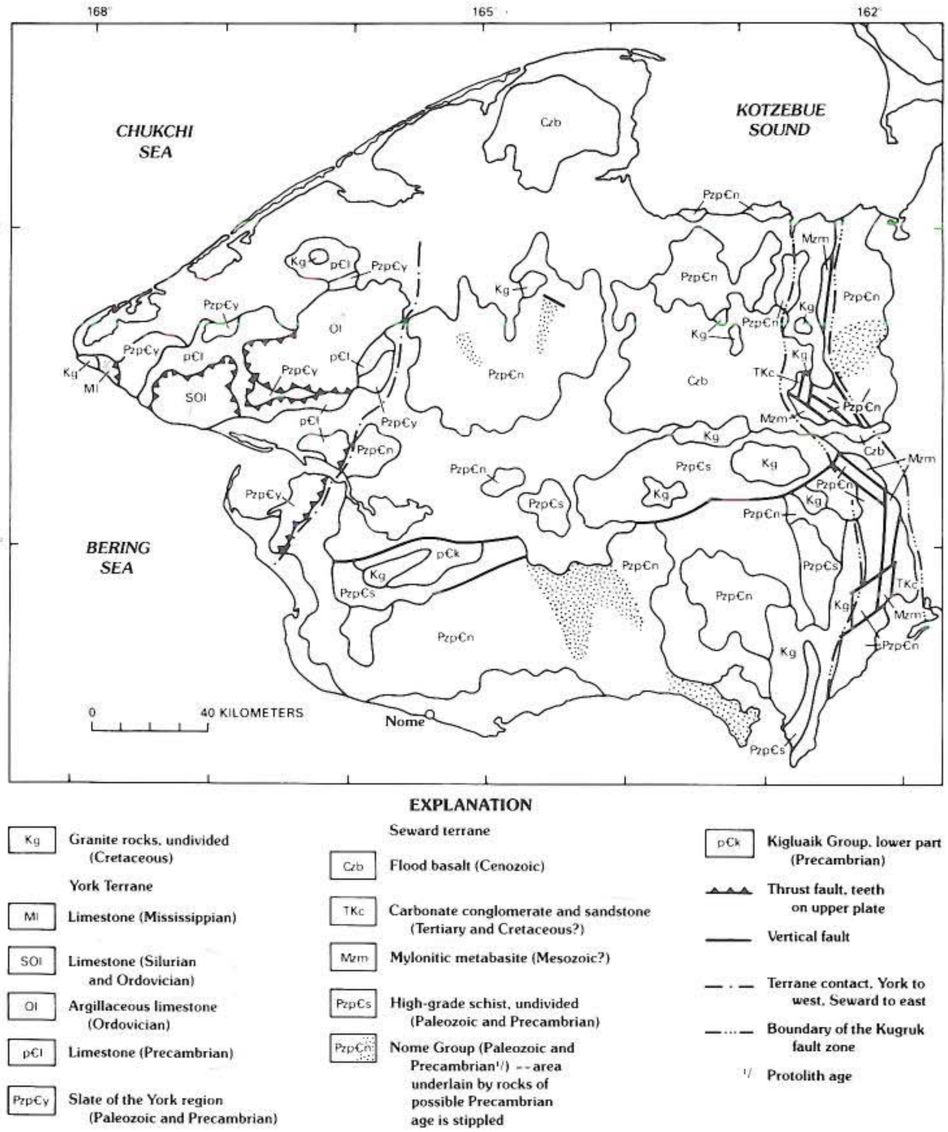

Figure 2. Simplified geologic map of Seward Peninsula, from Sainsbury (1972), Hudson (1977), and Till and others (1986).

Miller (1972) and Miller and Bunker (1976) conducted regional studies of the Cretaceous granitic rocks, and Hudson (1979) and Hudson and Arth (1983) studied geochemistry of the tin-granite suite in detail.

Tertiary deposits occur locally in the Seward terrane. Although they are poorly exposed, they have been studied in conjunction with flood basalts (Hopkins, 1963), coal deposits (Barnes, 1967), and uranium mineralization (Dickinson, 1984). Geothermal resources (hot springs), largely developed on plutonic contacts (Miller and others, 1975), have also drawn attention to Tertiary basins (Lockhart, 1984; Westcott and Turner, 1981). Studies of Quaternary deposits resulted from interest in the Bering land bridge and placer gold (Hopkins, 1967; Kaufman and Hopkins, 1986).

The Seward Peninsula is rich in mineral resources (Hudson and DeYoung, 1978; Puchner, 1986). Its gold deposits are well known, and it has substantial resources of tin and uranium. Numerous small base-metal vein deposits occur, and a stratabound Zn-Pb deposit has been postulated in the northern part of the peninsula (J. Briskey, written communication, 1985).

SEWARD TERRANE

The Seward terrane is dominated by low- and high-grade metamorphic rocks and at least three suites of granitic rocks (Fig. 2). Most common of the metamorphic rocks are the blueschist-facies schists of the Nome Group, which form the low rolling hills of the central peninsula. High-grade metamorphic rocks include the Kigluaik Group of the Kigluaik Mountains, large bodies of undivided schist and gneiss, and migmatite. These rocks, together with granitic bodies, are exposed in fault-bounded mountain ranges that transect the low-grade rocks from east to west (the Kigluaik and Bendeleben Ranges) and north to south (the Darby Range). Tertiary and Holocene basins filled with sedimentary and basaltic volcanic rocks are developed in the central and northern Seward terrane. Small amounts of basalt and coal-bearing sedimentary rocks occur in the eastern part of the terrane, commonly near the Kugruk fault zone. Within this north-south–trending fault zone, blocks of mylonitic metabasite, serpentinite, and Tertiary or Cretaceous(?) carbonate-clast conglomerate are juxtaposed with rocks of the Nome Group (Fig. 2).

Nome Group

The Nome Group, as defined by Moffit (1913), included two schist units with an interlayer of limestone. In his latest work, Sainsbury (1974) included all Nome Group rocks in the "slate of the York region." Recent mapping shows that the Nome Group is a metamorphic unit that includes two parts: (1) a coherent, mappable metamorphic stratigraphy; and (2) carbonate rocks, which have an indeterminate premetamorphic relation to that stratigraphy. Both parts are composed of early Paleozoic and possibly older protoliths that have undergone an episode of blueschist-facies metamorphism and deformation in the Jurassic (Forbes and others, 1984; Armstrong and others, 1986). the Nome Group, as used here, includes all lithologies that underwent blueschist-facies metamorphism and accompanying deformation. The Nome Group and related fossil data are discussed by Till and others (1986).

The stratigraphy of the metamorphic rocks includes four units mappable at a scale of 1:63,360: (1) a basal quartz-rich pelitic schist unit, (2) a mixed unit dominated by interlayered marble and quartz-graphite schist, (3) a mafic schist with calcareous components, and (4) an impure chlorite marble. These four units have a minimum combined thickness of 4.5 km; the schists share a common foliation and structural style and contain mineral assemblages that are stable under blueschist-facies conditions.

The pelitic schist forms tors several meters high and displays characteristic centimeter-thick layers and lenses of quartz that define at least two sets of isoclinal folds; the base is not exposed. These rocks were mapped as "tectonically metamorphosed York slate" by Sainsbury (1974).

The mixed unit is defined in part by its position between the more distinctive pelitic and mafic schist units, and also by the presence of quartz-graphite schist interlayered on a scale of 10 to 150 m with pure and impure marble and subordinate, thinner layers of pelitic and mafic schist. The quartz-graphite schist and marble thicken and thin significantly along strike on a scale of 5 to 10 km; pure marble dominates some exposure of the unit; quartz-graphite schist dominates others. The quartz-graphite schist of the mixed unit is equivalent to some of the rocks included by Sainsbury (1969) in the "slates of the York region." Mafic schist intercalated within the mixed unit is compositionally similar to parts of the overlying mafic schist unit. Metabasite boudins identical to those found in the mafic schist occur locally near the top of the mixed unit.

The mafic schist unit is composed of massive metabasite, mafic schist, calc-schist, chlorite-albite schist, and minor marble and quartzite. Massive, coarse-grained metabasite lenses and layers are conformably intercalated with schists of similar composition, as well as with quartz-poor calc-schist, pelite, semipelite and chlorite-albite schist. Glaucophane has been found in all major lithologies except pure marble and quartzite (Thurston, 1985). Repetitive interlayers of calc-schist and mafic schist occur locally in the unit. Two compositionally distinct types of metabasite have been recognized in the mafic schist: one exceptionally rich in iron and titanium, with abundant glaucophane and garnet; and one relatively rich in magnesium and aluminum, with actinolite and rare garnet (Forbes and others, 1984). The mafic schist unit is equivalent to the Casadepaga Schist of Smith (1910).

The impure chlorite marble directly overlies the mafic schist and consists of an orange-weathering, foliated marble, which characteristically includes lenses and thin layers of chlorite-albite schist. Blue amphibole forms inclusions in albite. Boudins of metabasite near the base of the impure marble are similar to the magnesium- and aluminum-rich metabasites in the mafic schist.

The presence of mafic material in the mixed unit, the mafic schist unit, and the impure chlorite marble suggests that the three lithologic units were part of the same sequence during the period of mafic igneous activity. Boudins in the mixed unit may represent feeder dikes to the volcanic protolith of the mafic schist, and boudins and lenses of mafic material at the base of the impure chlorite marble may have been produced at the waning stages of igneous activity.

Conodonts and radiolarians have been recovered from two of the upper three units of the metamorphic stratigraphy (Till and others, 1986). Radiolarians were observed in thin sections of finely laminated quartz-graphite schist from the mixed unit in the northern Darby Mountains. The age of the radiolarians has not been determined, but their presence indicates a Cambrian or younger age for that unit. Ordovician conodonts were found in marble near the top of the mixed unit; late Early to Middle Ordovician conodonts were recovered from the impure chlorite marble. Because these two units bracket the mafic schist, most of the metamorphic stratigraphy is Ordovician in age. The mafic igneous activity recorded in the mafic schist is therefore pre-late Early to Middle Ordovician in age. The mixed unit and mafic schist were intruded at Kiwalik Mountain by a granitic orthogneiss dated at 381 ± 2 Ma (J. Aleinikoff, written communication, 1983).

The original nature or age of the contact between the pelitic schist unit and the overlying unit is not known. No fossil or isotopic data constrain the age of the protolith of the pelitic schist unit. Gardner and Hudson (1984) indicate a Precambrian age for the unit on the basis of a structural analysis that suggested structures in the pelitic schist recorded at least one more deformational event than other units in the Nome Group. The pelitic schist unit may have been basement for Cambrian and Ordovician Nome Group rocks, or may have been juxtaposed with those rocks during the Jurassic.

The second part of the Nome Group, carbonate rocks with unknown relations to the premetamorphic stratigraphy, was divided by Dumoulin and Harris (1984) into several units on the basis of lithology and age. These rocks occur as isolated exposures and as fault slices; rarely, they are folded into the metamorphic stratigraphy. Till and others (1986) and Harris and Repetski (1987) provide fossil data and detailed lithologic descriptions of these rocks.

Cambrian, Ordovician, and Silurian dolostones, with sedimentary structures and fauna indicating shallow-water deposition, overlap in age with the metamorphic stratigraphy (Till and others, 1986); they may have formed in the same depositional basin. Deeper-water carbonate rocks, with sedimentary structures indicative of slope or basin environments, range in age from Cambrian to Devonian (Till and others, 1986; Ryherd and Paris, 1987).

Spatial relations suggest that Devonian rocks with sedimentary structures and fauna indicating shallow-water deposition may have been unconformably deposited on older Nome Group rocks (specifically the metamorphic stratigraphy and the deeper water Cambrian to Devonian carbonate rocks).

These lower Paleozoic units, along with carbonate rocks of unknown age, are included in the Nome Group on the basis of mineral assemblages, rock texture, and/or color alteration indices of conodonts. Depositional relations with other Nome Group rocks cannot be established, however, and they may not have formed in the same depositional basin as the carbonates in the metamorphic stratigraphy (principally in the mixed unit).

The flat-lying to gently dipping transposition foliation of the Nome Group is characterized by an axial planar schistosity, which is commonly parallel to lithologic layering and abundant intrafolial isoclinal folds. This foliation is the product of penetrative ductile deformation that presumably has significantly altered original geometric relations between lithostratigraphic units while leaving the stratigraphic succession largely intact. Lithologic contacts have been rotated parallel to the foliation in most exposures. Locally the transposition foliation crosses lithologic layering at high angles, presumably in the vicinity of the hinges of outcrop- to map-scale recumbent folds. In the northeast Seward Peninsula, in the vicinity of Kiwalik Mountain, the metamorphic stratigraphy is inverted relative to most exposures of the Nome Group; this may indicate that map-scale fold formation accompanied deformation. Stretching lineations (mineral aggregates, boudin axes) and isoclinal fold hinges that formed with the schistosity both have a north-south trend. All of the aforementioned structures were formed during regional blueschist-facies metamorphism, and indicate that the Nome Group schists were at a sufficient depth in the crust to undergo ductile deformation accompanied by significant extension along a north-south trend. Patrick (1988) used quartz petrofabrics and the geometry of rare sheath folds to determine that the deformation had northward vergence.

The metamorphic event occurred after the intrusion of the Kiwalik Mountain orthogneiss (381 ± 2 Ma) and before contact metamorphism of Nome Group rocks by the Kachauik and Windy Creek Plutons (100 to 110 Ma; Miller and others, 1972); that is, between the Late Devonian and middle Cretaceous. Armstrong and others (1986) obtained ages for the blueschist event ranging from 170 to 100 Ma using whole-rock-mica Rb-Sr isochrons. The same study obtained K-Ar mineral ages from rocks of the Nome Group ranging from 122 to 194 Ma, and suggested that a Jurassic metamorphic event was followed by partial resetting of the Rb-Sr isotopic system.

The timing of tectonic events studied in adjacent areas in northwestern Alaska probably can be related to this blueschist metamorphism event on the Seward Peninsula. In the western Brooks Range, sedimentation related to the Brooks Range orogeny began in the Middle Jurassic (Mayfield and others, 1983). Blueschist-facies rocks of the "schist belt," in the hinterland of the Brooks Range, have been related to the early phases of that orogenic episode (Gilbert and others, 1977; Hitzman and others, 1986). The predominantly lower Paleozoic platformal protolith sequence of the schist belt was metamorphosed in the Late Jurassic or Early Cretaceous (Armstrong and others, 1986). The similarities in protolith composition and metamorphic conditions of schist belt and Nome Group rocks have led to correlation of the two metamorphic terranes and the supposition that they formed in similar tectonic environments (Forbes and others, 1984; Patrick, 1988).

Kigluaik Group

Gneisses of the Kigluaik Group are the deepest crustal rocks exposed in northwestern Alaska and may be among the oldest. Moffit (1913) included in the group all schist and gneiss in the Kigluaik Mountains above greenschist facies in grade. These rocks can be divided into structurally (1) an upper section, amphibolite facies schists of the south flank of the Kigluaik Range (the Tigaraha Schist of Moffit); and (2) a lower section, granulite-facies gneiss of the Mt. Osborn area (the "heavily bedded limestone and basal gneiss" of Moffit). The upper section of the Kigluaik Group may be correlative with amphibolite-facies schists of the Bendeleben and Darby Ranges; no rocks correlative with the lower section of the group are known on the Seward Peninsula.

The upper section of the Kigluaik Group is composed of pelitic and calcareous schist, marble, and subordinate amphibolite and quartz-graphite schist. It is at least 3.5 km thick and locally includes foliated and nonfoliated granitic plutons. A massive pelitic and semipelitic biotite gneiss is the lowest lithologic unit in the upper section. It equilibrated above the second sillimanite isograd, and contains sillimanite and K-feldspar. The biotite gneiss is the only unit in the group from which an Rb-Sr age has been obtained (see below). Rocks from the upper part of the Kigluaik Group may be thermally upgraded equivalents of the Nome Group (Till, 1980; Thurston, 1985; Patrick and Lieberman, 1987). The contact between the Kigluaik and Nome Groups is commonly faulted, but may be a gradational metamorphic contact in the southwestern Kigluaik Mountains. Lithologic similarities and static metamorphic textures also provide possible evidence for this conclusion. High-grade equivalents of Nome Group units have been mapped in the Bedeleben and Darby Mountains (Till and others, 1986); these rocks may be correlative to the upper section of the Kigluaik Group.

The lower section of the Kigluaik Group consists of pure and impure marble, with interlayers of gneiss, mixed gneiss of pelitic, quartzo-feldspathic, mafic and ultramafic composition, and migmatite. Coarsely crystalline marble in the steepest faces of the cirques around Mt. Osborn ranges in composition from pure calcite marble to impure dolomitic marble containing olivine and diopside. Discontinuous layers and megaboudins of gneiss of varied composition are found below the marble-dominated section. Granulite-facies assemblages are common in these rocks, with two pyroxene-bearing assemblages in semipelitic and mafic gneiss and spinel peridotite in the ultramafic boudins. Mafic and ultramafic gneisses are most common at the base of the exposures and are interlayered with metasedimentary rocks. The only known alpine garnet lherzolite in North America occurs in these exposures and is partially recrystallized to spinel lherzolite. The migmatite is best exposed in a large cirque directly west of Mt. Osborn where it occupies the base of the exposed section of the Kigluaik Group. The upper contact of the migmatite locally cuts the foliation in the overlying gneiss. Large rafts of metasediment and peridotite derived from the mixed gneiss float in pegmatite and foliated granite.

The dominant structure in the Kigluaik Group is a transposition foliation parallel to lithologic layering. In the upper part of the Kigluaik Group, this is a schistosity; in the lower part of the group it is gneissic layering. Isoclinal introfolial folds with wavelengths of a few millimeters to tens of meters are present throughout the Kigluaik Group. Isoclinal fold hinges and stretching lineations are parallel and oriented along a north-south trend (Till, 1980). This foliation has been gently folded about an east-west axis to create a doubly plunging antiform with a culmination in the vicinity of Mt. Osborn.

Samples of pelitic and semipelitic gneiss from the biotite gneiss in the upper part of the Kigluaik Group yielded a whole-rock Rb/Sr isochron age of 735 Ma (Bunker and others, 1979). This age has been interpreted as the age of metamorphism for the Kigluaik Group; no confirmation of this interpretation has been obtained using mineral ages. This may represent the age of the protolith. The age of the migmatite is unknown. A geological argument can be made in favor of a Late Jurassic or Early Cretaceous age for the granulite-facies metamorphism (see "Tectonic history").

Biotite and hornblende separated from the upper Kigluaik Group have yielded K-Ar ages of 81 to 84 Ma (Westcott and Turner, 1981). Muscovite from pegmatite, which cuts the metamorphic fabric, yielded a K-Ar age of 84.0 ± 1.2 Ma (N. Shew, written communication, 1988). These ages represent minimum limits on the age of high-grade metamorphism.

High-grade metamorphic rocks of the Bendeleben and Darby Mountains

Amphibolite-, locally granulite-, and greenschist-facies rocks crop out in the Bendeleben and Darby Mountains. Lithologies include metapelite, marble, quartz-graphite schist, calc-schist, quartzo-feldspathic schist, amphibolite, and orthoamphibole-cordierite schist. These rocks may be in part correlative with the upper section of the Kigluaik Group; schist in the northern Darby Mountains is correlative with the Nome Group. Rocks in both mountain ranges have undergone both high- and low-pressure metamorphism.

In the eastern Bendeleben Mountains, sillimanite- and K-feldspar-bearing pelitic and quartzo-feldspathic schists occur in a large envelope of migmatite surrounding the Bendeleben Pluton; these schists record only one metamorphic event. In contrast, pelitic rocks in the western Bendeleben Mountains include an early high-pressure assemblage with kyanite and staurolite and late low-pressure assemblages with andalusite, sillimanite, and locally cordierite.

Regional mapping in the northern Darby Mountains has shown that Nome Group units can be recognized that have been upgraded from blueschist to greenschist facies. These units can be traced southward into the central Darby Range; the metamorphic grade of the rocks increases southward and reaches upper amphibolite or granulite facies in the vicinity of Mt. Arathlatuluk, where small stocks of anatectic granite crop out. A narrow contact aureole of andalusite-bearing schist is present around the Darby Pluton (95 Ma; Miller and Bunker, 1976) and overprints the upper amphibolite-facies metamorphism.

Protolithic ages of metasedimentary rocks in the Bendeleben and Darby Mountains are not known. However, where upgraded equivalents of Nome Group lithologies can be recognized, the protoliths are probably pre–Late Ordovician.

Biotite from an anatectic granitic stock in the southern Darby Mountains yields a K-Ar age of 102.6 ± 1.4 Ma (N. Shew, written communication, 1988), which is the minimum age limit for high-temperature metamorphism in the range. Geologic relations are consistent with this age, because relicts of the Late Jurassic to Early Cretaceous blueschist event were overprinted by the high-temperature event, and intrusion of the 95-Ma Darby Pluton occurred after the high-temperature metamorphism.

Ages of metamorphic events are difficult to discern in the Bendeleben Mountains. K-Ar mineral ages obtained from schist in the western Bendeleben Mountains and from the two plutons in the range fall in the range of 80 to 90 Ma (Westcott and Turner, 1981; Miller and Bunker, 1976). These ages record synchronous cooling of the intrusive bodies, their migmatitic carapaces, and the upper amphibolite-facies country rock.

Cretaceous magmatism

Large granitic plutons are associated with the high-grade metamorphic rocks of the Kigluaik, Bendeleben, and Darby Mountains (Fig. 2). These plutons range in composition from granodiorite to biotite granite but also include syenite and monzonite (Miller and Bunker, 1976; Miller, this volume, Chapter 16; Hudson, this volume); K-Ar ages of these plutons range from 100 to 80 Ma. Several alkaline subsilicic complexes and dike swarms occur in the southeastern Seward Peninsula; these are part of a belt of similar rocks of middle Cretaceous age, which extends some 1,200 km across western Alaska, the Bering Sea islands, and into eastern Siberia (Miller, 1972, 1989). Small late Cretaceous plutons and stocks of anorogenic origin are scattered across the northwestern Seward Peninsula and intrude both the York and Seward terranes. They consist of biotite granite and typically have tin deposits associated with them (Hudson and Arth, 1983).

Cenozoic sedimentary and basaltic rocks

Tertiary and Quaternary sedimentary rocks and basalt flows fill local basins and form large plateaus, respectively, on the Seward terrane. Sedimentary basins, tens of kilometers in diameter, locally may be very deep (Barnes and Hudson, 1977). Coal and uranium deposits occur in one or more of these basins (Dickinson, 1984). Geothermal activity at Pilgrim Springs, north of the Kigluaik Range, may be related to basin-forming faults (Westcott and Turner, 1981). Many of the Tertiary sedimentary rocks show signs of deformation, especially in the eastern part of the Seward terrane (Kugruk fault zone), where postdepositional deformation rotated coal-bearing strata to near-vertical dip. The maximum age of the strata is unknown, but pollen as old as Eocene have been collected from siltstone (T. Ager, written communication, 1985).

Alkalic and tholeiitic basalts form a plateau 46 by 25 km and an extensive maar and flow field in northern Seward Peninsula. Small flows occur in southern parts of the peninsula. Radiometric data indicate that flows as old as 29 Ma are present; the most recent flows are probably younger than 2,000 B.P. (Westcott and Turner, 1981; Hopkins, 1967, and unpublished data; Moll-Stalcup, this volume).

The Kugruk fault zone

A complex of north-south–trending vertical faults in the east part of the Seward terrane defines the Kugruk fault zone (Sainsbury, 1974; Fig. 2). Nome Group carbonate rocks from the most common slices in the fault zone. Blueschist-facies mylonitic metabasite, Tertiary or Cretaceous(?) carbonate- and mafic-clast conglomerate, altered tonalite, and rare serpentinite from mappable units in the zone. The presence of glaucophane and lawsonite in the mylonitic metabasite indicates that the zone may mark a significant tectonic boundary (Till, unpublished data). The carbonate- and mafic-clast conglomerate is a ridge-forming unit in the fault zone. It consists of material derived form bedrock units in the fault zone (principally the mylonitic metabasite and Paleozoic carbonate lithologies) but also contains material without analogues in the fault zone or in the surrounding Nome Group. The elongate Cretaceous Darby Pluton parallels the trace of the southern part of the fault zone, and it is possible that this is a zone of crustal weakness active from Cretaceous time to the present. Rocks of the Nome Group have been identified on both sides of the fault zone, so it cannot constitute the boundary between the Seward terrane and the Yukon-Koyukuk province (Dumoulin and Till, 1985).

YORK TERRANE

The York terrane consists mainly of Ordovician through Mississippian limestone, argillaceous limestone, dolostone, and fine-grained clastic rocks deposited in a shallow-water, normal marine to slightly restricted platform environment. Carbonate rocks of Ordovician age compose most of this sequence; some rocks are of unknown age and may be pre-Ordovician. York terrane carbonate rocks, especially those of Ordovician age, show many similarities in lithofacies and biofacies to the metacarbonate sequence of the western Baird Mountains Quadrangle in the southwestern Brooks Range (Dumoulin and Harris, 1987; Harris and Repetski, 1987). The color alteration indices (CAI) of most conodonts found in the York terrane range from 1.5 to 4.5 indicating that host rocks reached minimum temperatures of 60 to 300° C; locally higher values occur in rocks adjacent to plutonic intrusions (A. G. Harris, oral communication, 1985). Rocks of the York terrane have been deformed by thrust faulting, probably during the Brooks Range orogeny. Small stocks of 70- to 80-Ma biotite granite and mafic igneous rocks of unknown age locally intrude the sedimentary rocks. Tin deposits are associated with the granites; deposits at Lost River constitute the largest tin lode reserves in the United States (B. L. Reed, oral communication, 1985).

Ordovician rocks

Lower, Middle, and Upper Ordovician rocks occur in the York terrane. Lower Ordovician rocks are the most abundant and consist of a "shallow-water facies" and a "quiet-water facies" (Sainsbury, 1969). Rocks of the shallow water facies are interbedded quartz-carbonate sandstone and dolomitic lime mudstone, wackestone, and packstone. The sandstone is planar to cross-bedded, with locally well-developed oscillation and current ripples. The mudstones and wackestones are bioturbated, with bedding-plane feeding trails and subvertical burrows (Dumoulin, unpublished data). Other sedimentary features include intraclast conglomerates and oolites. This sequence may be as thick as 1,500 m (Sainsbury, 1969).

The sparse megafossil assemblage includes stromatolites, brachiopods, gastropods, and trilobite fragments (Sainsbury, 1969). Conodonts from a measured section at the mouth of Koteebue Creek indicate warm, shallow-water, normal marine conditions, and are of early Early Ordovician age (low Fauna D ofthe North American midcontinent faunal succession); they are very similar to correlative faunas from the Medfra Quadrangle, about 700 km southeast (A. G. Harris and J. Repetski, written communication, 1985).

Some rocks, lithologically similar to the Lower Ordovician shallow-water facies but considered by Sainsbury (1972) to be of Precambrian age, have recently been found to contain Early Ordovician conodonts. At one locality, bioclastic wackestones with 5 to 10 percent thin flasers of dolomitic mud are found; these rocks contain conodonts of early Early Ordovician age (low Fauna D) and indicate deposition in warm, shallow water (J. Repetski, written communication, 1985). Vandervoort (1985) reports additional collections of Early Ordovician conodonts from units previously thought to be Precambrian.

The other Lower Ordovician facies distinguished by Sainsbury (1969) consists of thick-bedded to massive micritic limestone with subordinate thinner interbeds of argillaceous limestone and locally abundant chert. As in the shallow-water facies, trace fossils are abundant. Unlike the shallow-water facies, these rocks lack ripple marks and sandstone interbeds, suggesting that deposition may have occurred in a quieter, perhaps somewhat deeper water environment with little input of terrigenous material (Sainsbury, 1969). Common rock types include bioclastic wackestones and pellet and intraclast grainstones (Dumoulin, unpublished data). The entire quiet-water sequence is thought to be about 2,200 m thick; the upper 70 m of this unit consists of a distinctive white to pinkish gray, blue-gray weathering, evenly bedded limestone with abundant trilobite fragments (Sainsbury, 1969).

Megafossils from this unit include cephalopods (Flower, 1941, 1946), trilobites, gastropods, brachiopods, and ostracodes (Sainsbury, 1969). These fossils indicate a Tremadocian or early Arenigian age for this unit (Dutro in Ross and others, 1982).

Conodonts collected from the quiet-water sequence support this age assignment. Collections from the middle and lower parts of the sequence contain conodonts of early Arenigian age (Fauna D) and indicate deposition on a shelf or platform chiefly under normal marine, although somewhat warm-water conditions. Collections from beds higher in the sequence contain conodonts of early middle Arenigian age diagnostic of slightly cooler water conditions (A. G. Harris and J. Repetski, written communication, 1985).

The entire Lower Ordovician sequence is characterized by 8- to 15-m-thick, shallowing-upward cycles (Vandervoort, 1985). The cycles appear to represent deposition in a range of subtidal to supratidal environments on a humid, pericontinental shelf.

Middle Ordovician outcrops are more limited in extent than those of the Lower Ordovician. The top of the quiet-water facies of the Lower Ordovician is succeeded, apparently conformably, by 7 to 30 m of fissile black shale with minor interbedded black limestone and dolomitic limestone. The shale sequence grades upward into flaggy, thin-bedded black limestone with shale partings. The Early/Middle Ordovician boundary seems to lie within this shale and limestone unit (Dutro in Ross and others, 1982). In the western part of the terrane, medium-bedded gray and brown limestone lithologically similar to older rocks also contains a Middle Ordovician fauna (Sainsbury, 1969; Dutro in Ross and others, 1982). Sainsbury (1972) considered the Middle Ordovician sequence to exceed 800 m in thickness.

Megafossils in Middle Ordovician rocks include cephalopods, graptolites, gastropods, and trilobites (Sainsbury, 1969; Dutro in Ross and others, 1982). Conodonts collected from the lower 50 m of the shale and limestone unit are of latest Arenigian to earliest Llanvirnian (earliest Middle Ordovician) age, and indicate cooler and deeper water conditions than those prevalent during deposition of the upper quiet water sequence (A. G. Harris and J. Repetski, written communication, 1985).

Upper Ordovician rocks crop out primarily along the Don River and are medium- to thick-bedded dark-gray limestone and dolomitic limestone, locally very fossiliferous. These rocks are wackestones and packstones, with pellets and skeletal hash in the micrite matrix (Dumoulin, unpublished data).

Megafossils from this unit include corals (Oliver and others, 1975), stromatoporoids, gastropods, ostracodes, trilobites, and brachiopods (Sainsbury, 1969). The trilobite fauna includes *Monorakis,* the only confirmed occurrence of this form outside of the northern Soviet Union, where it is widely distributed (Ormiston and Ross, 1979). Other evidence for a faunal tie between the Soviet Union and the area of the York terrane during the late Ordovician is provided by the coral and brachiopod faunas (Potter, 1984) and by conodont faunas from the Don River section. The conodont assemblage is of Late Ordovician age and is diagnostic of a warm, relatively shallow-water depositional environment; it includes forms thus far found only in Siberia and Alaska (A. G. Harris, written communication, 1988).

Silurian rocks

Silurian rocks are best exposed along the Don River where the Ordovician/Silurian boundary has been studied (Sainsbury and others, 1971). Light brown to dark gray limestones, dolomitic limestones, and dolostones contain local buildups of corals and stromatoporoids. The Silurian section consists chiefly of mudstones and bioclastic packstones and wackestones and contains more dolostones than does the underlying Ordovician section Some samples show well-developed fenestral fabric, suggesting deposition took place in very shallow water (Dumoulin, unpublished data). The Don River Silurian section may be as thick as 270 m (Sainsbury and others, 1971).

Megafossils from the Silurian portion of the Don River section include corals (Oliver and others, 1975), stromatoporoids, bryozoa, and brachiopods, interpreted to be of Middle and Late Silurian age by Sainsbury and others (1971). Biostratigraphically diagnostic conodonts from this locality are of Early and earliest Late Silurian age and indicate deposition in a warm, shallow-water environment (A. G. Harris, written communication, 1985).

Rocks of Silurian age also form a small klippe in the eastern part of the York terrane; these rocks were originally thought to be of Devonian age (Sainsbury, 1972). Megafossils at this locality indicate a Silurian age (W. Oliver, written communication, 1973). Conodonts from this locality are of middle Middle Silurian age and indicate deposition in a shallow-water environment (A. G. Harris, written communication, 1985).

Devonian? rocks

Only a single locality in the York terrane has produced fossils of possible Devonian age. Rocks in the north-central Teller quadrangle, originally shown by Sainsbury (1972) as part of the Lower Ordovician quiet-water facies, contain corals considered to be of probable Middle or early Late Devonian age by Oliver and others (1975). This age has been reevaluated, based on more recent worldwide collections, as Late Silurian (late Ludlovian)–early Late Devonian (Frasnian; A. Pedder, personal communication, 1988).

Mississippian rocks

Rocks of Mississippian age found near Cape Mountain, on the westernmost tip of the Seward Peninsula, consist of intensely deformed and recrystallized limestone intercalated with subordinate clastic rocks; the section is too deformed to allow an estimate of its thickness (Steidtmann and Cathcart, 1922; Sainsbury, 1972). A coral fauna of probable Late Mississippian age has been obtained from these rocks (Steidtmann and Cathcart, 1922).

Rocks of uncertain age

Two important units of uncertain age are found in the York terrane. There are extensive outcrops of slate, siltstone, graywacke, and subordinate carbonaceous limestone of the "slate of the York region" (Sainsbury, 1974) in the west, central, and south parts of the York terrane. This unit was thought by Sainsbury (1969, 1972) to be transitional upward into a locally strongly deformed, thin-bedded sequence of argillaceous and dolomitic limestone interbedded with silty claystone, called the Kanauguk Formation by Sainsbury (1974). Both the "slate of the York region" and the Kanauguk Formation were considered to be of Precambrian age because "the rocks are unfossiliferous and more deformed than the overlying Lower Ordovician limestones, contain quartz veins, and are intruded by numerous gabbros that do not intrude the Lower Ordovician limestones" (Sainsbury, 1972, p. 2).

The "slate of the York region" in the York terrane includes diverse lithologies, which may be grouped roughly into two sequences (Dumoulin and Till, unpublished data). One sequence consists of fine-grained schists and semischists of various, primarily pelitic and calcareous, composition; no fossils have yet been found to constrain their age. The other sequence is less metamorphosed and consists of locally calcareous, rhythmically intercalated mudstone, siltstone, and sandstone; clasts are mostly quartz and sedimentary lithic grains. Outcrop features include climbing ripples, cross beds, parallel laminae, convolute laminae, and graded beds, and suggest a turbidite origin for these rocks. A single conodont collection from this sequence is of middle Early through Late Ordovician age (A. G. Harris, written communication, 1985). Conodonts recovered from turbidite sequences may represent reworked older material, so it seems prudent to interpret this as a maximum age.

Rocks of the Kanauguk Formation are lithologically similar to rocks of the shallow-water facies of the Lower Ordovician; some limestones assigned a Precambrian age by Sainsbury (1972) have since been found to contain Lower Ordovician conodonts. Thus, much or all of Sainsbury's "Precambrian limestone unit" may be of Early Ordovician age and at least part of the "slate of the York region" is no older than middle Early Ordovician.

TECTONIC HISTORY

Lower Paleozoic rocks of the Seward Peninsula record a continental platform depositional environment punctuated by at least one episode of rifting. Iron- and titanium-rich metagabbros of the Nome Group are compositionally similar to modern ridge tholeiites (Forbes and others, 1984); the metagabbros intruded volcanogenic sediments and carbonate rocks during the Ordovician. Upper Paleozoic strata are not known on Seward Peninsula.

The Mesozoic history of the Seward Peninsula is dominated by convergent tectonics and plutonism. The Nome Group blueschist terrane is among the largest structurally coherent blueschist terranes in the world (Forbes and others, 1984). It is thought to have formed in or near a gently south-dipping subduction zone during collision of a volcanic arc with part of the North American continent in Late Jurassic to Early Cretaceous time (Box, 1985; Patrick, 1988).

The significance of the garnet lherzolite at the base of the Kigluaik Group is not well understood. The rare occurrence of alpine-type garnet lherzolite in crustal settings has caused many workers to infer that they are a relict of significant crustal upheaval (i.e., continental collision; Carswell and Gibb, 1980).

Upgraded equivalents of the Nome Group stratigraphy have been mapped in the Bendeleben and Darby Mountains (Till and others, 1986). Overprinting of higher-grade minerals directly onto blueschist-facies assemblages was noted by those workers in the northern Darby Mountains and by Patrick and Lieberman (1987) in the southern Kigluaik Mountains. The high-grade metamorphism in the Darby Mountains ended by 102 Ma (the age of small anatectic stocks that intruded the highest-grade rocks); there, the high-grade thermal overprint followed blueschist-facies metamorphism within 40 to 60 m.y. Patrick and Lieberman (1987) suggest that the high-grade rocks formed with restoration of a normal thermal gradient after subduction ceased, and they cite the metamorphic history of the Lepontine Alps as an analogue to the Seward Peninsula. Three periods of Late Cretaceous plutonism followed the crustal thickening episode.

In contrast to the deep crustal history of the Seward terrane, rocks of the York terrane show evidence of brittle deformation by thrust faulting (Sainsbury, 1969, 1972 and heating to relatively low temperatures (CAI 1.5 to 4.5; A. G. Harris, written communication, 1985). Sainsbury (1972) indicated a northward vergence for thrusting in the York terrane; structures with eastward vergence have also been observed (Till, unpublished data). (Sainsbury, 1972, also cited evidence for east-vergent folding in the Seward terrane.) More extensive microfossil dating, geologic mapping, and structural analysis are necessary before the age and nature of thrusting can be determined.

Confusion regarding the vergence of thrusting may be due to the occurrence of two separate events: one, synchronous with the development of the fold and thrust belt in the Brooks Range; and the other, in response to east-west compression centered in the Bering Straits area due to opening of the Atlantic Ocean. Folds with north-south–oriented axes, slightly overturned to the east, are found just east of Seward Peninsula in Albian and older sedimentary rocks, indicating that east-west compression occurred in post-Albian time (Patton and Tailleur, 1977). Patton and Tailleur (1977) postulated that an oroclinal bend formed between western Alaska and Chukotka (the Soviet Far East) in early Tertiary time, corresponding to the opening of the Atlantic Ocean. Their model would require a 90° counterclockwise rotation of Seward Peninsula relative to the western Brooks Range. Several lines of evidence argue against this model. Fold hinges and lineations formed during the Late Jurassic or Early Cretaceous in schist of the Seward terrane are parallel to their counterparts in the Brooks Range schist belt (Till, unpublished data). Quartz petrofabrics of Nome Group schist indicate that deformation on Seward Peninsula was northward-vergent (Patrick, 1988). Paleomagnetic studies (Plumley and Reusing, 1984) have been interpreted to preclude rotation of Seward Peninsula during the Tertiary. A major left-lateral strike-slip fault system trending northwest through Kotzebue Sound may instead be the major mechanism by which the Seward Peninsula moved into its present position.

Basin development and basaltic volcanism attest to the extensional nature of tectonics of the Seward Peninsula during Tertiary time.

SAINT LAWRENCE ISLAND

Saint Lawrence Island (Fig. 3) is situated in the shallow portion of the Bering Sea underlain by continental crust, 208 km from the Seward Peninsula and 64 km from the Chukotsk Peninsula of the U.S.S.R. The island is 5,000 km^2 in area, most of which is tundra-covered wave-cut platform. Streams are incised up to 50 m into the platform, and barren, rubble-covered mountains rise 300 to 600 m above it.

The island has been mapped at a scale of 1:250,000 (Patton and Csejtey, 1980), and topical studies have been completed on Pleistocene deposits (Hopkins and others, 1972) and plutonic rocks (Csejtey and Patton, 1974; Csejtey and others, 1971).

The bedrock of Saint Lawrence Island is composed of three lithologic belts, which are knit by mid-Cretaceous intrusives and overlain by a cover of Tertiary and Quaternary volcanic rocks.

The three lithologic belts include Paleozoic miogeoclinal sequence, a Permian to Triassic oceanic assemblage, and Lower Cretaceous andesitic rocks (Patton and Csejtey, 1980). The miogeoclinal assemblage, exposed in the eastern and central part of the island, is composed of a thick section of Devonian and Mississippian carbonate rocks and a condensed section of Triassic shale, limestone, and chert. The Permian to Triassic assemblage, in probable fault contact with the miogeoclinal rocks and exposed in the western and central part of the island, is composed of intensely deformed graywacke, grit, shale, and associated gabbro and diabase. Cretaceous volcanic rocks include flows, hypabyssal intrusives, and chiefly andesitic volcaniclastic rocks that range in composition from basalt to rhyolite. The Cretaceous rocks are also intensely folded and faulted (Patton and Csejtey, 1980). The three lithologic belts and a minor amount of undated but probable Precambrian or Paleozoic banded marly limestone were intruded

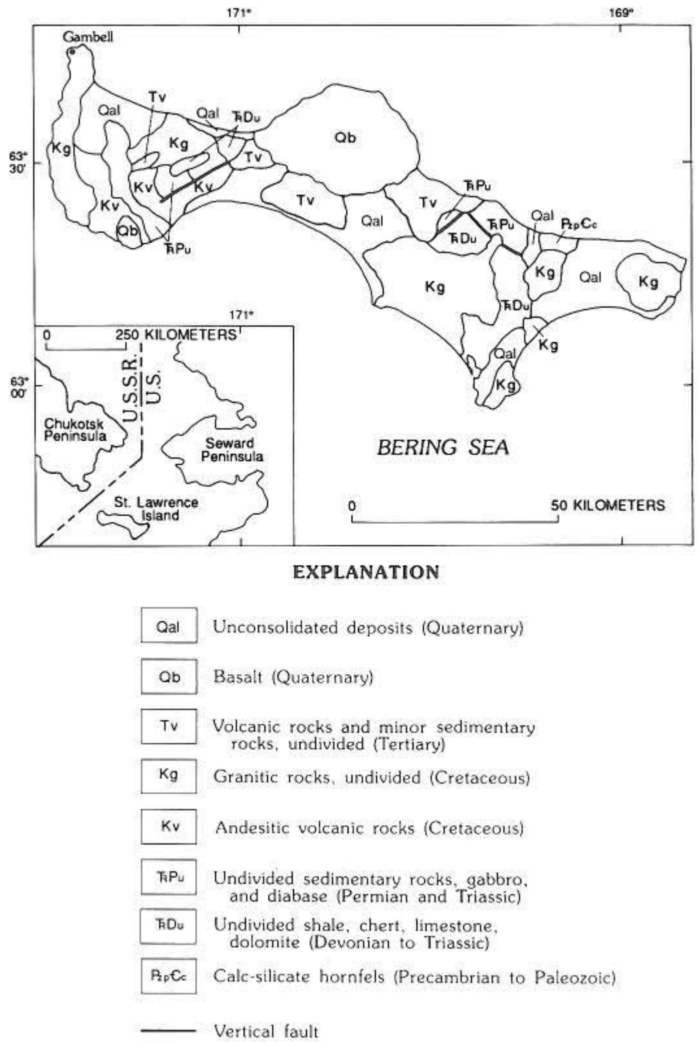

Figure 3. Geologic map of St. Lawrence Island, from Patton and Csejtey (1980).

during the Cretaceous by quartz monzonitic and lesser syenitic plutons, including small bodies of nepheline syenite that give K-Ar ages ranging from 93 to 108 Ma (Patton and Csejtey, 1980).

Tertiary volcanic rocks are of Paleocene and Oligocene age. Paleocene rocks are composed of sodarhyolite and basalt flows and hypabyssal intrusives with subordinate trachyandesite and andesite. Lignite float is found locally on these rocks. K-Ar ages of 62 to 64 Ma were obtained from the volcanic rocks (Patton and Csejtey, 1980). Oligocene rhyolitic and dacitic tuff with tuff-breccia and flows are intercalated with quartzose sandstone and conglomerate; plant fossils occur locally. These rocks are found on the eastern part of the island; correlative sedimentary rocks on the western part of the island include sandstone, coal, and tuff. A K-Ar age of 39.3 Ma was obtained from volcanic biotite; plant fossils of Oligocene age were collected from the sedimentary sequence (Patton and Csejtey, 1980). These Oligocene rocks are age-correlative to 37 to 43 Ma rhyolites and lesser basalt found in mainland western Alaska (Moll-Stalcup, personal communication, 1986; Miller and Lanphere, 1981).

Quaternary basalt cinder cones and flows form a large field in the north-central part of the island. These alkalic and theoleiitic basalts are part of the Bering Sea basalt field, which extends from the Seward Peninsula to the Pribilof Islands (Moll-Stalcup, personal communication, 1986; Moll-Stalcup, this volume).

Rocks of the Paleozoic to Mesozoic miogeoclinal sequence have been correlated with rocks in northwestern Alaska (Patton and Dutro, 1969). Devonian carbonate rocks with similarities to those found on St. Lawrence Island are found in the western and central Brooks Range. Shallow-water dolostones and marbles with particularly strong lithologic and faunal similarities occur in the Nome Group of the Seward terrane (Till and others, 1986). Mississippian and Triassic rocks of the island are correlative with the Alapah Limestone and Shublik Formation of the North Slope sequence (Patton and Dutro, 1969; Moore and others, this volume).

Saint Lawrence Island occupies a critical position in tectonic reconstructions of the Bering Straits region as a link between lithotectonic belts in Alaska and the northeast U.S.S.R. (Patton and Tailleur, 1977; Box, 1985).

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NOTES ADDED IN PROOF

Considerable advances have been made in understanding the metamorphic evolution of the Seward Peninsula. Conditions of the blueschist-facies metamorphism of the Nome Group were 460 ± 30°C and approximately 12 kilobars (Patrick and Evans, 1989); ductile fabrics formed during that event record high strain (Patrick, 1988). Amphibolite- to granulite-facies rocks of the Kigluaik Group reached peak metamorphic temperatures and pressures of 800°C and 8 kilobars (Lieberman, 1988). Magmatism was apparently synchronous with metamorphism and spanned the period 105–83 Ma (Amato and others, 1992). Interpretations of the tectonic setting of the high-grade metamorphism and the exhumation of the blueschists of the Nome Group remain controversial and have been assigned to both contractional (Patrick and Lieberman, 1988; Patrick, 1988) and extensional settings (Miller and Hudson, 1991; Miller and others, 1992).

Little work has been done on the protolith packages of the metamorphic units. However, a U-Pb zircon age from an orthogneiss body that intruded the lower part of the Nome Group indicates that the mixed unit is in part Proterozoic in age (B. Patrick, oral communication, 1993).

On Saint Lawrence Island, a new conodont collection shows that some of the Paleozoic carbonate rocks are at least as old as earliest Devonian and possibly older (J. Clough and A. G. Harris, oral communication, 1993).

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