FIELD GUIDE TO SOUTH-CENTRAL ALASKA'S ACCRETIONARY COMPLEX, ANCHORAGE TO SEWARD

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INTRODUCTION

This two-day trip, sponsored by the Geological Society of America and AAPG, will traverse the accretionary complex of south-central Alaska (Fig. 1) from Anchorage to Seward. On Day 1, we will stop at exposures along Turnagain Arm and in Whittier, and then drive to Seward for the night. On Day 2, we will tour the Resurrection Peninsula ophiolite by boat, and return to Anchorage. Each of the selected outcrops along Turnagain Arm has been described in previous Alaska Geological Society guidebooks (Clark, 1981; Winkler et al., 1984; Bradley et al., 1997). The Resurrection Peninsula has been featured in two earlier Alaska Geological Society guidebooks (Miller, 1984; Nelson et al., 1989) and also in the Geological Society of America's DNAG field guidebook series (Nelson, Miller, et al., 1987). Despite Alaska's vast size, roads are few, so the same places are visited time and again.

Turnagain Arm, just east of Anchorage, provides a readily accessible cross section through the Mesozoic part of south-central Alaska's accretionary complex. Nearly continuous exposures along the Seward Highway, the Alaska Railroad, and the shoreline display the two main units of the Chugach terrane: the McHugh Complex and Valdez Group. Turnagain Arm, an approximately east-west oriented fjord, is subject to frequent adiabatic ("Chinook") winds. In recent years, outcrops along the highway often have been free of snow and ice even in mid-winter. The last outcrop of the first day will be in Whittier, which lies at the end of what is claimed to be the longest highway tunnel in North America (2.5 miles long).

Resurrection Bay, a fjord south of Seward, features superb coastal exposures of the 57-Ma Resurrection Peninsula ophiolite. This ophiolite formed at a spreading ridge (Bol et al., 1992) which soon thereafter was subducted, leaving only a few preserved fragments. We believe that these are the last surviving pieces of the Resurrection Plate (Haeussler, Bradley, Wells, and Miller, 2003). Although the coast near Seward generally experiences high snowfall, shoreline exposures should be snow-free by early May.

TECTONIC SETTING

Alaska's Pacific margin is underlain by two parallel composite terranes (Plafker et al., 1994). On the inboard side is the Wrangellia composite terrane, which includes the Peninsular terrane; this belt has been the site of episodic arc magmatism since the Early Jurassic. Farther outboard and paired with it is the Chugach-Prince William composite terrane, an accretionary complex during the same time interval. These terranes are thought to have collided as a group with the North American margin in latest Jurassic to Early Cretaceous time (e.g., Trop et al., 2002). The Border Ranges Fault forms the boundary between the Wrangellia and Chugach terranes; it began as a subduction thrust but has been reactivated in various places as a strike-slip or normal fault (e.g.,

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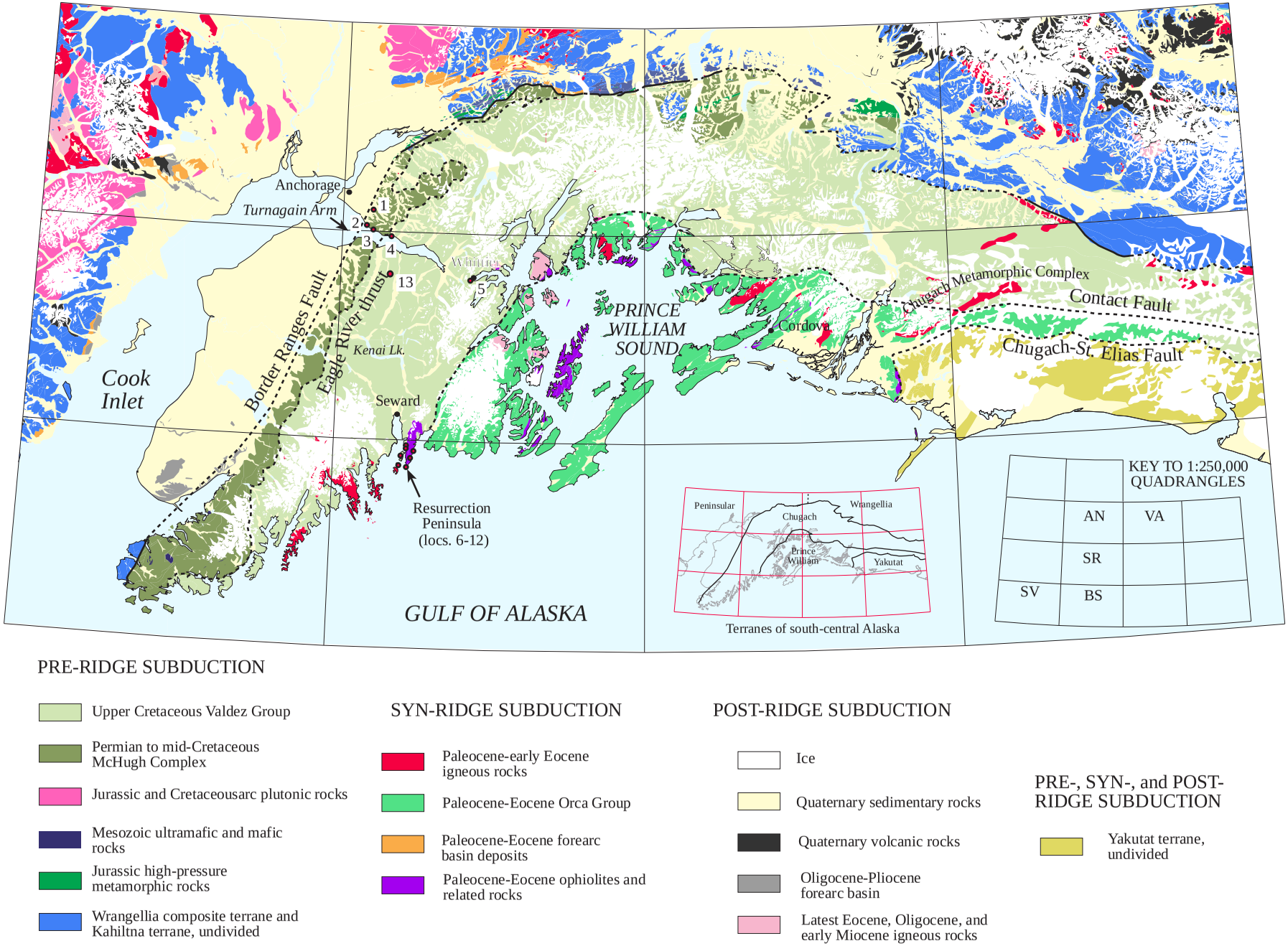

PRE-RIDGE SUBDUCTION

- Upper Cretaceous Valdez Group
- Permian to mid-Cretaceous McHugh Complex
- Jurassic and Cretaceousarc plutonic rocks
- Mesozoic ultramafic and mafic rocks
- Jurassic high-pressure metamorphic rocks
- Wrangellia composite terrane and Kahiltna terrane, undivided

SYN-RIDGE SUBDUCTION

- Paleocene-early Eocene igneous rocks
- Paleocene-Eocene Orca Group
- Paleocene-Eocene forearc basin deposits
- Paleocene-Eocene ophiolites and related rocks

POST-RIDGE SUBDUCTION

- Ice
- Quaternary sedimentary rocks
- Quaternary volcanic rocks
- Oligocene-Pliocene forearc basin
- Latest Eocene, Oligocene, and early Miocene igneous rocks

PRE-, SYN-, and POST-RIDGE SUBDUCTION

- Yakutat terrane, undivided

Figure 1. Generalized geologic map of south-central Alaska, from Bradley et al. (2003) and sources cited therein. Numbers 1-13 are field trip stops. Abbreviations for 1:250,000-scale quadrangles: AN, Anchorage; BS, Blying Sound; SR, Seward; SV, Seldovia; VA, Valdez.

Little and Naeser, 1989). The Mesozoic part of the accretionary wedge is referred to as the Chugach terrane, the Cenozoic part as the Prince William terrane. The distinction between Chugach and Prince William terranes appears artificial (Dumoulin, 1987, 1988) but the names are entrenched and are retained for now.

McHUGH COMPLEX

The McHugh Complex constitutes the older, inboard part of the Chugach terrane in south-central Alaska. On Kodiak Island, equivalent rocks are called the Uyak Complex (Connelly, 1978). Along Turnagain Arm (Fig. 2), the northwestern part of the McHugh Complex is a mélange composed of fragments and disrupted beds of graywacke, mafic volcanic rocks, and chert, in a phacoidally cleaved matrix of argillite and tuff (Fig. 3A). The southeastern part is mainly composed of siliciclastic rocks, including boulder and cobble conglomerate, graywacke, and argillite.

The McHugh Complex is characterized by moderate to intense stratal disruption, which resulted in tectonic juxtaposition of varied rock types. Where mapped in the Valdez (Winkler et al., 1981) and Seldovia (Bradley et al., 1999) quadrangles, fault slices of chert, basalt, and graywacke have been traced tens of kilometers along strike. The Anchorage and Seward quadrangles have not yet been mapped in equivalent detail, but the map pattern is probably comparable. At all scales from thin section to hand sample to outcrop, chert, basalt, and graywacke occur as more competent objects in argillite matrix. The predominant mode of early deformation was layer-parallel fragmentation; breakup of relatively competent beds, such as chert and graywacke, was accompanied by flowage of argillite (± tuff) matrix into gaps. The resulting fragment foliation is the most conspicuous fabric element in the McHugh Complex; along Turnagain Arm, it strikes north-northeast and in most places dips steeply northwest. The foliation is commonly displaced across narrow (up to a few centimeters wide), early ductile shear zones. When the dominant foliation is restored to horizontal, the ductile shear zones show a very consistent sense of slip, seaward side down (Bradley and Kusky, 1992). Clark (1973) reported prehnite-pumpellyite metamorphic facies assemblages in the McHugh Complex along Turnagain Arm. The primary mélange foliation, ductile shear zones, and prehnite-pumpellyite metamorphism are all believed to have formed during the subduction-accretion processes.

Although exposures of basalt in the McHugh Complex are common in some parts of the Chugach terrane, they are sparse and small along Turnagain Arm. Mafic rocks include pillow lava, pillow breccia, massive flows, and diabase dikes, and gabbro in small fault-bounded bodies. On trace-element discriminant plots, various McHugh Complex basalts plot in the N-MORB, E-MORB, and "within-plate" fields (Nelson and Blome, 1991). Age control for the basalt is provided by radiolarians as noted below (mid-Triassic to mid-Cretaceous) (Bradley et al., 1999). A gabbro from the McHugh Complex in the Seldovia quadrangle yielded a concordant $^{206}Pb/^{238}U$ zircon age of 227.7 ± 0.6 Ma (earliest Carnian in the Late Triassic; R. Tucker, written commun., 2000).

Rhythmically interbedded chert and argillite form one of the most distinctive parts of the McHugh Complex. The chert is typically gray or green, less commonly black or red, and is interbedded on the scale of a few centimeters with argillite. Large chert sections, as are exposed along the shoreline of Kachemak Bay in Seldovia quadrangle, are intensely faulted and disharmonically folded. As noted in the road log for Turnagain Arm, Late Triassic, Early

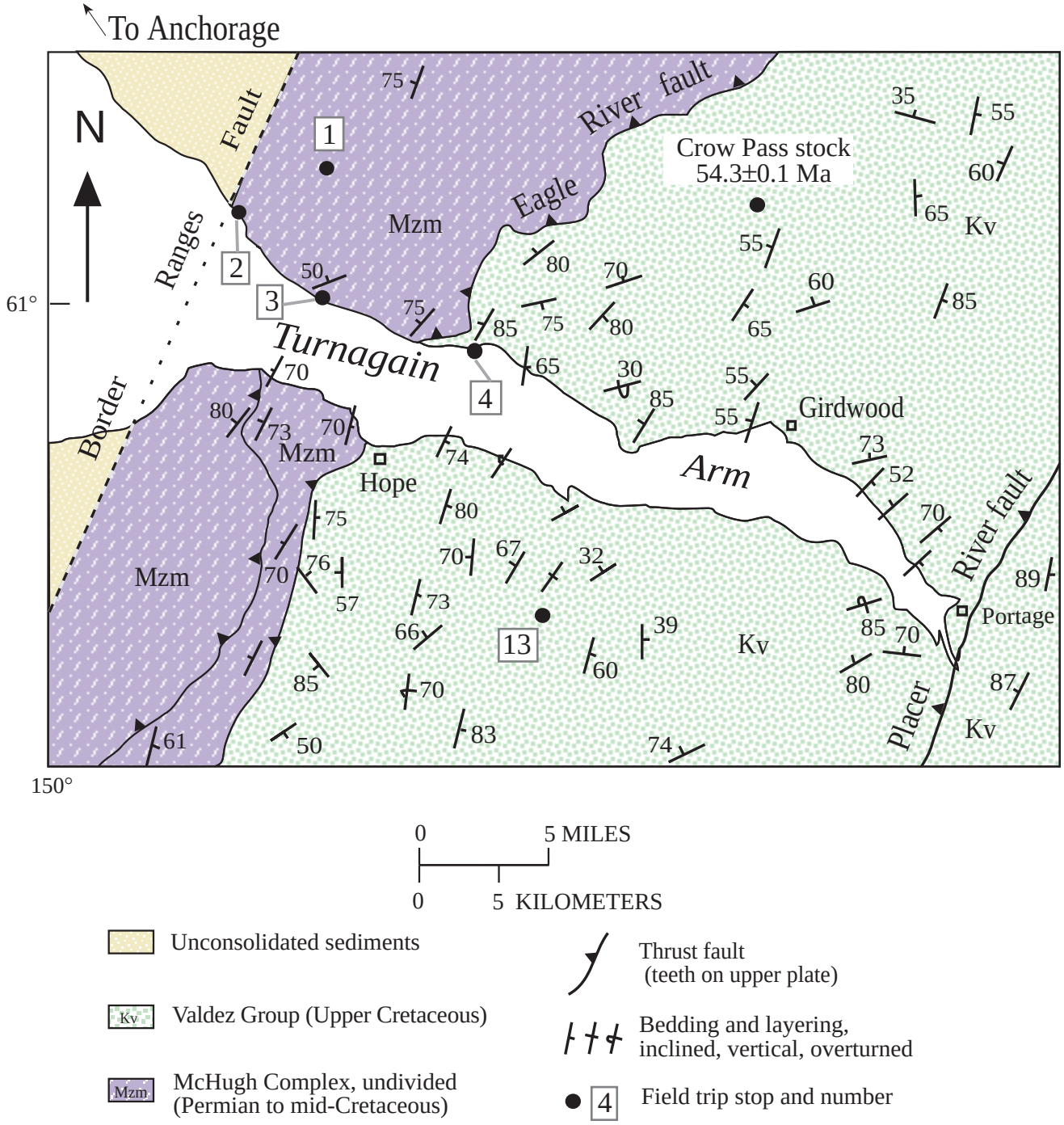

Figure 2. Generalized geologic map of Turnagain Arm, south-central Alaska.

Figure 3A. McHugh Complex conglomerate, at Stop 3 along Turnagain Arm. Clasts include intermediate plutonic rocks, sandstone, argillite, limestone, and greenstone. Note severe deformation of some clasts (e.g., black argillite) but not others (e.g., intermediate plutonic rocks).

Figure 3B. Valdez Group turbidites from Stop 4 along Turnagain Arm. Coherent bedding is common in the Valdez Group but rare in the McHugh Complex.

Jurassic, and Early Cretaceous radiolarians have been reported from the field trip area (Karl et al., 1979; Nelson, Blome, et al., 1987). Equivalent rocks in the Seldovia quadrangle provide more complete age control and some indication of the stratigraphic succession. At several places in that area, radiolarian chert depositionally overlies pillow basalt (Bradley, Kusky, et al., 2000). Precise radiolarian age data show that the base of the chert varies in age from Ladinian (middle Triassic) to Albian-Aptian (middle Cretaceous) (Bradley et al., 1999).

Graywacke and associated conglomerate are major components of the McHugh Complex, especially along Turnagain Arm. The graywacke and conglomeratic graywacke typically occur as massive, structureless bodies. Bedding is seldom observed and even where visible, it generally cannot be traced far (as will be seen at Stop 3). Graywacke is matrix-supported and poorly sorted; the clast population consists of predominant chert and volcanic rock fragments, and subordinate quartz and feldspar. McHugh Complex conglomerates, such as will be seen at Stop 3 (Fig. 3A), contain a broad mix of clast types that are discussed in the description of that stop. At one location in the Seldovia quadrangle, graywacke conformably overlies ribbon chert that yielded Pliensbachian (Early Jurassic) radiolarians (Bradley et al., 1999). Regionally, graywacke is considered to range in age from Pliensbachian through Early Cretaceous.

Limestones within the McHugh Complex form tectonic blocks, typically occurring as severely extended strings of boudins. In the Seldovia quadrangle, Permian fusulinids and conodonts have been recovered from several blocks (Stevens et al., 1997; B. Wardlaw and A. Harris, written commun., 1994). The fusulinids and conodonts are of shallow-water, tropical, Tethyan affinity; the fusulinids are quite distinct from those found within the Wrangellia terrane.

Figure 4 shows a depositional model developed by Connelly (1978) for the Uyak Complex, an equivalent of the McHugh Complex on Kodiak Island. According to this model, the McHugh basalts were formed by seafloor spreading, the overlying cherts were deposited on the ocean floor as it was inexorably conveyed toward a trench, and the argillite and graywacke record deposition on the downgoing plate in and near the trench, just prior to subduction-accretion. In addition to the oceanic-plate component, it is possible that parts of the McHugh Complex were deposited on the overriding plate, for example, on the inner trench slope. The timing of subduction-accretion is not well known, but probably spanned much of the Jurassic and Cretaceous. A discontinuous belt of Early Jurassic (Bradley, Kusky, et al., 2000) high-pressure metamorphic rocks lies along the landward margin of the McHugh Complex. In terms of the Connelly (1978) model, the tectonic limestone blocks might represent the tops of seamounts that were decapitated at the subduction zone. If so, some of the ocean floor that was offscraped to form the McHugh Complex must date back to Paleozoic time.

In addition to these various relatively coherent rocks, much of the McHugh Complex consists of what we refer to as "mesoscale mélange"—blocks of chert, graywacke, greenstone, and rare limestone in a phacoidally cleaved, argillaceous matrix. Genetic studies of McHugh mélanges have not been undertaken in Turnagain Arm, but detailed mapping of McHugh mélange exposures in Seldovia quadrangle (Bradley and Kusky, 1992; Kusky and Bradley, 1999) show that mélange zones mark the fault boundaries between blocks of the various competent lithologies. That mélange flowed along fault zones is suggested by a diverse mix of blocks that don't always match either fault wall. There is no evidence that soft-sediment, near-surface deformation related to gravitational slope failure was an important process in the structural development of the McHugh Complex.

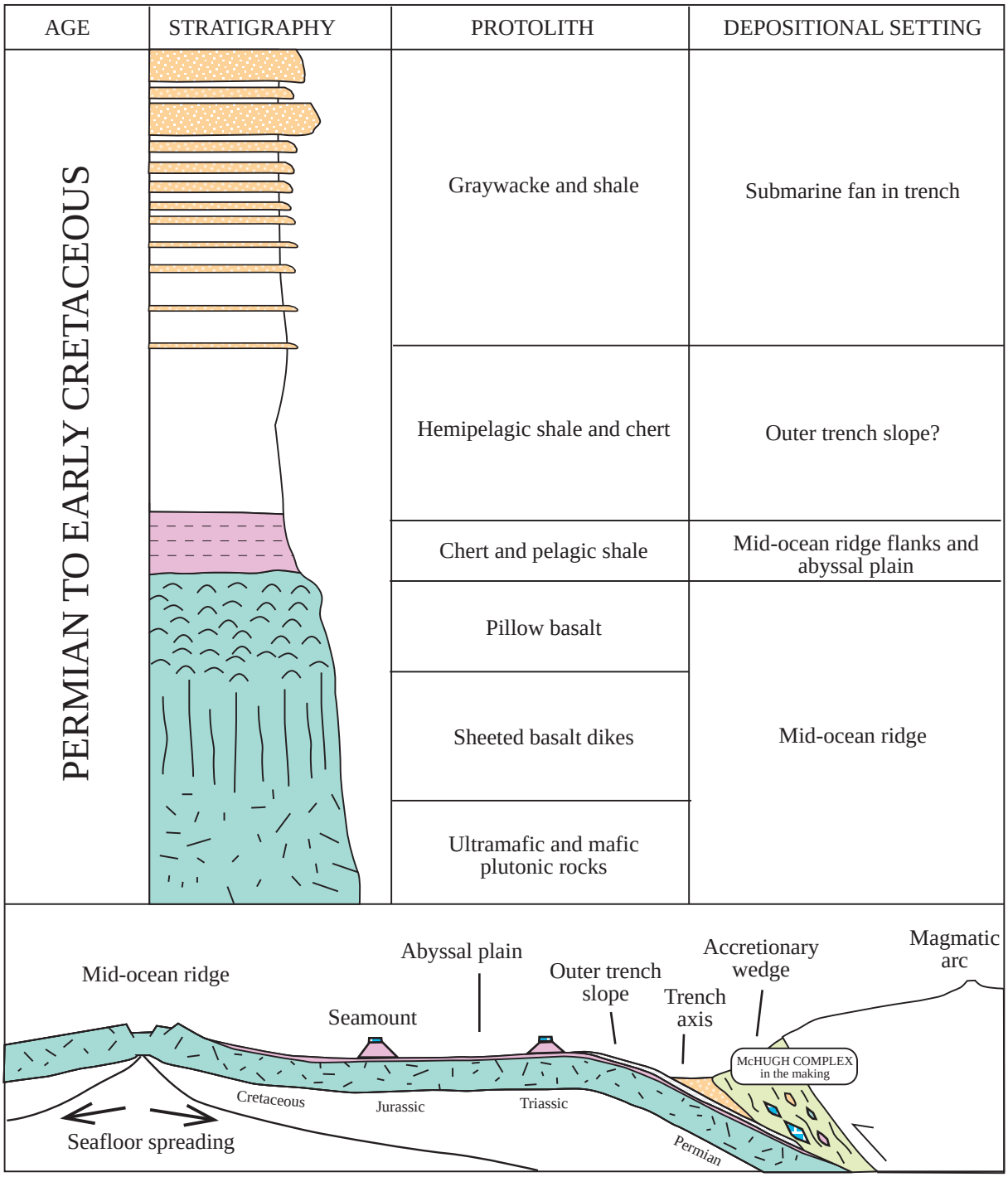

Figure 4. Conceptual model for genesis of the diachronous igneous and sedimentary components of the McHugh Complex, modified from Connelly (1978). The presence of Permian limestones, interpreted as decapitated seamounts, requires that some of the subducted oceanic crust was Permian or older.

VALDEZ GROUP

The seaward part of the Chugach terrane is underlain by the Valdez Group of Late Cretaceous (Campanian? to Maastrichtian) age. Age control is based on a handful of inoceramid bivalves. The Valdez Group includes medium- and thin-bedded graywacke turbidites (Fig. 3B), black argillite, and minor pebble to cobble conglomerate. Sandstones of the Valdez Group are moderately well sorted, and consist mostly of quartz and feldspar, some volcanic fragments, and rare chert. In the Seldovia quadrangle, one conglomerate clast of an intermediate plutonic rock yielded a U-Pb SHRIMP zircon age of about 223 Ma (Bradley, unpub. data, 2004). The Valdez Group was probably deposited on the downgoing plate in a deep-sea trench (Nilsen and Zuffa, 1982), and accreted shortly thereafter. Most of the Valdez Group consists of relatively coherent strata that were deformed into regional-scale tight- to isoclinal folds and cut by a slaty cleavage.

The McHugh Complex and Valdez Group are juxtaposed across a regional-scale thrust fault, which in the area of Turnagain Arm is called the Eagle River Fault. Beneath the fault is a mélange of partially to thoroughly disrupted Valdez Group turbidites. This monomict mélange, which is quite distinct from the polymict mélanges of the McHugh Complex, can be traced for hundreds of kilometers in the footwall of the Eagle River Fault and its along-strike equivalents (Kusky, Bradley, Haeussler, and Karl, 1997).

RESURRECTION PENINSULA OPHIOLITE

A nearly complete ophiolite section is exposed on the Resurrection Peninsula (Fig. 5). From east to west, base to top, the igneous sequence consists of: (1) gabbro, minor plagiogranite, and local pods of ultramafic rock, (2) sheeted dikes, and (3) pillow basalt (Fig. 5); original stratigraphic tops are to the west. Siliciclastic sedimentary rocks interbedded with the pillow lavas suggests the spreading center was relatively near a continental source (e.g., Bol et al., 1992). The ophiolite is interpreted to be 57 Ma based on a U/Pb zircon age from plagiogranite (Nelson et al., 1989). Serpentinized peridotite and pyroxenite occur as pods in the gabbro and, on the east side of the peninsula, as fault-bounded slices in Valdez Group flysch. Dikes and small plugs of plagiogranite intrude the upper parts of the gabbro. The Resurrection Peninsula ophiolite lacks a tectonized peridotite basal section, which would complete the classic ophiolite sequence (Coleman, 1977). Rocks of this mafic-ultramafic complex have undergone ocean floor metamorphism, which is typical of ophiolite sequences (Coleman, 1977), and is marked by low greenschist facies minerals and no penetrative fabric. The contact between the Resurrection Peninsula ophiolite and the surrounding sedimentary rocks is demonstrably a thrust fault on the northeast (Miller, 1984; Nelson, Miller, et al., 1987; Nelson et al., 1989), but the western contact with sedimentary rock is problematic (see below).

The ultramafic rock of the Resurrection Peninsula ophiolite is almost entirely serpentinized. It occurs as pods and small bodies in the gabbro, and as fault-bounded slices in the Valdez Group (Fig. 5; Fig. 6A). In most places enough relict texture and mineralogy remains to recognize original clinopyroxenite, dunite, and harzburgite in thin sections (Miller, 1984). The irregular pods within the gabbro range from a few meters to a few tens of meters across. In the Valdez Group, the serpentinized rock occurs along more linear trends and shows slickensides and many fractures that are locally filled with calcite (Miller, 1984). As also pointed out by Tysdal et al.

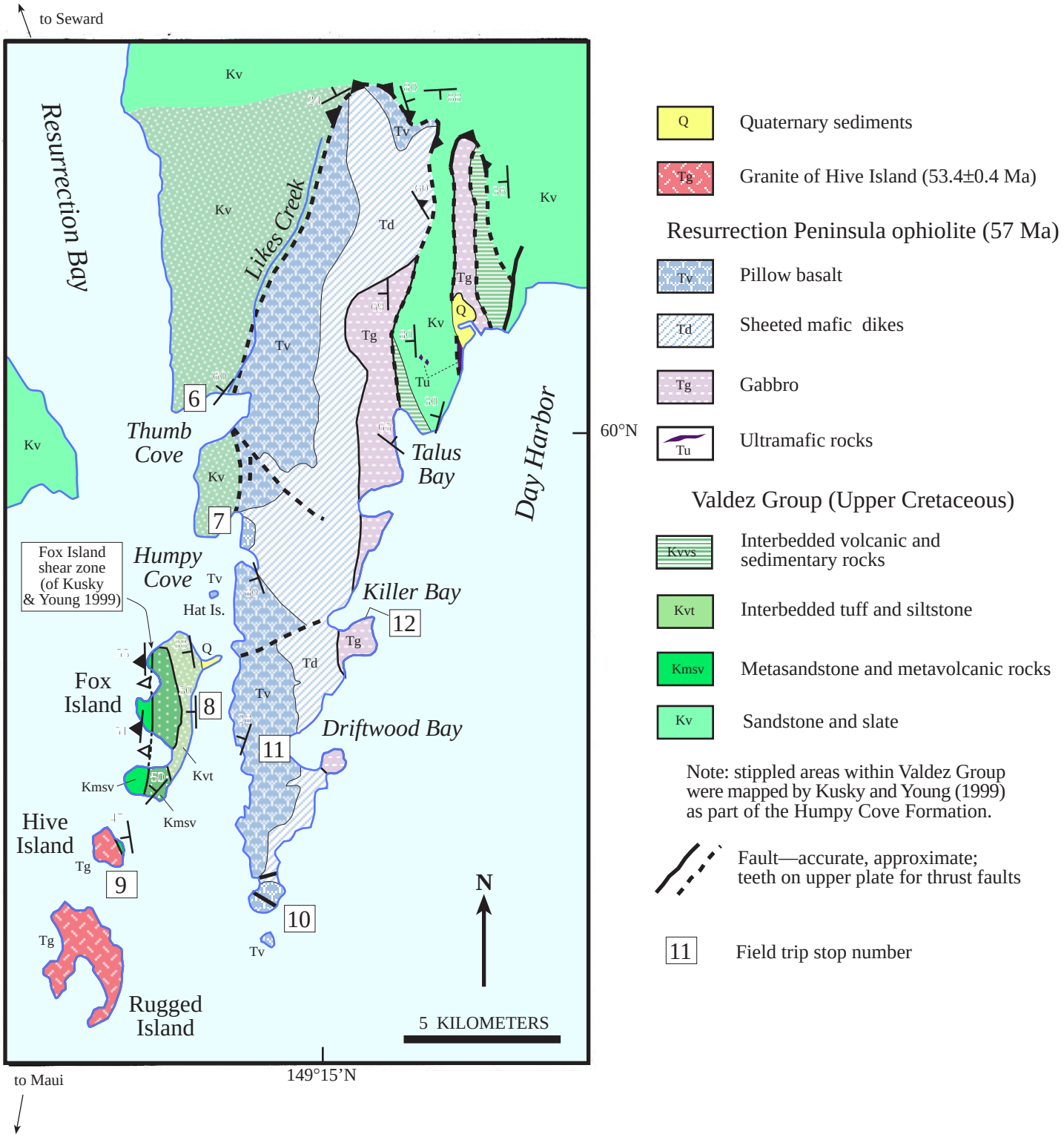

Figure 5. Geologic map of the Resurrection Peninsula, from Miller (1984) and Nelson et al. (1989), but including the Fox Island shear zone of Kusky and Young (1999).

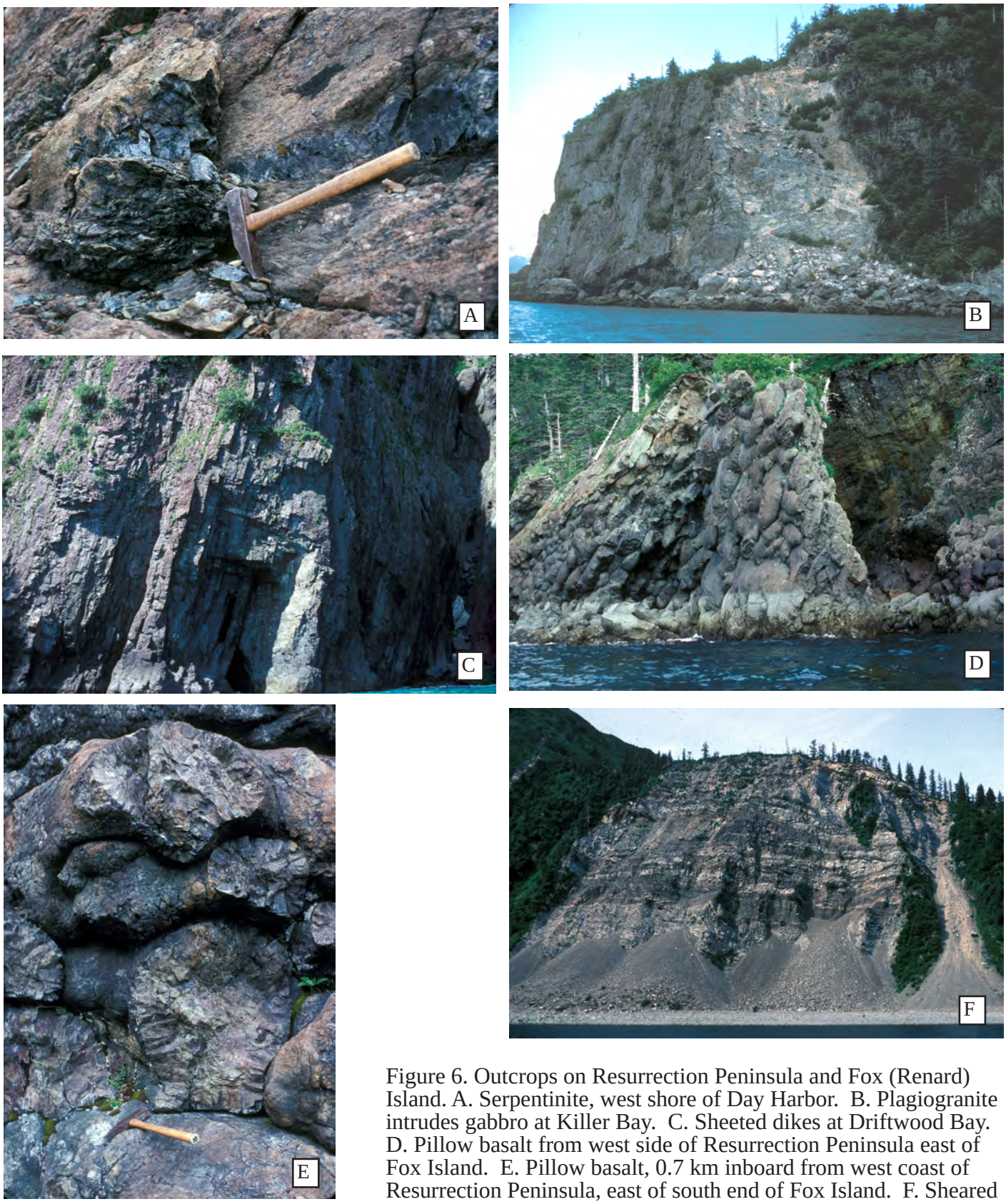

Figure 6. Outcrops on Resurrection Peninsula and Fox (Renard) Island. A. Serpentinite, west shore of Day Harbor. B. Plagiogranite intrudes gabbro at Killer Bay. C. Sheeted dikes at Driftwood Bay. D. Pillow basalt from west side of Resurrection Peninsula east of Fox Island. E. Pillow basalt, 0.7 km inboard from west coast of Resurrection Peninsula, east of south end of Fox Island. F. Sheared Valdez Group metatuff, metasandstone, and argillite, east side of Fox Island.

(1977), the exposure is poor in the northeastern part of the mafic sequence, and the ultramafic rock may be more abundant than shown on the map.

The gabbro occurs as two distinct bodies (herein referred to as the eastern gabbro and the western gabbro) that are separated by a fault-bounded block of interbedded volcanic and sedimentary rocks of the Valdez Group. Locally, well developed, west-dipping magmatic mineral layering occurs in the structurally lowest part of the western gabbro (Miller, 1984). This mineral layering can be observed in four well exposed shoreline outcrops between Driftwood Bay and Talus Bay. At these outcrops, alternating light- and dark-gray layers reflect different proportions of pyroxene and feldspar. The layered gabbro grades westward into massive gabbro containing an increasing percentage of mafic dikes which then grades westward into the sheeted dike unit. Typical exposures consist of medium- to coarse-grained, subophitic clinopyroxene gabbro; plagioclase ranges from about An_{60-70} (optically determined). The clinopyroxene is always at least in part altered to actinolite plus chlorite (uralitized), the plagioclase is sericitized, and secondary epidote is common (Miller, 1984). Stratigraphically higher in the succession, primary hornblende gabbro (plagioclase composition of about An_{55}) crops out, and pods and dikes of hornblende-bearing plagiogranite are present. The largest exposure of plagiogranite is near the south entrance to Killer Bay (Fig. 5, Stop 12; Fig. 6B). At this location, a 20-m-wide stockwork zone of light-gray hornblende plagiogranite intruded gabbro. In a zone extending at least 3 km to the north, other small dikes and pods of plagiogranite intrude the gabbro unit. The plagiogranite is fine to medium grained and contains about 15 percent green hornblende that is slightly altered to actinolite and chlorite. Other constituents include sodic plagioclase, quartz, and intergrain alteration minerals such as epidote and prehnite (Miller, 1984). In a few places, prehnite occurs in veins 1-2 mm wide, cutting both the plagiogranite and the country rock. Also, mafic dikes become more numerous stratigraphically higher in the gabbro. They range from 10-50 cm in width and consist chiefly of fine-grained diabase containing primary hornblende altered to actinolite plus chlorite, plagioclase microlites with brown cryptocrystalline alteration, and interstitial very fine-grained epidote or sphene (Miller, 1984).

Petrographic characteristics indicate that the eastern gabbro body was originally part of the same sequence as the western gabbro, but was stratigraphically lower. The eastern gabbro does not show mineral layering in outcrop (perhaps due to poor exposure), but it does show cumulate textured clinopyroxene gabbro. Additionally, relatively more pods of ultramafic rock are found in the eastern gabbro than in the western gabbro. These aspects, along with no increase in the abundance of basaltic dikes and lack of plagiogranite, would place the eastern gabbro lower in an idealized ophiolite section than the western gabbro (Miller, 1984). Therefore, the sense of displacement between the two gabbro bodies is up to the east.

The gabbro complex grades westward into sheeted dikes that form the rugged crest of the central part of the peninsula (Fig. 6C). The dikes generally trend north-northwest to north-northeast and generally dip steeply to the east, although at the southern end of the peninsula, the dikes trend west-northwest. Cross-cutting dikes locally intrude pre-existing dikes at low angles. Most dikes are from 30 cm to 1 m thick and one-sided chilled margins indicate sheeted intrusion. Aphanitic, porphyritic, and diabasic textures are common even across one outcrop. Low-grade metamorphic assemblages are displayed and include actinolized clinopyroxene, brown-altered plagioclase, interstitial chlorite and epidote, and occasional veins of prehnite (Miller, 1984). The sheeted dikes grade westward to pillow basalt through a mixed zone of increasing pillow screens and volcanic breccia.

Pillow basalt and subordinate massive basalt and broken-pillow breccia make up most of the western flank of the Resurrection Peninsula. The pillow basalt forms a west-dipping sequence and contains minor interbedded siltstone; strike is approximately north and dips are 30-45° to the west. Discrete pillows average 0.5 m in diameter; beautifully preserved forms include long thin pillows, tubes, and stubby and budded pillows (Fig. 5, Stop 6; Figs. 6D, E). Interpillow spaces are locally filled with red and green chert (lacking radiolarians). In thin section the pillows show axiolitic and glassy textures indicating quick cooling. Amygdules are filled with secondary chlorite, zeolites, and in some places epidote; prehnite veins are sparse (Miller, 1984). Siliceous siltstone is locally interbedded with the pillow lavas and shows clay alteration of detrital grains, consistent with the low pressure, hydrothermal, ocean floor metamorphism of the ophiolite sequence (Miller, 1984). Trace-element abundances from basalts of the Resurrection Peninsula ophiolite and related ophiolites in Prince William Sound are broadly MORB-like but show some enrichment in incompatible elements (Lytwyn et al., 1997; Nelson and Nelson, 1993; Crowe et al., 1992). These trends can be explained in terms of contamination of normal MORB with sediment in a near-trench setting. A paleomagnetic study suggested that when it formed, (1) the ophiolite was 13±9° south of its present position with respect to cratonic North America, and (2) the strike of the ridge was about 026±12° (Bol et al., 1992, but see Haeussler, Bradley, Wells, and Miller, 2003).

We now turn to the contact relationships between the ophiolite and adjacent largely sedimentary rocks. As stated previously, the northern and northeastern contact (Fig. 5) is clearly a thrust fault separating the 57-Ma ophiolite from the 70-65 Ma Valdez Group (Miller, 1984; Nelson et al., 1989). The western contact is problematic and interpreted as either depositional or faulted. Tysdal and Case (1979) thought the contact was depositional and that both the ophiolite and a purportedly capping sedimentary section were Cretaceous in age and part of the Valdez Group. Kusky and Young (1999) also interpreted a depositional contact on the west side, but given the 57 Ma age of the ophiolite, they regarded the sedimentary section as late Paleocene to early Eocene and named these rocks the "Humpy Cove Formation" of the Orca Group[3]. Noting a significant difference in metamorphic grade across this boundary, Miller (1984), Nelson, Miller, et al. (1987), and Nelson et al. (1989) interpreted this same contact as likely a fault separating Valdez Group from the ophiolite. The contact itself can be most closely approached in Humpy Cove (Fig. 5, Stop 7). Here, Kusky and Young (1999) reported that turbidites overlie the pillow basalts; however, cleavage along this contact obliterates bedding and leaves open a fault interpretation (Miller, 1984). All workers agree that some siliciclastic rocks are demonstrably interbedded with the pillow lavas; the question is how much original sedimentary section is preserved above the pillow lavas—a minor amount, or as Kusky and Young (1999) suggested, ~2.5 km worth?

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[3] As further support of their conclusion that the rocks on the west are not Valdez Group, Kusky and Young (1999) cited the work of Dumoulin (1988), who reported that the Valdez Group sandstones of the "Resurrection Peninsula area" differed from most Valdez Group samples in that they were finer grained, better sorted, and had higher feldspar content. The problem is, the "Resurrection Peninsula area" samples cited by Dumoulin (1988) were from northeast of the mafic complex, <u>not</u> from the problematic strata, hence have no bearing on Kusky and Young's argument.

Fox Island[4], which lies a few kilometers west of Resurrection Peninsula (Fig. 5, Stop 8), is a key area for interpreting the purportedly capping sedimentary rocks. On the east side of the island a cliff exposure shows vari-colored units of interbedded tuff and siltstone (Fig. 6F). These rocks are metamorphosed to biotite zone of greenschist facies, or approximately middle greenschist facies (Miller, 1984). Although metamorphosed and locally semischistose, original protoliths are easily identified as silty argillite, sandy siltstone, tuff, and rare flows or sills. On the west side of the island metasandstone and metavolcanic rocks are strongly semischistose and of even higher metamorphic grade. The metasandstones show abundant syndeformational secondary biotite, lesser muscovite, and minor chlorite, indicating metamorphic grade well into the upper greenschist facies (Miller, 1984). The metavolcanic rocks have secondary assemblages (hornblende, plagioclase, and minor biotite) indicative of amphibolite facies metamorphism (Miller, 1984). The metamorphic grade displayed in the rocks of Fox Island is distinctly higher than that of the ophiolite and its demonstrably interbedded sedimentary rock, but is completely consistent with that displayed by Valdez Group rocks on trend to the north of the Peninsula.

Fox Island may also provide evidence bearing on the emplacement of the ophiolite, but this too is controversial. Kusky and Young (1999) mapped a shear zone on the west side of Fox Island (Fig. 5), which in their view separates Cretaceous Valdez Group on the west from the purported ophiolite cover rocks on the east. They documented a west over east shear sense and argued that the ophiolite was emplaced into the accretionary prism along this structure. This interpretation rests on two others: 1) that the rocks to the west of the shear zone are Cretaceous Valdez Group, and 2.) that rocks to the east depositionally overlie the ophiolite. Extrapolating offshore to the south, they interpreted this shear zone to be intruded by the 53.4 Ma (Bradley, Parrish, et al., 2000) granite of Hive Island (Kusky and Young, 1999)—something that is not self-evident from the map pattern. Thus they bracketed the emplacement of the ophiolite into the accretionary prism between 57 Ma and 53.4 Ma. These numbers fed, in turn, into conjectures about the travel distance of the ophiolite from an original location well to the south in the northern Pacific. All this rests, however, on the interpretation that the rocks of eastern Fox Island are the sedimentary cap of the ophiolite section, a concept that we question. Attempts to determine the age of these problematic strata by dating the tuffs have failed. We plan to next try U-Pb geochronology on detrital zircons from Fox Island sandstone hoping to solve the dilemma; Paloecene zircons would support the Kusky and Young (1999) interpretation.

NEAR-TRENCH INTRUSIVE ROCKS

Paleocene to early Eocene (~61-50 Ma) near-trench intrusive rocks are a key part of the tectonic history of the Chugach-Prince William composite terrane. Hudson et al. (1979) recognized and named these intrusive rocks the Sanak-Baranof belt. Intrusive bodies range from narrow dikes to giant granitic batholiths 10 by 100 km (Fig. 7A, B). Remarkably, the plutons intrude turbidites that only a few million years earlier were being deposited along the trench axis. The plutons show a remarkably smooth time-transgressive age pattern: from 61 Ma in the west to 51-50 Ma in the east, for some 2200 km along strike (Bradley et al., 2003 and older references therein). For these and other reasons, the Sanak-Baranof plutonic belt is regarded as marking the site of a trench-ridge-trench triple junction, which migrated from west to east (e.g., Bradley et al., 2003).

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[4] The island is variously known as Fox Island and Renard Island, the French name for fox. The U.S. Geological Survey uses the latter name on its topographic maps.

A

B

C

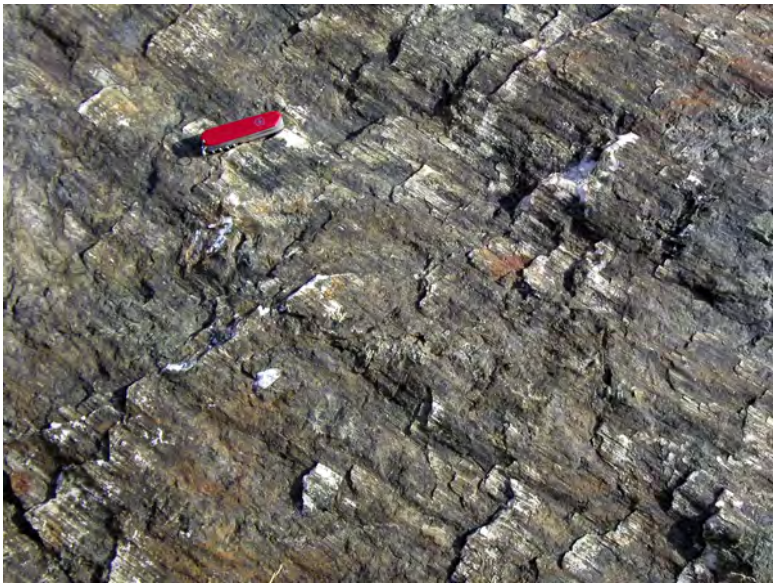

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Figure 7. Photographs of features attributed to early Tertiary ridge subduction. A, Dike swarm cutting McHugh Complex, Seldovia quadrangle. B, Chernof pluton, Seldovia quadrangle. C, Beauty Bay gold mine, Seldovia quadrangle. D, Slickensided fault surface, Stop 3, Turnagain Arm.

METAMORPHISM

A high-P, low-T metamorphic belt is the most landward part of the Chugach terrane. This metamorphism in Seldovia quadrangle has been dated at 191-192 Ma ($^{40}Ar/^{39}Ar$ white mica and amphibole; Bradley, Parrish, et al., 2000). The high-pressure belt is discontinuous and does not crop out on the field-trip transect. The McHugh Complex was regionally metamorphosed, presumably during subduction-accretion, to prehnite-pumpellyite facies. This "event" was presumably a long one that happened before Maastrichtian time, but it has not been directly dated. The Valdez Group was regionally metamorphosed to low greenschist facies during the latest Cretaceous to early Paleocene (pre-Sanak-Baranof intrusives). As described above, the presumable Valdez Group units on Fox Island exhibit mid and upper greenschist facies. The 61-50 Ma near-trench intrusive pulse was accompanied by contact metamorphism to cordierite grade, and by development of a few regional-scale, low-P, high-T metamorphic highs. The largest of these is the Chugach Metamorphic Complex east of Cordova (Fig. 1; Pavlis and Sisson, 1995), but others are known as well (see Fig. 14 of Bradley et al., 2003).

GOLD MINERALIZATION

The Chugach terrane hosts hundreds of early Tertiary lode gold occurrences, prospects, and a few small mines (Fig. 7C). The Kenai Peninsula has had a long history of mineral activity. The first report of gold was in 1848 when a Russian surveyor discovered gold in the Russian River area. Placer gold was mined by the Russians near Kenai Lake in the early 1850's (Jansons et al., 1984). By the late 1800's placer gold miners were active in the Kenai Peninsula area and lode gold was subsequently discovered on the Kenai Peninsula, near Girdwood, in northern Prince William Sound, and near Valdez (Nelson and Miller, 2000). The first notable production came in 1911 from mines in the Falls Creek area (Jansons et al., 1984). The most recent significant lode gold production from the region was in the 1930's and 1940's. Little or no production from lode gold deposits has taken place since 1956, but small-scale placer mining continues today. Over the nearly 100 years of mining, the northern Kenai Peninsula has produced ~133,800 oz. of placer gold and an additional 30,000 oz. of lode gold (Huber and Kurtak, 2001).

The gold-bearing veins occur along brittle faults that truncate accretion-related structures, mostly in the Valdez Group, but also rarely in the McHugh Complex (Haeussler, Bradley, and Goldfarb, 2003). Quartz, calcite, and ankerite are the typical gangue minerals, arsenopyrite is the most abundant sulfide, and tiny grains of native gold can still be found on many mine dumps. Goldfarb et al. (1986) invoked a regional-scale ore-forming process because the vein mineralogy, isotopic data, and fluid inclusion composition, salinities, and homogenization temperatures are consistent over a very wide area. The mineralizing fluids were most likely metamorphic in origin (Goldfarb et al., 1986). The forearc setting of these gold occurrences is unusual, as forearcs are normally relatively "cold" places not noted for hydrothermal activity. $^{40}Ar/^{39}Ar$ dating of sericite from a number of gold mines has shown that mineralization took place at 57-53 Ma in the Kenai Peninsula area, and at 52-49 Ma in southeastern Alaska (Haeussler et al., 1995). Near-trench gold mineralization thus was essentially coeval with the diachronous pulse of near-trench magmatism, and by implication, the two processes occurred in the same tectonic setting.

BRITTLE FAULTING

Throughout southern Alaska, rocks of the Chugach terrane are cut by abundant late brittle faults (Bradley and Kusky, 1990; Kusky, Bradley, and Haeussler, 1997) (Fig. 7D). Along Turnagain Arm, these late faults include sets of: (1) dextral and sinistral strike-slip faults, (2) synthetic and antithetic thrust faults, and (3) synthetic and antithetic normal faults. The faults are typically spaced every few meters to tens of meters, and can be seen at most of the outcrops visited on this trip. The thrust faults shortened the wedge subhorizontally nearly perpendicular to strike; then, the somewhat younger normal and strike-slip faults extended the wedge nearly parallel to orogenic strike. The three fault sets are characterized by quartz + calcite + chlorite slickensides; curved slickenlines on some faults of each set reveal that displacement directions changed over time. We believe that this resulted from progressive changes (relative to the rocks, at least) in the orientation or magnitude of principal stresses during exhumation of the accretionary wedge, while the faults were active (Kusky, Bradley, and Haeussler, 1997). Although none of the brittle faults along Turnagain Arm have been dated, the similarity in their character with the gold-quartz veins suggests that they were active at the time of near-trench magmatism (Haeussler, Bradley, and Goldfarb, 2003). As noted above, most of the gold-quartz veins in the Chugach terrane—including those that have been dated—occupy strike-slip and normal faults that resemble the ones along Turnagain Arm.

LATE CRETACEOUS TO PALEOGENE PLATE-TECTONIC INTERPRETATIONS

The Paleocene and early Eocene history of the Gulf of Alaska margin can be best explained as the consequence of ridge-trench interaction, or "ridge subduction."[5] The Sanak-Baranof plutons were diachronously intruded along the entire ~2000-km length of the accretionary complex, from 61 Ma in the west to 51-50 Ma in the east. It is difficult to envision a plausible mechanism, other than ridge subduction, for generating intrusions in a near-trench setting with such a large-scale trend in ages; alternative ideas have been discussed at length by Bradley et al. (2003). Other geologic effects of ridge subduction are shown in Figure 8.

Global plate reconstructions imply that a spreading center was subducted somewhere along the western margin of North America during the early Tertiary (Atwater, 1989). Most of what is known about the ancient seafloor of the northwestern Pacific is from the mirror-image magnetic anomaly pattern embedded in surviving parts of the Pacific Plate. The magnetic stripes require that three oceanic plates existed (Pacific, Kula, and Farallon), that the Kula-Farallon Plate boundary was a ridge, and that this ridge must have trended toward the western margin of North America. The Sanak-Baranof near-trench plutonic belt thus might have marked the place where the Kula-Farallon ridge met the margin (e.g. Bradley et al., 1993) (Fig. 9A). However, evidence for simultaneous ridge subduction in Alaska and Washington (Fig. 9B) suggests a more complex plate geometry involving an additional ridge and thus a fourth oceanic plate (Fig. 9C). We have called this the Resurrection Plate (Haeussler, Bradley, Wells, and Miller, 2003; Miller et al., 2002) after the Resurrection Peninsula, its largest surviving piece. The post-50-Ma history of ridge subduction has recently been discussed by Madsen et al. (2006).

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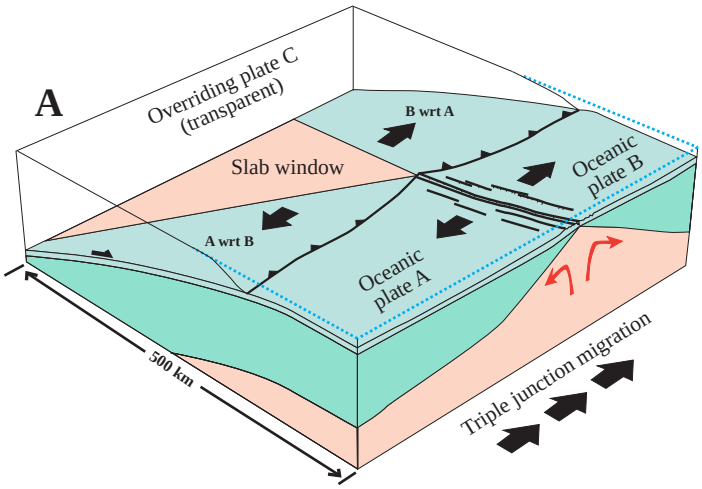
[5] Ridge subduction in this case refers to the subduction of the boundary between two diverging oceanic plates. Over time, the continued separation of these plates at depth leads to the opening of a "slab window" (Thorkelson, 1996).

Figure 8. Block diagrams illustrating the geologic effects of ridge subduction in south-central Alaska. (A) The overriding plate has been rendered transparent to schematically show the opening of a slab window downdip of the spreading center. The gap between subducting, diverging slabs allows hot asthenosphere to come into contact with the normally cold base of the forearc. Beneath the arc axis, where dehydration reactions would ordinarily give rise to arc magmas, there is no plate to dehydrate and hence a gap in arc magmatism over the slab window, as shown in (B).



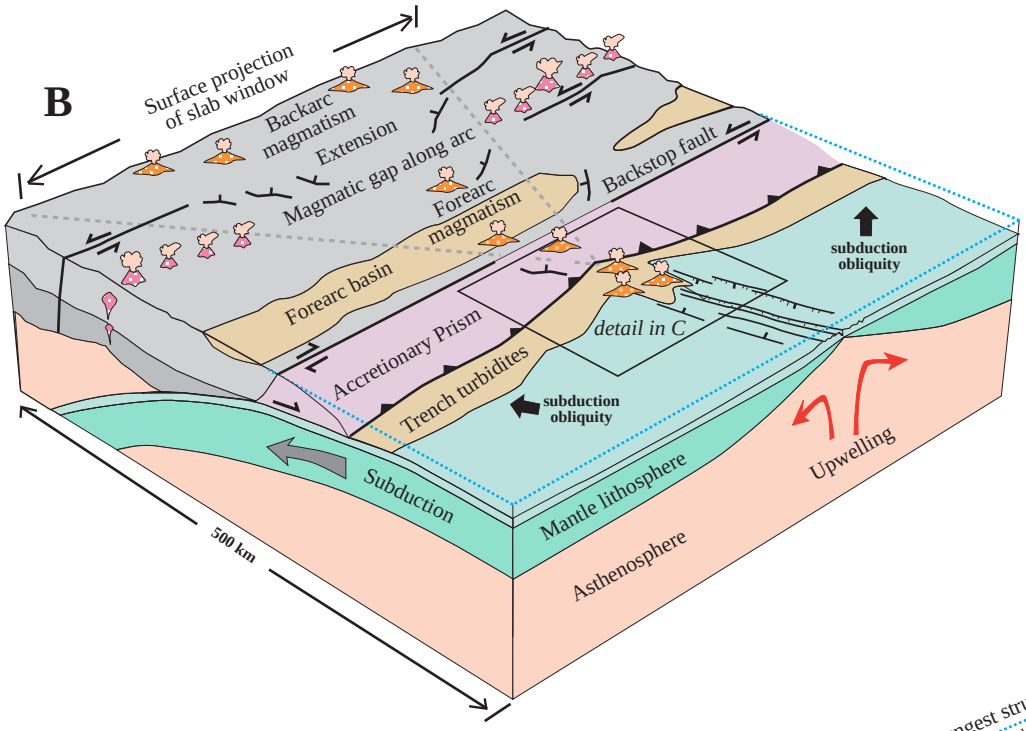
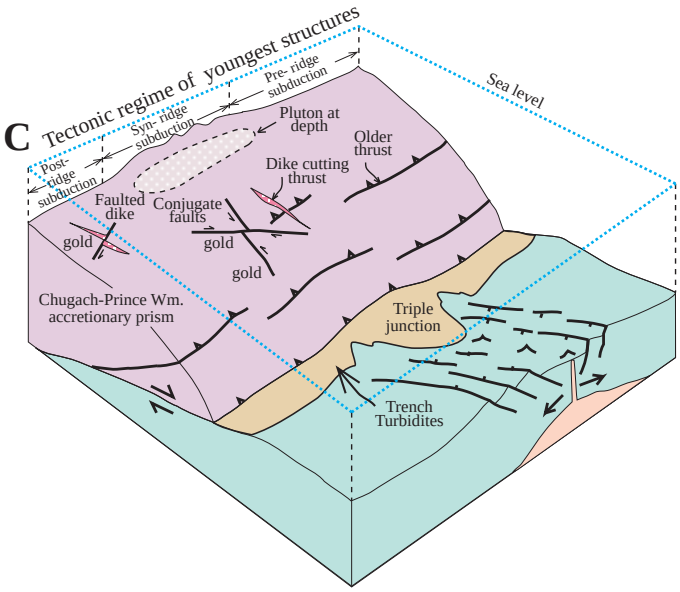


Figure 8B. Igneous activity along the arc just prior to the magmatic lull might perhaps be related to the thin, hot, trailing edge of the subducted plate or to the first interactions between slab-window asthenosphere and the base of the arc. Margin-parallel strike-slip faults are a response to oblique convergence that affects the upper plate differently on either side of the triple junction. Transtension drives subsidence of the forearc basin.

Figure 8C. In the triple junction area, ocean-floor basalts formed by seafloor spreading are interbedded with trench turbidites. Depending on bathymetric details, turbidity flows might either be funneled into or blocked from the axial valley of the ridge, and accordingly, both Besshi- and Kuroko-type massive sulfide mineralization are possible in the triple junction area. The presence of near-trench, forearc volcanoes is inferred by analogy with the Woodlark triple junction; only near-trench plutons and dikes remain at the crustal levels now exposed in Alaska. Brittle faults, some of them mineralized by gold-quartz veins, facilitate orogen-normal shortening (i.e. the narrowing of the prism) and orogen-parallel extension.



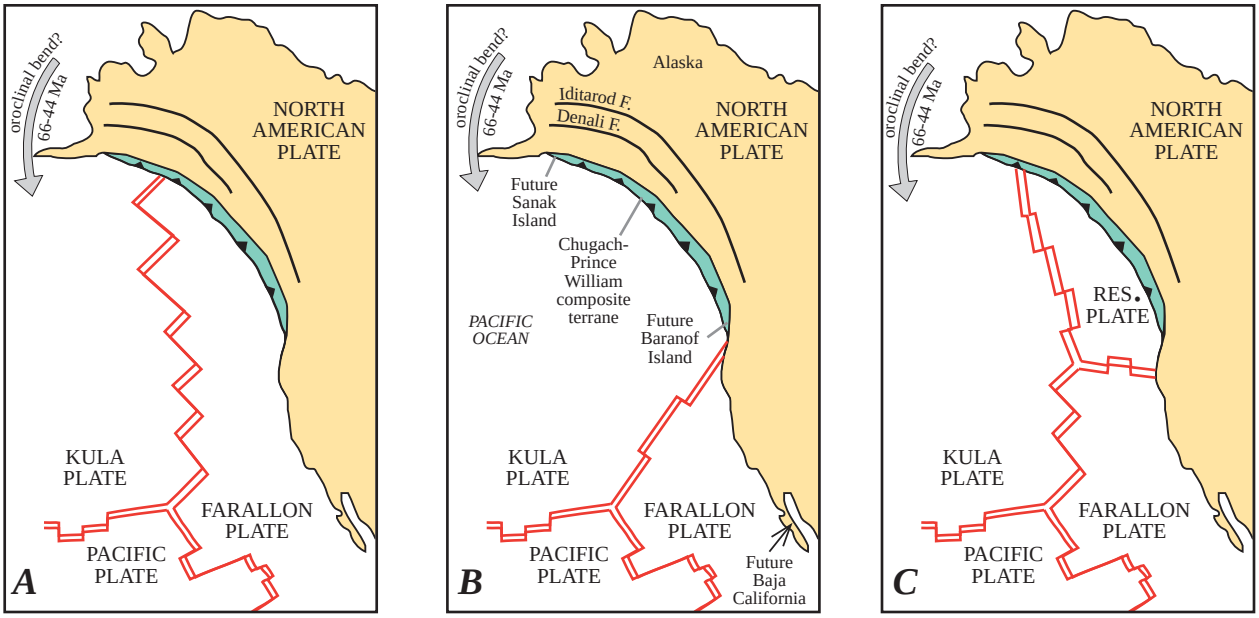


Figure 9. Alternative plate models for 57 Ma. A. Kula-Farallon ridge subducted beneath Chugach terrane. B. Kula-Farallon ridge subducted beneath Washington. C. Preferred model: Kula-Resurrection was ridge subducted beneath the Chugach terrane while the Resurrection-Farallon ridge was subducted beneath Washington.

ROAD LOG FOR DAY 1

The route description begins at the O'Malley Road overpass on the Seward Highway in Anchorage. (This is about 5 miles south of the intersection of the Glenn Highway and Seward Highway.) From here there are two options depending on the weather. If the visibility is good we will drive east to Stop 1, and then retrace the route back to O'Malley and Seward Highway. If the visibility is bad, we will skip Stop 1 and drive south from here on the Seward Highway.

Mile 0. Intersection of Seward Highway and O'Malley Road.

To Stop 1: Drive east on O'Malley Road towards the Chugach Mountains.

Mile 2.0. Alaska Zoo on the left.

Mile 3.6. Turn right onto Hillside.

Mile 4.5. Turn left onto Upper Huffman.

Mile 5.3. Turn right onto Toilsome Hill Road.

Mile 6.0. Road turns to gravel.

Mile 7.2. Turn left into the Glen Alps parking lot. A sign reads "Entering Chugach State Park." There is an Alaska State Parks parking fee.

STOP 1. GLEN ALPS OVERLOOK

Walk to the scenic overlook, usually a windy place.

The subduction of the Pacific Plate beneath the North American Plate controls the tectonics of southern Alaska. The Pacific Plate is about 40 km beneath our feet, and is currently subducting beneath Anchorage at a rate of 54.3 ± 1.4 mm/yr toward an azimuth of 344° (using the relative motion poles of DeMets et al., 1994). Three main parts of a classic subduction zone can be seen from here: the Aleutian volcanic arc, the Cook Inlet forearc basin, and the Chugach terrane accretionary complex.

The Aleutian volcanic arc is located approximately 100 km above the subducting Pacific Plate, and it stretches about 2500 km from Hayes Volcano (north of Mt. Spurr) almost to the end of the Aleutians. The distance between the toe of the accretionary complex and the magmatic arc is a wider-than-average 460 km. Forty-two volcanoes have been active since 1760, accounting for more than 265 eruptions during that time. Eighty volcanic centers show evidence for Holocene activity. Arc magmas range from basalts to rhyolites, but are dominantly andesites. Mt. Spurr to the north and Mt. Redoubt to the south are the two closest volcanoes visible from here.

The Chugach and Kenai Mountains consist of accreted ocean-floor rocks of the Chugach terrane. As discussed above, these rocks originated in the paleo-Pacific, and were accreted to the continental margin, by some combination of offscraping and underplating, during the

Jurassic and Cretaceous. The Border Ranges fault (originally a subduction-zone thrust, subsequently reactivated) lies in the lowlands along the mountain front.

Between the volcanic arc and the accretionary complex is the Cook Inlet forearc basin, which has a long, discontinuous history of sedimentation since Triassic time. Tertiary strata in Cook Inlet have been deformed into open, assymetric, and fault-cored folds that Haeussler et al. (2000) have attributed to transpression. The folds are important structural traps of oil and gas in the Cook Inlet basin—the first petroleum province extensively developed in Alaska. Haeussler et al. (2000) argued that these folds are still active and may be important seismic hazards in the Anchorage area.

The oldest known glaciation in the Anchorage area is thought to have covered the top of Mt. Susitna (4,396 ft.)(Karlstrom, 1964) and possibly the top of nearby Flattop Mountain (3,510 ft.)(Schmoll et al., 1984)—both summits should be visible from this overlook. The lowlands of the Anchorage bowl are largely underlain by sediments of the Bootlegger Cove Formation, dated at 13,690 to 14,900 radiocarbon yr. B.P. (Schmoll et al., 1972). Updike and Ulery (1986) regarded these sediments as fan-delta deposits, related to an ice lobe located just west of Anchorage. The Elmendorf Moraine, is visible to the north as a band of low, hummocky hills, formed during the last glacial maximum between 12,000 and 14,000 radiocarbon yr. B.P., when a large ice lobe extended down the Matanuska Valley (Schmoll et al., 1984).

On a clear day, Mt. McKinley (Denali), 20,320 feet tall, can be seen 150 km to the north-northwest.

Retrace path to the New Seward Highway.

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Set or reset odometer to zero at the O'Malley-Seward Highway overpass. Begin mileage log for Turnagain Arm.

Mile 0. Drive south on the New Seward Highway toward Turnagain Arm.

Mile 2.6. Rabbit Creek Road overpass on the New Seward Highway. Just ahead, the road narrows to two lanes and is notoriously dangerous because of unsafe passing.

Mile 5.8. East end of Potter Marsh. This is the approximate buried trace of the Border Ranges fault system, which separates the Peninsular and Chugach terranes (Fig. 1).

Mile 6.0. Turn left into Weigh Station.

STOP 2. McHUGH COMPLEX, PILLOW BASALT AND MÉLANGE.

Park at the west end of weigh station parking lot, and carefully cross the Seward Highway and railroad tracks to an outcrop of pillow basalt along the shoreline, opposite the stop sign. This is about as good as pillows get in Turnagain Arm; in Seldovia quadrangle, there are dozens of more spectacular exposures of likely correlative rocks. This basalt has not been

dated, but is probably Middle Triassic to mid-Cretaceous, by extrapolation from the Seldovia quadrangle.

From here, walk westward along the shore a few hundred feet to the next outcrop, which consists of mesoscale mélange. Enclosed in a phacoidally cleaved argillite matrix are isolated blocks, and variably pulled-apart layers of greenstone, chert, limestone, and graywacke. The tall roadcut opposite this outcrop also consists of mesoscale mélange, but the clean weathered surface exposed here shows more detail. Note the large limestone block in the roadcut.

Return to vans, continue along Seward Highway. The type locality of the McHugh Complex extends from here at the Weigh Station, past the next stop at Beluga Point, to a short distance west of Falls Creek.

Mile 6.7. In the roadcut on the left is a conspicuous lozenge of red, ribbon-bedded radiolarian chert. Five species of radiolarians indicate a Berriasian to Valanginian (Early Cretaceous) age for this chert (Karl et al., 1979).

Mile 7.4). There are several different paleontologically dated cherts in this part of the McHugh Complex. A gray chert lens in a 150-meter road cut of predominantly massive greenstone accompanied by minor chert and argillite yielded Berriasian to Valanginian radiolarians (Nelson, Blome et al., 1987). A hundred meters or so to the south is a 10-meter-thick lens of red and gray ribbon-bedded chert, in basalt, that yielded radiolarians of late Pleinsbachian to Toarcian (Early Jurassic) age (Nelson et al., 1986). Four meters to the south is another block of red ribbon chert containing Berriasian to Hauterivian (Early Cretaceous) radiolarians (Nelson, Blome, et al., 1987).

Mile 7.6. Gray chert in centimeter-scale lenses and boudins occurs in a sheared argillite matrix. This chert contains radiolarians of late Carnian to early Norian (Late Triassic) age (Nelson, Blome et al., 1987).

Mile 9.1. Cross McHugh Creek, from which the McHugh Complex took its name.

Mile 10.6. Turn right into parking lot at Beluga Point, and park at the east end of the lot.

STOP 3. BELUGA POINT

Keeping an eye out for yahoos, cross the highway to cliffs of McHugh Complex sandstone and boulder conglomerate. Clasts found here include greenstone, argillite, chert, limestone, sandstone, siltstone, and gabbro. A plutonic clast from a nearby outcrop yielded a hornblende K/Ar age of 146±7 Ma (M.A. Lanphere, quoted in Clark, 1981). Elsewhere in the Anchorage quadrangle, Clark (1981) reported clasts of glaucophane schist and serpentinite in conglomerate of the McHugh Complex. Conodonts from a limestone cobble in McHugh conglomerate 2 km southeast of Beluga Point are Meramecian to Morrowan in age (Late Mississippian to Early Pennsylvanian; Nelson et al., 1986); this clast could have been shed from the Strelna Formation of the Wrangellia terrane (Nelson et al., 1986).

Sandstones of the McHugh Complex typically are poorly sorted, matrix-supported, and consist mostly of chert and volcanic rock, plus less abundant quartz and feldspar. A comprehensive study of the McHugh Complex conglomerate clasts—using the full arsenal of modern techniques—would be of enormous value in plate reconstructions.

On the lower cliff face, about opposite the information kiosk, are the slickensided surfaces of several late brittle faults that are widespread throughout Chugach terrane on the Kenai Peninsula (Bradley and Kusky, 1990; Kusky, Bradley, and Haeussler, 1997). The most conspicuous of these is an oblique dextral-normal fault.

Return to vans, continue along Seward Highway.

Mile 15.3. Falls Creek. Here we cross the Eagle River Fault, a low-angle thrust. Driving eastward we leave the McHugh Complex in the hangingwall and enter the Valdez Group in the footwall. For about the next kilometer, the Valdez Group is quite disrupted. Bedding, where preserved, is generally inverted, suggesting that disruption may be the consequence of shearing on the overturned limb of a recumbent footwall syncline. In the same structural position in the Seldovia quadrangle, the belt of thoroughly disrupted Valdez Group has been mapped as the "Iceworm mélange" (Kusky, Bradley, Haeussler, and Karl, 1997). All roadcuts from here to Seward will be Valdez Group.

Mile 16.3 First coherent section of Valdez Group.

Mile 16.7. Park on the right at a pullout, located in the middle of a broad bend to the left.

STOP 4. VALDEZ GROUP NEAR INDIAN

Scramble down a tricky footpath, cross the railroad track, and descend toward the water's edge, keeping to the left of the prominent headland. The wave-washed outcrops are of thin- and medium-bedded turbidites of the Valdez Group, most of which display partial Bouma sequences (Tcde). Sedimentary structures are quite well preserved, to a degree seldom seen in the McHugh Complex, but fairly common in parts of the Valdez Group. Paleocurrents here, based on single-tilt restoration of cross laminae (n = 7), are toward about 215°, roughly parallel to the present structural grain and to the inferred paleotrench axis. Outcrops to the left and right are of thick, locally conglomeratic sandstones.

Structural elements here are distinctly different from what we saw earlier in the McHugh Complex. The strata are well bedded, and are preserved in coherent sections amenable to sedimentological study and bed-by-bed section logging. Faults are present but beds can be matched across them. Slaty cleavage is well developed; it is strongly refracted in the sandstone layers. Thin sandstone dikes locally cut the slates; they are nearly parallel to cleavage and have been interpreted to show that the Valdez Group was not well lithified when deformation commenced (Clark, 1981).

Around the point toward the west is a felsic dike belonging to the Sanak-Baranof near-trench intrusive suite. The outcrops are low and nearly covered with riprap. The dike shows a pronounced tectonic foliation; of hundreds of dikes that we have examined in the Chugach terrane, only a handful are foliated. The dike is undated but presumed to be about 54 Ma.

Although dikes are uncommon along Turnagain Arm, they are abundant not far to the north in the valley of Eagle River and to the south in the Hope-Sunrise mining district (directly across Turnagain Arm).

Return to vans, continue along Seward Highway.

Mile 19.4. Parking lot on the left just before the bridge over Bird Creek. Huge, new outcrops of Valdez Group, not yet studied.

Mile 22.1. The recently relocated railroad tracks on the right bury a classic exposure of a submarine channel and overbank succession in the Valdez Group, which was Stop 5 of Bradley et al. (1997). The best parts of the outcrop are gone.

Mile 22.9. Outcrops below the railroad tracks have yielded the Late Cretaceous (Maastrichtian) pelecypod *Inoceramus kusiroensis*. This corresponds to Stop 12 of Clark (1981).

Mile 24.8. Bird Point parking lot access, on the right. Bird Point is a thick channel sandstone of Valdez Group that juts out into the middle of Turnagain Arm. Gold was mined here between 1916 and 1919, from quartz-carbonate veins in the Valdez Group, exposed below high-tide level.

Miles 26.3-28.1. Huge new, unstudied exposures of Valdez Group.

Mile 31.3. Girdwood. The dead spruce forest was killed by saltwater incursion soon after the 1964 earthquake.

For those doing this field trip at a later date, a side trip hiking up to Crow Pass is highly recommended, to see small near-trench intrusions and gold mineralization related to the subduction of the Kula-Resurrection ridge. From the Seward Highway turnoff to Girdwood, drive northeast on the Alyeska Highway 2 miles and turn left onto the Crow Creek Road where the main road swings toward the east. It is about 5.8 miles farther to the parking lot for the Crow Pass Trail. Placer gold operations are still active near the point where the road narrows to a single lane. From the parking area, the trail switchbacks up a moderate grade to Crow Pass, about 3 miles. Along the way, you will see remarkably undeformed sedimentary rocks of the Valdez Group, and on the mountainsides above, light-colored felsic dikes and small intrusions of the Sanak-Baranof belt. At 1.7 miles from the parking area you will reach the ruins of the Monarch Mine, a lode gold mine that operated from 1906-1948. A number of characteristics of the gold-quartz veins can be seen at the entrance to the collapsed upper adit of the Monarch Mine, a short distance above the trail. The mineralized structure is a right-lateral strike-slip fault, marked by a brecciated zone between 40 cm and 2 m in width. Faulting probably occurred during and after mineralization, as is indicated by slickensides on quartz-vein surfaces. Visible free gold has been found in outcrop here. Haeussler et al. (1995) reported a $^{40}Ar/^{39}Ar$ plateau age of 54.1±0.1 Ma on white mica from the Crow Pass intrusion, uphill from the Monarch Mine, and a plateau age of 54.3±0.1 Ma on white mica from a gold-quartz vein at the nearby Jewel Mine. These dates provide a clear link between near-trench magmatism and gold mineralization. Orientations of all the structures in the Crow Pass area, including faults hosting the gold-quartz veins, are roughly 90° clockwise of equivalent structures along Turnagain Arm and in the Prince William Sound region. Thus, some vertical-axis rotation appears to have affected this region after gold mineralization. Two references are helpful: the geologic map and report by Park (1933), and a description of the mine workings by Hoekzema et al. (1986).

Mile 34.6. Old roadcut on the left side of the highway exposes Valdez Group cut by a felsic dike, which probably is around 54 Ma. If you wish to forego the $12 trip through the tunnel to Whittier and Stop 5, this is the next best thing.

Mile 41.5. Ruined buildings, a casualty of the 1964 earthquake, in the settlement of Portage. After the earthquake, trains could run here only at low tide. A pilot car would go first, checking that the tracks were okay, and then the train would proceed.

During the 1964 there was about 2 m of subsidence here, which killed the large trees. The small alders and birch growing among the dead trees indicates that post-seismic rebound has been sufficient for salt-water intolerant plants to live here again.

Mile 42.9. Turn left on road to Whittier.

Mile 49.6. Whittier tunnel toll booth. After paying the toll, vehicles will line up at the staging area for tunnel. The tunnel is one-way and also serves trains. For travel to Whittier (in May), the tunnel is open every hour on the half hour; for the return from Whittier, it is open every hour on the hour. We will aim for the 2:30-2:45 PM slot to Whittier and the 4:00-4:15 PM slot for the return.

Mile 52.5. Leave tunnel.

Mile 53.5 (approx). Right end of a half-mile-long roadcut of Valdez Group.

STOP 5. NEAR-TRENCH INTRUSION IN WHITTIER

This roadcut is the best road-accessible exposure of a near-trench intrusive in south-central Alaska. It is a fine-grained felsic body, perhaps 10-20 meters across, intruding the Valdez Group. U-Pb zircon geochronology so far has been unsuccessful due to inheritance, but judging from other ages from this portion of the Sanak-Baranof belt, it is probably about 54 Ma.

If time permits, we will also stop elsewhere along this huge roadcut to see the Valdez Group, while waiting for the next tunnel opening for outbound traffic.

The 1964 M9.2 earthquake unleashed numerous submarine landslides. The largest here was likely located in the northwestern corner of the bay. Wave run-up of 105 feet was measured on the cliffs directly across the bay. The locally generated tsunami waves in Passage Canal caused extensive damage of the harbor and port facilities (Kachadoorian, 1966).

Mile 54. Turn around at the east end of the roadcuts. Retrace route through the tunnel to Seward Highway in Portage, about 11 miles. Turn left on Seward Highway toward Seward, about 80 miles. From here to Seward, all outcrops are sandstone and slate of the Upper Cretaceous Valdez Group.

BOAT LOG FOR DAY 2

Seward lies at the head of Resurrection Bay, a fjord nearly 1,000 m deep that was carved by outlet glaciers of a large icecap to the north in the Kenai Mountains. Seward was badly damaged during the 1964 earthquake, and 13 people were killed (Lemke, 1966). Strong ground motion lasted 3 to 4 minutes. Seward was hit by tsunamis generated by the earthquake and also by local submarine landslides. A strip of waterfront 15 to 300 meters wide slid into the bay. When the submarine slide let go, it initially caused water level to drop; then the wave popped back up and flooded the town. Oil tanks ruptured and the floating oil caught fire. A train that included 40 filled oil tankers caught fire in a chain reaction of exploding cars. The first seismic sea wave arrived about 25 minutes after the earthquake. Wave run-up was about 10 m; areas as far as 2 km inland were inundated.

Until our first stop, we will be flanked on both sides of the bay by Valdez Group.

STOP 6. THUMB COVE

Spectacular shoreline exposures of pillow lavas. The unexposed contact between pillow lavas to the east and sedimentary rocks to the west follows Likes Creek, which enters the sea here on the north side of the cove. Kusky and Young (1999) interpreted the sedimentary section to depositionally overlie the pillow lavas. Miller (1984), Nelson, Miller, et al. (1987), and Nelson et al. (1989) interpreted this as a fault contact.

STOP 7. HUMPY COVE

The contact between the sheeted dikes and the pillow lavas is exposed at the head of the cove. The contact between the pillow lavas and turbidites is interpreted by some as depositional and by others as a fault; lithologies on both sides of the contact are fracture cleaved. Several copper prospect pits, dating from the 1920's, are visible. Economic massive-sulfide deposits are known from broadly equivalent rocks in Prince William Sound but not from the Resurrection Peninsula ophiolite.

STOP 8. FOX ISLAND

The western part of the island is underlain by metasedimentary and metavolcanic rocks (upper greenschist facies). These lie in thrust contact ("Fox Island shear zone" of Kusky and Young, 1999) over lower grade, mid-greenschist facies sedimentary rocks of the eastern part of the island. The latter include tuffs that so far have defied attempts at U-Pb dating. This is unfortunate because it would likely settle the affiliation of these rocks: Miller (1984), Nelson, Miller, et al. (1987), and Nelson et al. (1989) mapped them as Upper Cretaceous Valdez Group whereas Kusky and Young (1999) interpreted them as uppermost Paleocene to lowermost Eocene (post-57-Ma).

STOP 9. HIVE ISLAND

The Hive Island pluton is exposed on two islands that guard the entry to Resurrection Bay. It is a biotite granodiorite, dated at 53.4±0.4 Ma ($^{40}Ar/^{39}Ar$ biotite). Zircons from a U/Pb sample from this pluton, collected by Peter Haeussler (USGS), are plagued by inheritance (R. Friedman,

written communication, 2005). Kusky and Young (1999) believed that the projected map trace of the Fox Island shear zone is truncated by the pluton, thus constraining the age of subduction-accretion of the ophiolite.

STOP 10. CAPE RESURRECTION

Pillow lavas, locally interbedded with argillite and siltstone. The cliffs are a kittiwake rookery.

STOP 11. DRIFTWOOD BAY

Layered gabbro to the south and sheeted dikes to the north.

STOP 12. KILLER BAY

This is the largest exposure of plagiogranite in the Resurrection Peninsula ophiolite. A U-Pb zircon age of 57 Ma was reported by Nelson et al. (1989), but supporting details were never published and resampling has not yielded datable material.

OPTIONAL STOP, DAY 2, RETURN TRIP TO ANCHORAGE

STOP 13. GOLD PANNING AT SIXMILE CREEK

If time allows on the drive from Seward to Anchorage, we will stop at Sixmile Creek, part of which is open to recreational gold panning. To get to the panning site on Sixmile Creek, we will take the Seward Highway north to Hope Junction. From there we turn left toward Hope, and park at mile 2.3. A pamphlet by Huber and Kurtak (2001) describes in detail access to this site and others on the Kenai Peninsula that are open to recreational gold panning. Joe Kurtak, of the U.S. Bureau of Land Management, will help out here during the May, 2006 field trip. If you are doing this trip on your own, we strongly recommend getting a copy of the Huber and Kurtak (2001) pamphlet to ensure you choose a legal panning site.

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