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Subduction initiation, subduction accretion, and nonaccretion, large-scale material movement, and localization of subduction megaslip recorded in Franciscan complex and related rocks, California

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ABSTRACT

The Franciscan complex of California provides the ideal field laboratory to examine the rock record of subduction. This field trip guide describes a two-day field trip of the 2013 Geological Society of America Cordilleran Section Meeting. The field stops include a stop along the Panoche Road in the southern Diablo Range, and four in the San Francisco Bay region: one on the southern margin of the California State University East Bay campus in the Hayward Hills, one, with additional optional stops, at El Cerrito quarry north of Berkeley, one at Ring Mountain on Tiburon Peninsula, and the final stop at Rodeo Cove of the Marin Headlands. The geology seen at these stops provides insight into subduction initiation processes, subduction accretion and subduction erosion, accretion of ocean plate stratigraphy, mélangé evolution, multiple burial-exposure cycles, exhumation of high-pressure metamorphic rocks, localization of subduction megathrust slip, large-scale subduction complex architecture, and spreading ridge deformation. The field trip stops will captivate visitors with their scenic beauty as well as their interesting geology.

INTRODUCTION

Many regard the Franciscan complex of coastal California (Fig. 1) as the type subduction complex (Hamilton, 1969), and a benchmark locality for high-pressure–low-temperature (HP-LT) metamorphic rocks (Ernst, 1970), as well as one of the premier localities of mélangé (Hsü, 1968; Cloos, 1982, 1985; Raymond, 1984; Cowan, 1985). Accordingly, the Fran-

ciscan complex and related rocks represent one of the world's finest natural laboratories in which to study convergent plate-margin processes. This field trip will view a spectrum of outcrops that provide the rock record for past and emerging studies. Here, I will summarize some aspects of Franciscan complex geology before describing the geology seen at the various stops and the relevance of this geology to subduction and orogenic processes. Many competing proposals exist for

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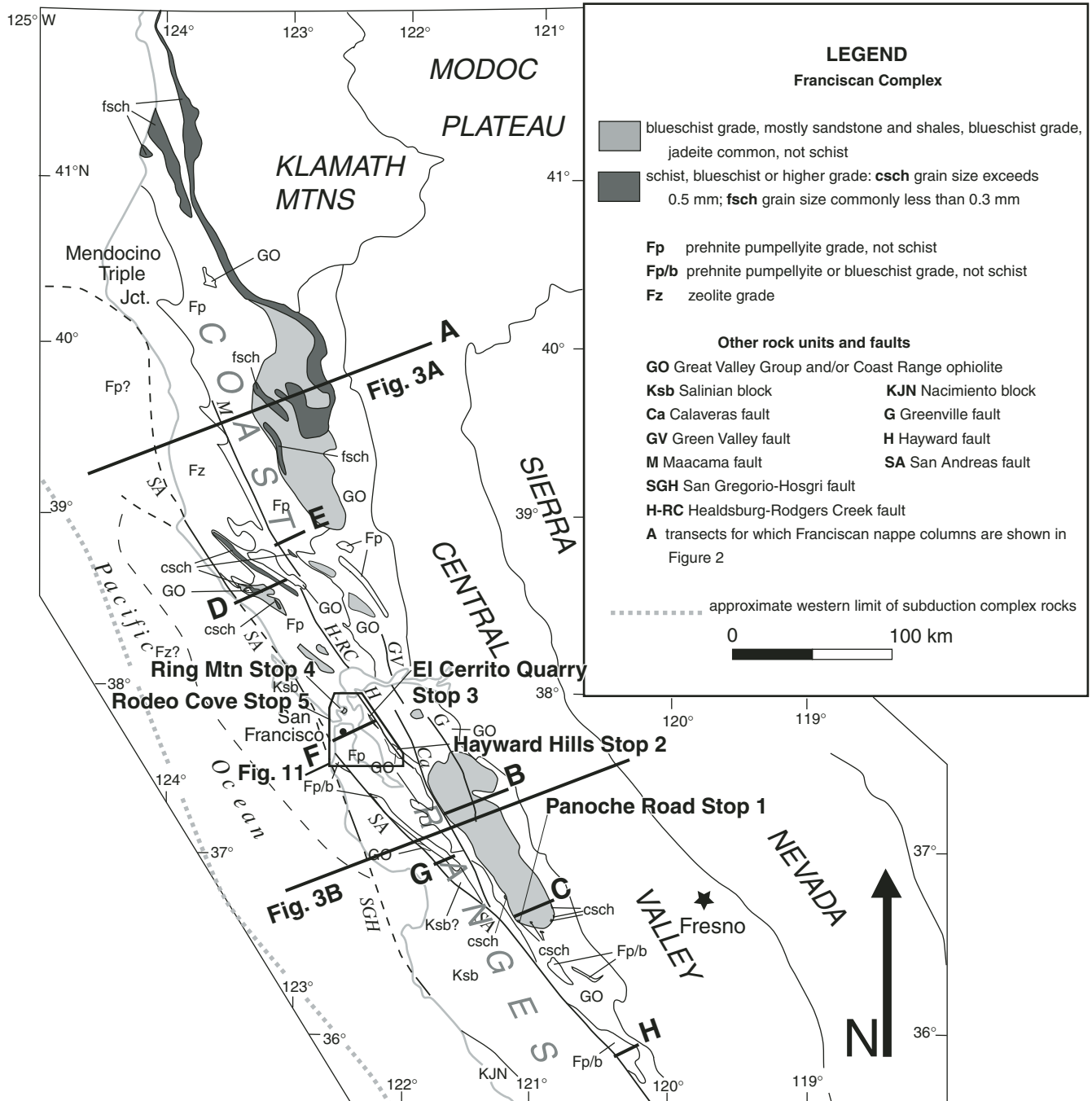


Figure 1. Distribution of Franciscan and other basement rocks of central and northern California, showing Franciscan rocks of different metamorphic grades and field trip stop locations. Figure is modified from Wakabayashi (1999a).

the various features of the Franciscan complex, and I remind the reader that this guide and my presentation at various field trip stops reflect my own opinions, which may differ markedly from those of others. Although I attempt to summarize a spectrum of views in this guide, I fear that this presentation will be far from complete in this regard, so I apologize in advance for any omissions.

The Franciscan complex of California formed during a period of over 140 m.y. of continuous east-dipping subduction in coastal California from ca. 165 Ma to ca. 20 Ma, and most of the rocks of the Franciscan complex were metamorphosed and accreted between ca. 165 Ma and ca. 50 Ma (e.g., Blake et al., 1988; Wakabayashi, 1992a; Wakabayashi and Dumitru, 2007; Ernst, 2011; Dumitru et al., 2013). At least 25% of the exposed

rocks have undergone HP-LT blueschist- or higher-grade metamorphism as a result of burial by partial subduction (Terabayashi and Maruyama, 1998; Wakabayashi, 1999a; Ernst and McLaughlin, 2012). Franciscan subduction terminated with ridge subduction, triple junction migration, and conversion to a transform plate boundary (the dextral San Andreas fault system), rather than by collision (e.g., Atwater, 1970; Furlong, 1984; Wakabayashi, 1996).

The Coast Range ophiolite, consisting of serpentinitized ultramafic rocks, gabbro, basalt, and other plutonic and volcanic rocks (Hopson et al., 1981, 2008), structurally overlies the Franciscan complex and is depositionally overlain by well-bedded sandstones and shales of the Great Valley Group, which are generally coeval with the clastic sedimentary rocks of the Franciscan complex (Dickinson, 1970). In contrast to the Franciscan complex, burial metamorphism of the Great Valley Group is limited to zeolite-facies metamorphism in the lower Great Valley Group (Dickinson et al., 1969), although some olistostrome units at or near its base contain high-pressure metamorphic blocks identical to those found in the Franciscan complex (Hitz and Wakabayashi, 2012) (see stop 2). The Coast Range ophiolite underwent seafloor hydrothermal metamorphism ranging from zeolite grade to greenschist or higher grade, but burial metamorphism does not exceed zeolite grade (Evarts and Schiffman, 1983).

The Coast Range ophiolite exhibits island-arc chemistry and is thought by many to have formed over a nascent subduction zone (Shervais and Kimbrough, 1985; Stern and Bloomer, 1992; Shervais, 1990, 2001; Giaramita et al., 1998), although Hopson et al. (1981, 2008) proposed a mid-ocean-ridge origin for most of the ophiolite, with a later stage of arc volcanism. From west to east, the three subparallel geologic provinces of the Franciscan complex, Great Valley Group, and the Sierra Nevada batholith (Fig. 1) represent, respectively, the subduction complex, forearc basin deposits, and the magmatic arc of an ancient arc-trench system associated with eastward subduction (Dickinson, 1970).

Franciscan lithologies are predominantly detrital sedimentary rocks (mainly sandstones or “graywackes”; these two terms will be used interchangeably in this guide) with subordinate serpentinite, basaltic volcanic rocks and chert, and minor limestone (Bailey et al., 1964; Blake et al., 1988; Coleman, 2000; Wakabayashi, 2004a, 2012a). The detrital rocks are primarily offscraped and underplated trench-fill sediments (Dickinson, 1970). The volcanic rocks represent remnants of seamounts (ocean-island basalt [OIB]), other oceanic rises, and mid-ocean-ridge-generated basalts (MORB), and the pelagic rocks represent their cover strata, components of the upper part of the subducted oceanic plate offscraped and underplated at the trench (Hamilton, 1969; MacPherson, 1983; Shervais, 1990). Some volcanic and pelagic rocks in the Franciscan complex originated from olistostrome blocks shed from the upper plate of the subduction system, or from previously accreted and exhumed Franciscan material (MacPherson et al., 1990; Erickson, 2011; Wakabayashi, 2012a). Some Franciscan serpentinites may represent intact sheets scraped off of the subducting plate (Coleman,

2000; Wakabayashi, 2004a; Prohoroff et al., 2012), whereas other Franciscan serpentinites (serpentinite mélanges) appear to have originated from serpentinite-rich debris flows shed into the trench, possibly from the upper plate (Wakabayashi, 2012a) (see stops 1, 2, and 4).

Franciscan rocks have been traditionally classified as internally “coherent” mappable bodies (not block-in-matrix) or mélange units that consist of a matrix with included blocks (e.g., Maxwell, 1974; Cowan, 1974; Blake et al., 1984, 1988; Wakabayashi, 1992a), although the distinction may be more difficult to apply than once thought (Wakabayashi, 2012b), as we will see on this field trip, particularly at stop 1.

Regional Franciscan structure includes a stack of refolded thrust nappes made up of both coherent and mélange units (Blake et al., 1984; Seiders, 1991; Wakabayashi, 1992a, 1999a) (Figs. 2 and 3). The map pattern of belts of similar Franciscan units (e.g., Jennings, 1977) that trend northwest-southeast is a consequence of the folding of the nappe stack about subhorizontal fold axes, followed by erosion. Although postsubduction folding, dip-slip faulting, and strike-slip faulting associated with the transform plate boundary (San Andreas fault system) have affected the California Coast Ranges, the regional nappe structure appears to reflect accretionary complex processes (Wakabayashi, 1992a). The regional cross sections of Figure 3 depict the Franciscan complex at two different latitudes with postsubduction San Andreas fault system dextral slip restored. These cross sections show that the along-strike variation in the various Franciscan regions, illustrated in the contrasting nappe stacks in Figure 2, is still significant after restoration of postsubduction strike-slip faulting (Wakabayashi, 1999a). In addition to the thrust nappe architecture of the Franciscan complex, a first-order regional feature is the east-vergent thrust fault that locally places Franciscan units over Great Valley Group and Coast Range ophiolite, along the eastern margin of the Coast Ranges (Fig. 3) (Wentworth et al., 1984; Unruh et al., 1991, 1995; Wakabayashi and Unruh, 1995). This arc-vergent faulting began in the Late Cretaceous and was reactivated in the late Cenozoic transform regime (Wentworth et al., 1984; Unruh et al., 1995; Wakabayashi and Unruh, 1995).

Many previous papers have divided the Franciscan complex into the Eastern, Central, and Coastal belts, following Berkland et al. (1972), and the three-part division is useful in the northern Coast Ranges (e.g., Maxwell, 1974; Blake et al., 1988; Terabayashi and Maruyama, 1998; Ernst and McLaughlin, 2012; Dumitru et al., 2013). Unfortunately, this scheme does not work well for other parts of the Franciscan complex, such as the Diablo Range and San Francisco Bay area, and attempts to apply this nomenclature to those areas have resulted in much confusion, especially to researchers from outside of California (for more details, see Wakabayashi, 2011a, p. 120–121). Accordingly, I divide the Franciscan complex in this field trip guide, and in my other papers, on the basis of metamorphic grade and accretionary age for individual nappes. The comparison and correlation of the belt nomenclature to accretionary age and metamorphic grade throughout the complex are illustrated in Figures 1 and 2 for the

belt designations given for the nappes in column A in Figure 2 (northern Coast Ranges, where originally defined).

Metamorphism in most coherent units ranges from zeolite facies to epidote blueschist (Blake et al., 1988). Exceptions are rare high-grade slabs that display eclogite, amphibolite, and garnet-amphibolite assemblages (Wakabayashi and Dumitru, 2007). Blocks-in-mélange span the same range in metamorphism as the coherent units (e.g., Blake et al., 1988). It is useful to define “high-grade” metamorphic rocks to include blocks-in-mélange (referred to as “high-grade blocks”) and their rare coherent equivalents, to include amphibolites, eclogites, and coarse blueschists, for these have special significance and have drawn attention out of all proportion to their volumetric abundance.

Collectively, the high-grade metamorphic rocks make up far less than 1% of Franciscan metamorphic rocks, but they are very widely distributed, cropping out at hundreds of localities along the length of the subduction complex (Coleman and Lanphere, 1971). Many, if not most, eclogites and essentially all coarse blueschists have relics of earlier amphibolite-facies metamorphic assemblages (Moore and Blake, 1989; Wakabayashi, 1990). High-grade rocks are entirely metabasites with minor metacherts, whereas lower-grade coherent metamorphic rocks are mostly metagraywackes and metashales, with lesser proportions of metabasites and metacherts (Blake et al., 1988; Coleman and Lanphere, 1971).

Many high-grade blocks are partly encased in “rinds” composed of minerals such as actinolite, chlorite, talc, and phengite: This rind apparently formed by metasomatic reaction between the block and surrounding ultramafic rocks under conditions of decreasing metamorphic grade (e.g., Coleman and Lanphere, 1971; Moore, 1984; Catlos and Sorensen, 2003). High-grade metamorphism evolved along a counterclockwise pressure-temperature (P - T) path (P on the positive y -axis) (Wakabayashi, 1990; Krogh et al., 1994; Tsujimori et al., 2006; Page et al., 2007), whereas metamorphism in lower-grade metamorphic rocks took place under a mildly clockwise or hairpin P - T trajectory (Maruyama and Liou, 1988; Maruyama et al., 1985). Franciscan metamorphic rocks lack a thermal overprint, suggesting exhumation to shallow crustal levels prior to subduction termination (Ernst, 1988). The ridge subduction and conversion to a transform margin were associated with the termination of subduction and raised geothermal gradients in the former subduction complex (Furlong, 1984). Rocks at depths of 10 km or more have probably been overprinted with greenschist- and higher-grade metamorphism assemblages within the past 20 m.y. (Cloos and Dumitru, 1987; Wakabayashi, 1996, 2004b). The lack of thermal overprint on Franciscan metamorphic rocks thus reflects the small amount of exhumation since subduction termination (Dumitru, 1989), as well synsubduction exhumation of the rocks at the surface today.

The high-grade rocks are the oldest metamorphic rocks in the Franciscan complex, yielding Ar/Ar hornblende ages of 157–168 Ma (Ross and Sharp, 1988; Wakabayashi and Dumitru, 2007; Shervais et al., 2011), Lu-Hf garnet ages of 147–169 Ma

(Anczkiewicz et al., 2004), and lower-temperature metamorphic or cooling ($^{40}\text{Ar}/^{39}\text{Ar}$ or K/Ar white mica) ages of 138–159 Ma (Wakabayashi and Deino, 1989; Wakabayashi and Dumitru, 2007; Ukar et al., 2012). The age of high-temperature metamorphism in the high-grade rocks is slightly younger than the 165–172 Ma crystallization age of the Coast Range ophiolite (Hopson et al., 1981, 2008; Shervais et al., 2005).

Ages of lower-grade blueschist metamorphism range from ca. 135 Ma to 80 Ma (Wakabayashi and Dumitru, 2007). Peak temperatures of Franciscan metamorphism decrease with decreasing age from amphibolite high-grade blocks and slabs with temperatures of over 600 °C (Wakabayashi, 1990), to 300–350 °C for epidote blueschists (Blake et al., 1988; Maruyama and Liou, 1988) at ca. 154 Ma (Wakabayashi and Dumitru, 2007), to 150–250 °C for lawsonite blueschist-facies rocks (Ernst, 1993; Maruyama et al., 1985) at 80–120 Ma (Wakabayashi and Dumitru, 2007), and ultimately to zeolite and prehnite-pumpellyite temperatures of <250 °C (Blake et al., 1988) for the last 95 Ma. The overlapping age ranges for blueschist-facies and prehnite-pumpellyite-facies rocks probably result from along-strike variation in accretion and exhumation (Figs. 2 and 3).

Estimates of metamorphic pressures of high-grade metamorphic rocks range from ~5 kbar for the lowest- P amphibolites (Wakabayashi, 1990; but see statements in description of stop 4) to 20–25 kbar for eclogites (Tsujimori et al., 2006; Page et al., 2007), with lower-grade blueschist giving pressures of ~6 to >10 kbar (Brown and Ghent, 1983; Maruyama et al., 1985; Ernst, 1993; Ernst and McLaughlin, 2012), whereas sub-blueschist-grade rocks were metamorphosed at pressures at 4 kbar or less (Blake et al., 1988). Metamorphic ages of Franciscan rocks generally approximate the time of subduction, because exhumation of these rocks occurred while subduction and refrigeration were ongoing (Ernst, 1988).

The high-grade rocks may have formed from the upper part of oceanic crust subducted and accreted beneath hot suboceanic upper-mantle material at the initiation of subduction, based on their old age, their age relative to the Coast Range ophiolite, lithology, and counterclockwise P - T evolution, whereas lower-grade Franciscan rocks formed during subsequent subduction and accretion (Cloos, 1985; Wakabayashi, 1990). The high-grade rocks thus are analogous to sheets of high-grade rocks found beneath ophiolites called metamorphic soles (Platt, 1975; Wakabayashi, 1990; Wakabayashi and Dilek, 2000). High-grade rocks in the Franciscan complex, coherent and blocks-in-mélange, have geochemical characteristics reflecting island-arc protoliths (Saha et al., 2005; Wakabayashi et al., 2010) (see additional discussion on stop 4).

Lower-grade coherent volcanic rocks of the Franciscan complex have MORB or OIB chemistry (Shervais and Kimbrough, 1987; MacPherson et al., 1990; Shervais, 1990; Ghatak et al., 2012), whereas some lower-grade blocks-in-mélange (fine-grained blueschist and lower in grade) have island-arc chemistry in addition to others that have OIB or MORB chemistry (MacPherson et al., 1990). That lack of coherent low-grade

meta-igneous rocks of island-arc chemistry suggests that the occurrence of these rocks as blocks-in-mélange resulted from submarine debris flows derived from the upper plate (MacPherson et al., 1990; Wakabayashi, 2012a; see also stops 1, 2, and 4).

At least four mechanisms have been proposed for exhumation of the coherent high-pressure rocks of the Franciscan complex (and other coherent high-pressure rocks of the world): (1) buoyancy relative to the mantle beneath which the rocks were subducted (Ernst, 1970), (2) thrust faulting and erosion (Ring and Brandon, 1994), which requires two stages of thrust faulting to produce the metamorphic contrasts seen (Cowan et al., 1989), (3) synsubduction extension within the subduction wedge (Platt, 1986), and (4) cross-sectional extrusion, wherein a high-pressure nappe is exhumed relative to flanking rocks by a thrust fault below it and a normal fault above it (Maruyama et al., 1996). Future structural and geochronologic studies in the Franciscan complex promise to further explore this issue. Exhumation of high-grade blocks will be discussed in the description for stop 4.

Whereas metamorphic ages closely approximate the age of accretion of some metamorphic units, most of the Franciscan complex is unsuitable for radioisotopic dating owing to the lack of datable metamorphic minerals. The depositional ages of the graywackes may approximate accretionary timing because they were deposited in the trench shortly before subduction (Wakabayashi, 1992a). Determination of depositional ages once relied upon very sparse fossils, some of which have been recently shown to be reworked (Dumitru, 2012), whereas U-Pb detrital zircon geochronology has significantly enhanced and refined this fragmentary geochronologic record (e.g., Ernst et al., 2009; Snow et al., 2010; Dumitru et al., 2010, 2013; Dumitru, 2012). Although the youngest detrital zircons from a given rock represent a maximum age of deposition, these appear to closely approximate the age of deposition, owing to the similarity of metamorphic ages and youngest detrital zircon populations, and the continuous activity of the inboard magmatic arc (e.g., Dumitru et al., 2010).

The ages of accretion for most of the Franciscan complex, determined by metamorphic ages and detrital zircon chronology, become younger structurally downward, ranging from ca. 165 Ma for the structurally highest and rare high-grade coherent rocks to ca. 35 Ma (Ernst et al., 2009; Snow et al., 2010; Dumitru et al., 2010, 2013; Dumitru, 2012) (Fig. 2). The structurally downward younging of accretion ages reflects progressive offscraping of material from the downgoing plate, as proposed by Karig and Sharman (1975) for general accretionary complex evolution. Very little accretion took place between subduction initiation at ca. 165 Ma and ca. 120 Ma in the northern Coast Ranges (Dumitru et al., 2010), 100 Ma in the San Francisco Bay area (Wakabayashi, 2012a), and ca. 115 Ma in the southern Diablo Range (this study; revising Wakabayashi, 2012a; see stop 1 discussion). This suggests a long period of nonaccretion or subduction erosion following initiation of subduction.

The preservation of small sheets of coherent high-grade material at the highest structural level suggests general nonac-

cretion rather than subduction erosion, because the high-grade sheet probably had an original structural thickness of less than 500 m (Wakabayashi, 2012a). However, the possibility of accretion followed by subduction erosion cannot be dismissed for this period in Franciscan history. Nonetheless, the amount of forearc removal by subduction erosion associated with Franciscan subduction appears vastly less than the Catalina Schist of offshore southern California, where the subduction complex appears juxtaposed against the base of the arc (Grove et al., 2008).

Whereas the accretionary age recorded or approximated by metamorphic ages and depositional ages of clastic sedimentary rocks gets younger structurally downward, the age of formation of the oceanic rocks (pelagic sedimentary rocks and volcanic rocks) in the subduction complex does not, as illustrated in [Figure 4](#). Because of the difference in the tectonic significance of the age of clastic sedimentary rocks versus oceanic rocks in a subduction complex, depiction of convergent plate-margin assemblages with stratigraphic columns based on formational age may lead to misunderstanding of the associated tectonic processes. Stop 5 visits the part of the Franciscan complex, the Marin Headlands, that displays the largest disparity between accretionary and formational age.

Based on plate-motion reconstructions (Engebretson et al., 1985), the >100 m.y. period of subduction recorded by the exposed Franciscan complex indicates that more than 10,000 km of oceanic plate subducted during the accretion of the subduction complex. The Franciscan complex should preserve the >10,000 km of slip on paleosubduction megathrust horizons between the nappes that accreted at different times ([Figs. 2 and 5](#)). Because mélanges commonly intervene between coherent accreted nappes of the Franciscan complex (Maxwell, 1974; Blake et al., 1984; Wakabayashi, 1992a), it comes as little surprise that Franciscan mélanges have been equated to the paleomegathrust horizons, and their block-in-matrix fabric (“tectonic mélange”) has been attributed to tectonic dismemberment during extreme displacement and strain as a “subduction channel” (e.g., Cloos and Shreve, 1988a, 1988b). In contrast, others have proposed that the block-in-matrix fabric formed as a result of submarine sliding (“sedimentary mélange”), creating a sedimentary breccia that was subsequently overprinted by tectonic strain (e.g., Cowan, 1978; Aalto, 1989; Wakabayashi, 2011a, 2012a).

Based largely on relationships at El Cerrito quarry (stop 3; see stop description), I have proposed (2011a, 2011b, this study) that most paleomegathrust displacement is accommodated along the upper contact of, rather than within, mélanges ([Fig. 5](#)). However, it is unclear how representative the El Cerrito quarry exposures are of general Franciscan accretionary structures.

It is likely that the Franciscan complex includes mélanges formed by at least three different processes, including tectonic strain, submarine sliding, and diapirism. Becker and Cloos (1985) identified a diapiric mélange in the San Simeon area along the central California coast. On our field trip, we will see several examples of what have interpreted as sedimentary mélanges in the Franciscan complex (stops 1, 3, and 4) (Wakabayashi,

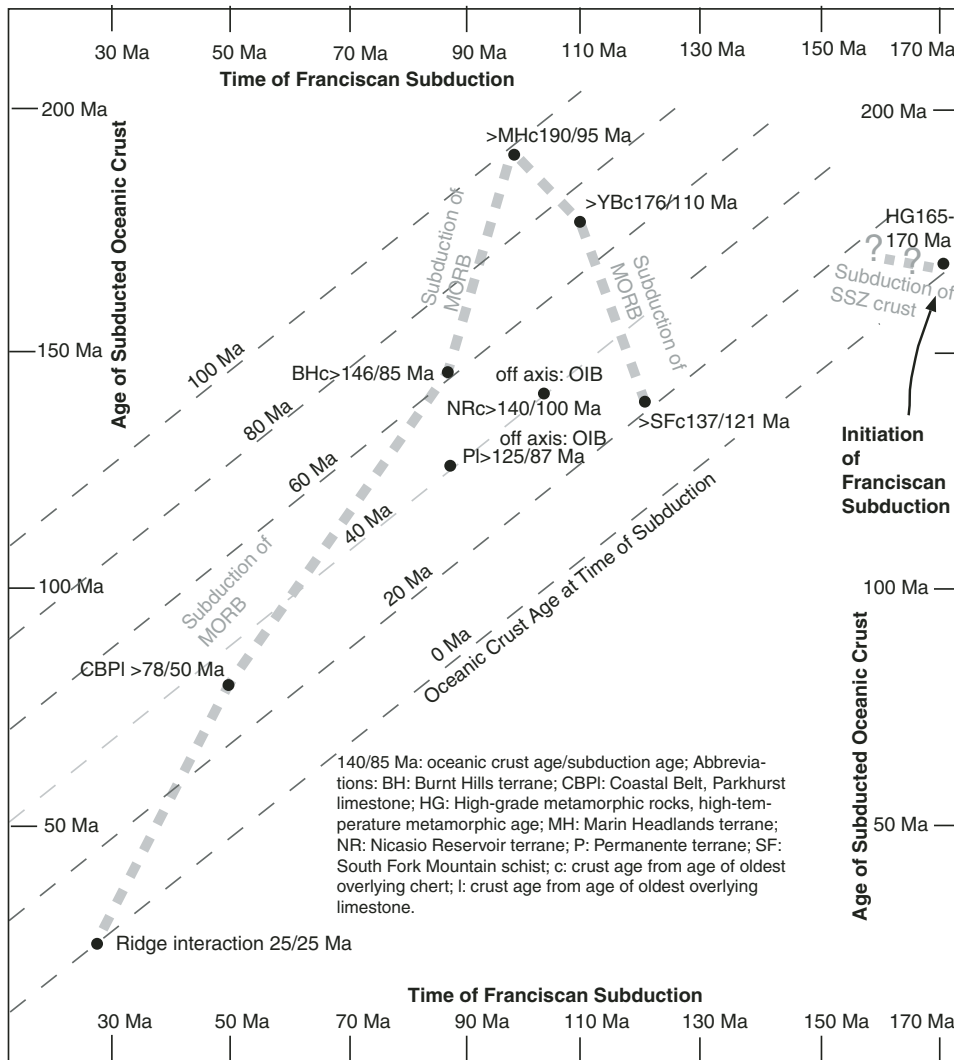


Figure 4. The age of subducted oceanic crust versus time of subduction at the Franciscan trench (thick dashed gray line on diagram), updated from Wakabayashi et al. (2010), with some adjustment to the Coastal Belt and Yolla Bolly accretion ages. An earlier plot illustrating this concept was done by Murchey and Blake (1992). MORB—mid-ocean-ridge basalt; SSZ—suprasubduction zone.

2011a, 2011b, 2012a, with some updates presented here), a sedimentary mélange in the Great Valley Group (stop 2) (Hitz and Wakabayashi, 2012), a tectonic mélange or tectonic mélanges in the Franciscan complex (stop 5), which formed during imbrication of a coherent unit within the subduction complex (Meneghini and Moore, 2007) (Fig. 5), and a possible diapiric mélange (stop 5).

To conclude this introduction to the Franciscan complex, Figure 6 shows a series of cross-sectional cartoons illustrating a speculative tectonic history of the Franciscan complex, as presented in Wakabayashi (2012a). In this model, collisional termination of a pre-Franciscan west-dipping subduction zone resulted in initiation of east-dipping Franciscan subduction within nascent arc crust, now represented by the Coast Range ophiolite and the high-grade rocks, similar to earlier proposals of Moores (1970) and Schweickert and Cowan (1975). This model, however, conflicts with more widely held views on subduction initiation and generation and emplacement of arc-related ophiolites, including the Coast Range ophiolite. In the latter alternative models, east-

dipping Franciscan subduction initiated within mid-ocean-ridge-generated crust, and the Coast Range ophiolite formed early above the nascent Franciscan subduction zone as the slab hinge rolled back (e.g., Stern and Bloomer, 1992; Shervais, 2001). In one of these alternatives, the high-grade rocks formed as a result of ridge (mid-ocean-ridge) subduction after the initiation of Franciscan subduction (Shervais, 2001).

Initiation of subduction was followed by a long period of nonaccretion before significant accretion took place, based on the ages of accreted units (Dumitru et al., 2010; Wakabayashi, 2012a). The material delivered to the trench and accreted included a significant component of previously subducted and accreted Franciscan material that had been exhumed and delivered back to the trench in submarine landslides (olistostromes), along with a significant amount of material derived from the upper plate (Coast Range ophiolite and basal Great Valley Group) (Wakabayashi, 2011a, 2012a) (see stop 1 and 2 descriptions).

It is important to realize that the cartoons in Figure 6 are designed to illustrate processes that formed what we see at the

Paleomegathrust localized on upper mélangé contacts within accretionary prism

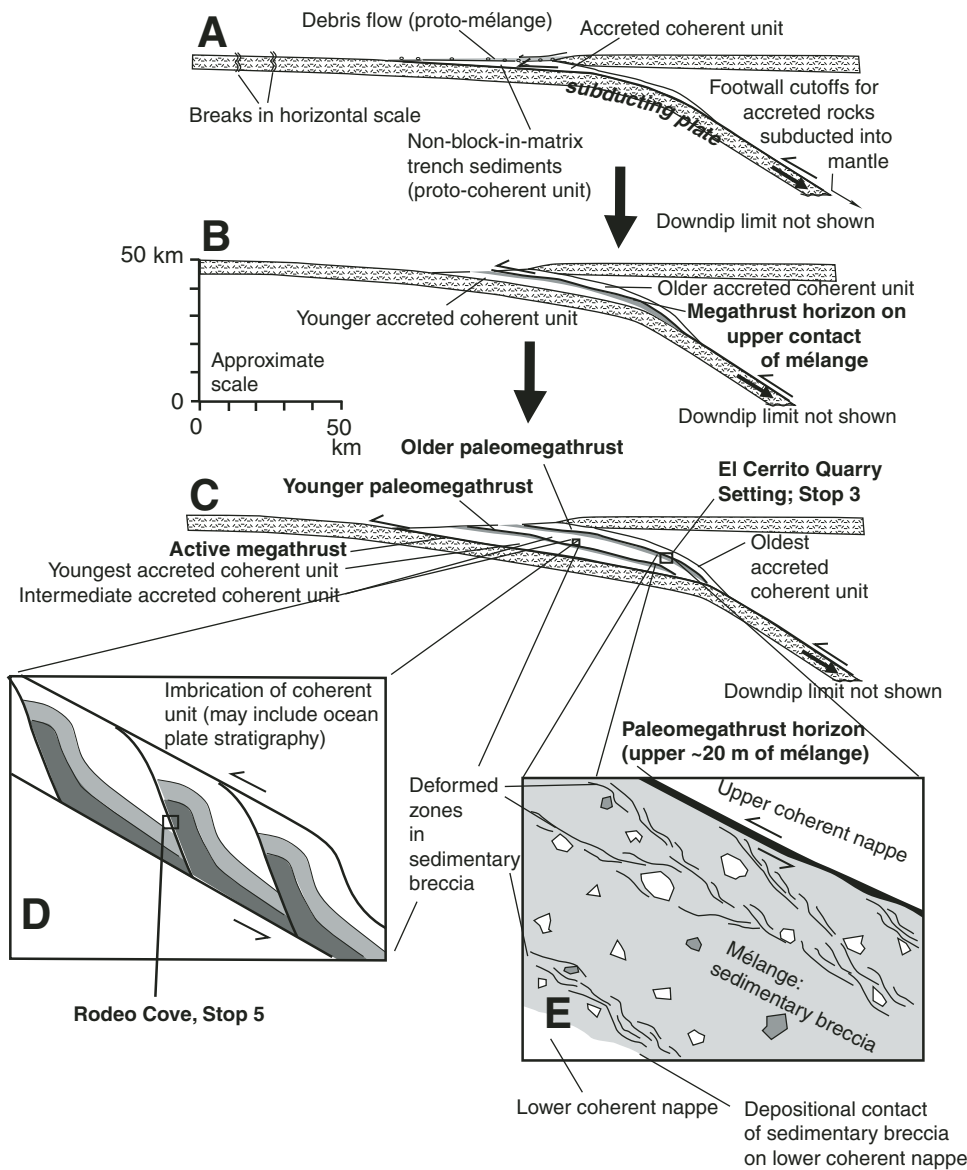


Figure 5. Cross-sectional cartoons showing subduction, accretion, and the preservation of paleomegathrust horizons within an accretionary complex. The sequence of frames from A to C shows how megathrust displacement with hundreds or thousands of kilometers can be accommodated on the upper contacts of mélangé zones. The mélangés themselves originate as submarine landslides (sedimentary breccia) and are variably deformed during subduction accretion. Inset in D shows a schematic of imbrication of a coherent (non-block-in-matrix) unit. Such imbrication can be recognized if the coherent unit consists of ocean plate stratigraphy, such as the basalt, chert, graywacke of the Marin Headlands (stop 5, Fig. 20), but it is difficult to recognize in coherent units made of graywacke and shale because of the lack of marker horizons. Inset in E shows schematic enlargement of the area adjacent to the paleomegathrust based on field relationships at El Cerrito quarry (stop 3). The subduction slip is accommodated in a 10–20-m-thick zone atop the mélangé, and the mélangé is a sedimentary breccia in depositional contact atop the lower coherent Alcatraz nappe. Local zones of deformation overprint the sedimentary features; whereas there is a general increase in deformation structurally upward, it is not progressive, for a deformed zone occupies some of the lower part of the mélangé near the basal contact.

current level of exposure in the Franciscan. Accordingly, the relatively small volume of accreted material in the Franciscan was probably associated with a much larger volume of accreted material at shallower levels (for example, compare to Hamilton, 1979), which has now been removed by erosion to the current exposure level of the subduction complex.

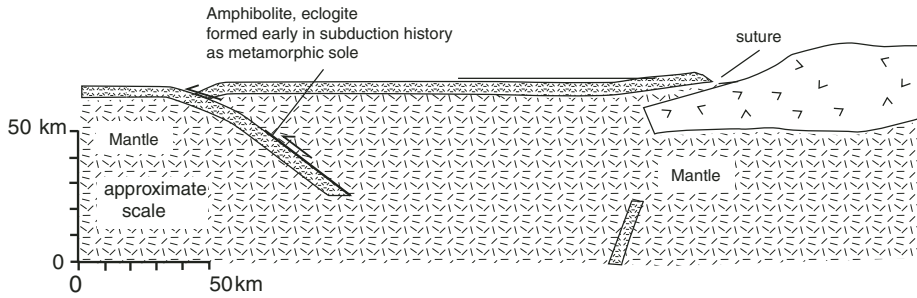
Stops 2, 3, and 4 have been described in previous field trip guides (Wakabayashi, 1989, 1992b, 1999c, 2005). A comparison with this guide illustrates a large change in the understanding and interpretation of the geology that has taken place in the last 20 or so years, both as a result of changes in understanding the geologic process in general, as well as acquisition of new data at the specific localities.

STOP DESCRIPTIONS AND ROAD LOG

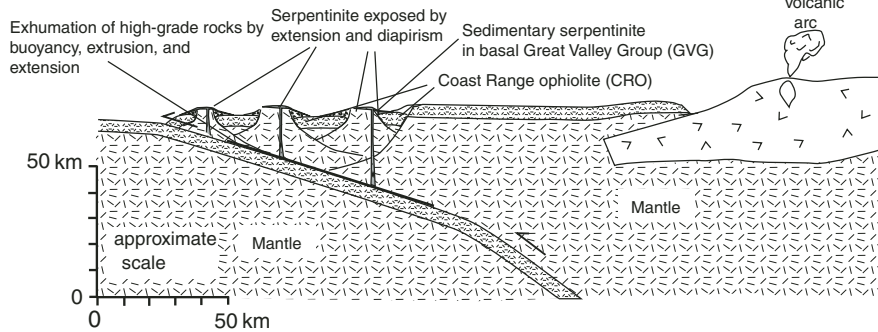
Drive from Fresno to Stop 1 on the Panoche Road (~92 Miles; 148 Kilometers)

From the junction of State Route (SR) 99 and 180, take State Highway 180 westbound (set odometer to 0.0). Just before reaching the town of Mendota, the highway curves right (northward). At 32.6 miles (52.5 km) at the southern edge of the town of Mendota, turn left (west) onto Belmont Avenue (following signs that says route to Interstate Highway 5). Drive westward 11 miles (17.7 km) (43.6 miles [70 km] past SR99) and turn right (north, T-intersection) at Fairfax Avenue. Drive north 2 miles (3.2 km)

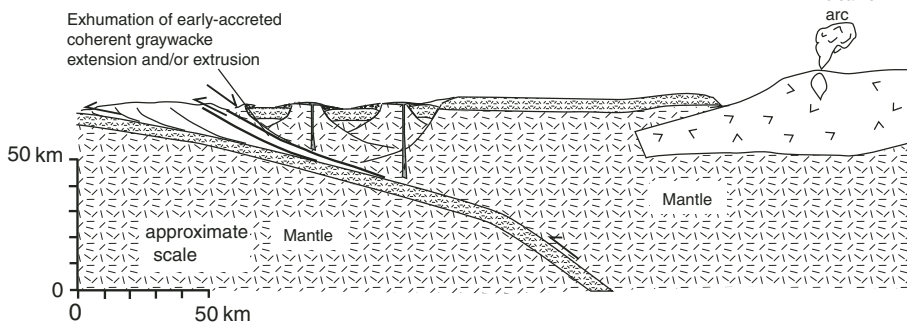
A Ca. 165 Ma subduction initiation



B Ca. 150-135 Ma early serpentinite sedimentation



C Ca. 130?-100 Ma early clastic accretion



D Ca. 100 Ma first preserved sedimentary serpentinite in Franciscan

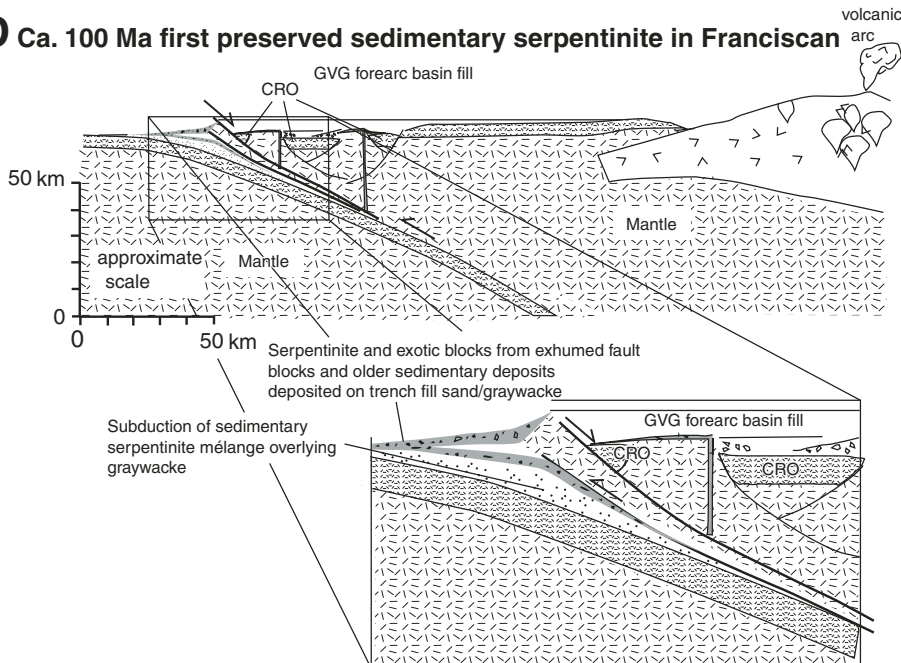


Figure 6. Schematic cartoons illustrating the tectonic evolution of the Franciscan complex. Cartoons represent tectonic history of the Franciscan complex in the Ring Mountain, Tiburon Peninsula area (stop 4), with subduction erosion between initiation of subduction at ca. 165 Ma and the first preserved accretion of the serpentinite mélangé and underlying graywacke at ca. 100 Ma. Note that at ca. 100 Ma, serpentinite and high-grade blocks are eroded and transported to the trench from exposed fault blocks, which may include coherent slices of high-grade material, as well as reworked sedimentary serpentinites that had been deposited on the forearc crust. Arc-derived clastic sediment (graywacke protolith) reaches the trench via submarine canyons out of the plane of this section. Figure is revised from Wakabayashi (2012a).

(45.6 miles [73.4 km] past SR99) and then turn left (west) onto Shields Avenue (County Road J-1, following signs to Interstate 5). Drive west on Shields Avenue and cross Interstate 5 after 9 miles (14.5 km) (54.6 miles [87.9 km] past SR99). After crossing Interstate 5, this road is called Mercy Hot Springs Road. Drive another 20.9 miles (33.6 km) past Interstate 5. The road climbs gradually up a drainage with abundant flights of terraces, and then a bit more steeply before heading through a narrow gap (this ridge is called Glaucophane Ridge after some rather large blocks or slabs of coarse blueschist that crop out on it), and descending to Panoche Valley, where it then heads southward across the valley and meets Panoche Road at a T-intersection (as noted above 20.9 miles [33.6 km] past I-5 and 75.5 miles [121.5 km] past SR99), where we turn right (westward) onto Panoche Road. Drive westward on this narrow country road for an additional 16.4 miles (26.4 km) to stop 1 (total distance from I-5 is 37.3 miles [60 km]; total distance from SR99 is 91.9 miles [147.9 km]). Just before reaching stop 1, note exposures of serpentinite on the right (northern) side of the road and pull to the right of the road and park just short of a repaired and regraded landslide. Parking area location: 36.65718°N latitude, 121.1167°W longitude.

Additional Geologic Notes for Drive along Panoche Road to Stop 1

Panoche Road crosses the Diablo Range, a regional-scale antiformal structure with the structurally low Franciscan subduction complex in the core, flanked by Great Valley Group forearc basin strata on the east and west (Fig. 1), with local Coast Range ophiolite remnants intervening between the Franciscan and Great Valley Group. The Franciscan core along the road consists of coherent metagraywackes of at least two different units as well as mélanges. The most extensive metagraywacke unit along the road is the structurally lowest unit (Wakabayashi and Dumitru, 2007), consisting of well-bedded sandstone and shale that lacks a strong foliation and cleavage. This unit appears to lack jadeitic clinopyroxene (or it may be very scarce), but lawsonite is common. The locally thinly bedded character of this unit, as well as fairly common detrital biotite suggest a correlation to the Burnt Hills terrane to the north, which has yielded Santonian-Campanian fossils (depositional age ca. 85 Ma) (Elder and Miller, 1993).

The Panoche Road crosses the Diablo Range not far north of the southern margin of Franciscan exposures. This area is noteworthy in part because of the presence of the largest high-grade coherent slabs in the Franciscan complex (Wakabayashi and Dumitru, 2007). Unfortunately, these slabs are on private land, so they are not a part of this field trip. The largest such body is the Willow Spring slab, 2.1 by 1.2 km in map dimension, which crops out ~1 km north of the Panoche Road near Panoche Pass (Wakabayashi and Dumitru, 2007). This slab is ~8 km east of stop 1 of our field trip. Another such unit, named the Antelope Creek slab by Wakabayashi and Dumitru (2007), is located a few kilometers east of the Willow Spring slab and has been the sub-

ject of several petrologic and geochronologic studies (Hermes, 1973; Ross and Sharp, 1988; Anczkiewicz et al., 2004).

Ernst (1965) published a study of the field geology and metamorphic petrology of Franciscan rocks along and flanking the Panoche Road. His work also included a color geologic map, which remains the most detailed and accurate map of this area. In the years since then, studies of Franciscan geology in the Diablo Range have shifted northward, including the studies of Ernst himself (e.g., Ernst, 1970, 1993; Ernst et al., 2009), so the Panoche Pass area has received comparatively little attention in the last four decades, with the exception of high-grade slabs near Panoche Pass and the recent work I have done on the rocks we will see at stop 1 (Wakabayashi, 2012a). Most Franciscan field trips in the Diablo Range have visited exposures along the much busier Pacheco Pass Road (e.g., Bennison et al., 1991). Whereas the Pacheco Pass Franciscan exposures are excellent, the Panoche Road exposures benefit from having far less traffic, making for a safer and quieter field trip experience.

Stop 1. Panoche Road at Western Margin of Franciscan Exposures of Diablo Range (36.65718°N, 121.11670°W)

This stop views blueschist and higher-grade Franciscan rocks on the western margin of the Diablo Range antiform near a contact with Great Valley Group forearc basin rocks (Fig. 7). Rocks viewed at the stop include exposures of interbedded jadeite-bearing graywacke and sedimentary breccias, with the latter grading into mélange with increasing strain. The breccia units include a full spectrum of exotic blocks and clasts, such as serpentinite, amphibolite, and abundant felsic volcanics. These exposures provide good evidence for the derivation of Franciscan mélanges with exotic blocks from sedimentary breccia. Wakabayashi (2012a) described this locality in detail, but this guide provides significant updates, particularly the description of outcrops of interbedded sedimentary breccia and metagraywacke north of the road, as well as the recognition of the possible Burnt Hills terrane sandstones (see following).

Because these outcrops are situated on the western margin of the Franciscan core of the Diablo Range antiform, these rocks should represent the structurally highest horizons in the Franciscan in this area. However, the contact with the Great Valley Group appears to truncate internal Franciscan structure. In addition, the foliation along the western exposures at this locality (Ernst, 1965; this study)—from the western part of the large serpentinite body (north-central part of Fig. 7) westward—dips northeast, so that the serpentinite is the structurally highest unit in this area, and it structurally overlies a stack of metavolcanic, metagraywacke, and breccia/mélangé units (Fig. 7). This stack may contain at least one major accretionary boundary (paleo-megathrust horizon). The structurally lowest of these contacts may place a metagraywacke slab (ms directly SW of the large mv exposure but north of the photo area delimited for Fig. 9 on Fig. 7) over the breccia, mélange, and graywacke, which make up the southwesternmost Franciscan exposures.

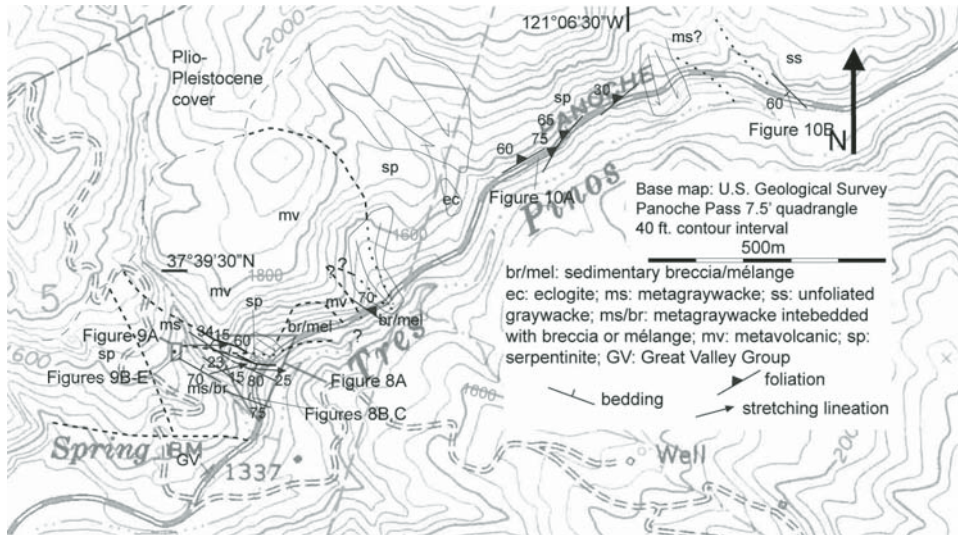


Figure 7. Geology near the western margin of Franciscan exposures along Panoche Road. Figure is revised from Wakabayashi (2012a).

Breccia matrix collected from along the road (from the outcrop of Fig. 8C) has yielded a maximum depositional age of 104 Ma, whereas the metagraywacke slab structurally above it has yielded a maximum deposition age of 114 Ma, based on U-Pb detrital zircon ages (T. Dumitru, unpub. data, 2011, personal commun.). The metavolcanic unit appears to be structurally above the metagraywacke slice. It is likely that this unit tectonically overlies the graywacke rather than being deposited on the latter as a large block. The contact of the metavolcanic unit over the graywacke may represent a major accretionary contact, or it is an ocean floor imbricate zone within an otherwise coherent unit (for example, as schematically shown in Fig. 5D; see also stop 5). Although the serpentinite appears sedimentary (see following), the nature of the contact between it and the (structurally or stratigraphically) lower metavolcanic unit cannot be discerned because lack of suitable exposures. The foliation within the eastern part of the serpentinite body dips northwest, indicating a synformal structure (Fig. 7). The unfoliated, thinly bedded sandstone at the eastern limit of the road exposures on Figure 7 dips steeply southwest, structurally beneath the serpentinite, but this unit does not resemble the foliated metagraywacke units west of the serpentinite body. In addition, the thinly bedded, unfoliated sandstone lacks jadeite, but it has neoblastic lawsonite, abundant detrital biotite, and resembles the Burnt Hills terrane, which has yielded fossils suggestive of a ca. 85 Ma depositional age. This indicates later accretion for this sandstone at structurally lower levels than the units west of the serpentinite. These relations suggest that the eastern contact of the serpentinite may be a late fault that cuts the Franciscan nappe stack.

Road cuts expose blueschist-facies sedimentary breccia, including clast-supported breccia (Figs. 8A, 8B, and 8C). The interbedding of sedimentary breccia and metagraywacke is not as apparent along the road-cut exposures as in a narrow south-flowing gully north of the road (Fig. 9; see location on Fig. 7). As seen in Figures 9C and 9D, metagraywacke grades into sedimen-

tary breccias, with an increase in the proportion of gravel-sized clasts. The hike up the gully is somewhat rugged, as is the hike to the ridge-top viewpoint from which photo Figure 9A was taken. Ticks are atrociously numerous in this gully during the spring. The gradation between sedimentary breccia (which includes a full range of exotic clasts) and deformed mélangé matrix viewed along the road and the interbedding of the sedimentary breccia and metagraywacke demonstrate the sedimentary origins of the block-in-matrix character of these mélanges.

The largest block of serpentinite within the breccia/mélangé units itself has blocks of amphibolite and garnet-amphibolite/eclogite within it (Fig. 8A). The serpentinite matrix of this block appears to be recrystallized serpentinite sandstone-breccia-conglomerate similar to the serpentinite matrix we will observe on day 2 of the field trip (stop 4) at Ring Mountain of Tiburon Peninsula. Smaller clasts (tens of centimeters in size and smaller) of recrystallized serpentinite sandstone can be found within the breccia exposures (Fig. 8C). Many clasts in the breccia have fabrics truncated at clast boundaries, and these pre-depositional fabrics are associated with blueschist-facies or, rarely, higher-grade assemblages. New blueschist-facies minerals grow across the original clast boundaries, and aragonite veins cut across both the clast boundaries and also postdate much of the deformation of the breccia. This indicates that many of the clasts in the breccia were partly subducted and exposed at least twice (once before deposition in the breccia and then together in the breccia). The high-grade blocks in the serpentinite lens may have been partly subducted and exhumed to the surface three times (Wakabayashi, 2012a).

Coherent felsic volcanic rocks are not found within the Franciscan complex, but such rocks are common in the Coast Range ophiolite (Hopson et al., 1981, 2008). Accordingly, the abundance of felsic volcanic clasts indicates that a significant fraction of the clasts in the breccia was derived from the upper plate of the subduction system, whereas the larger proportion of clasts

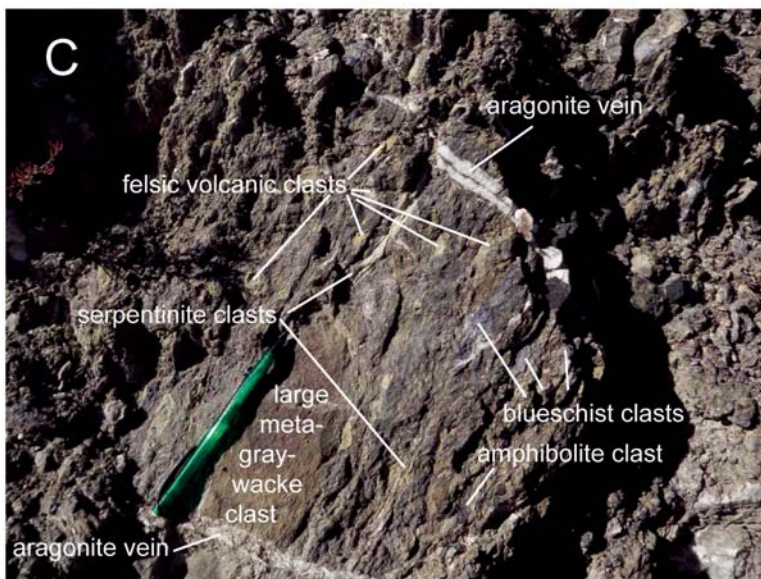
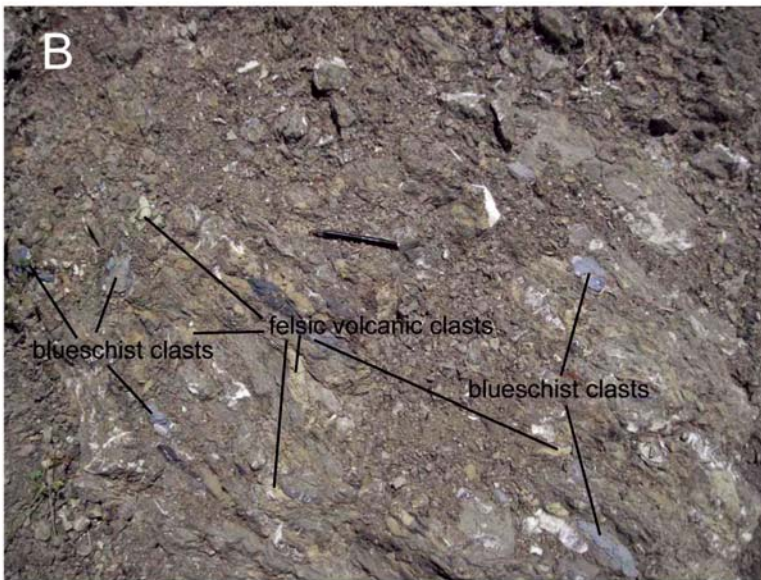
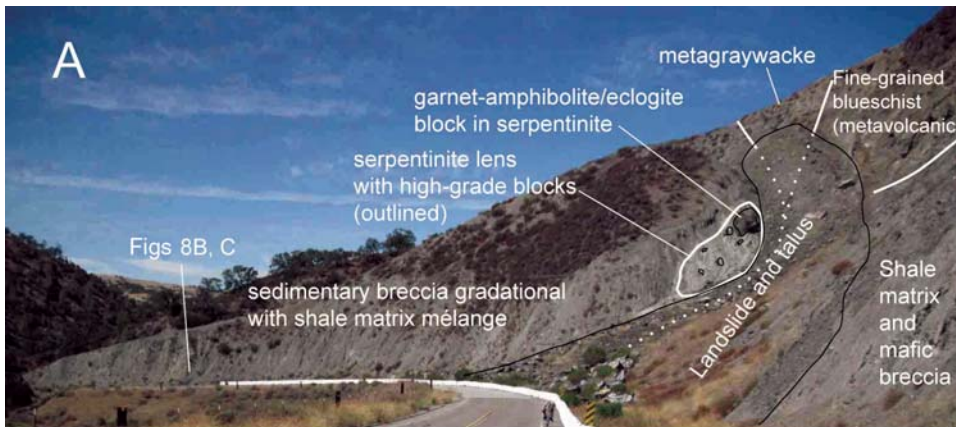


Figure 8. (A) Shale matrix breccia and mélange, including block of serpentinite mélange, western margin of Franciscan exposures along Panoche Road. Locations of photos in B and C are shown. (B) Close-up of sedimentary clast-supported breccia along Panoche Road. Most clasts are metavolcanic and metagraywacke. A few representative clasts are labeled. (C) Another close-up of the sedimentary breccia that includes an amphibolite clast, serpentinite clasts, and late aragonite veins. See Figure 7 for location.

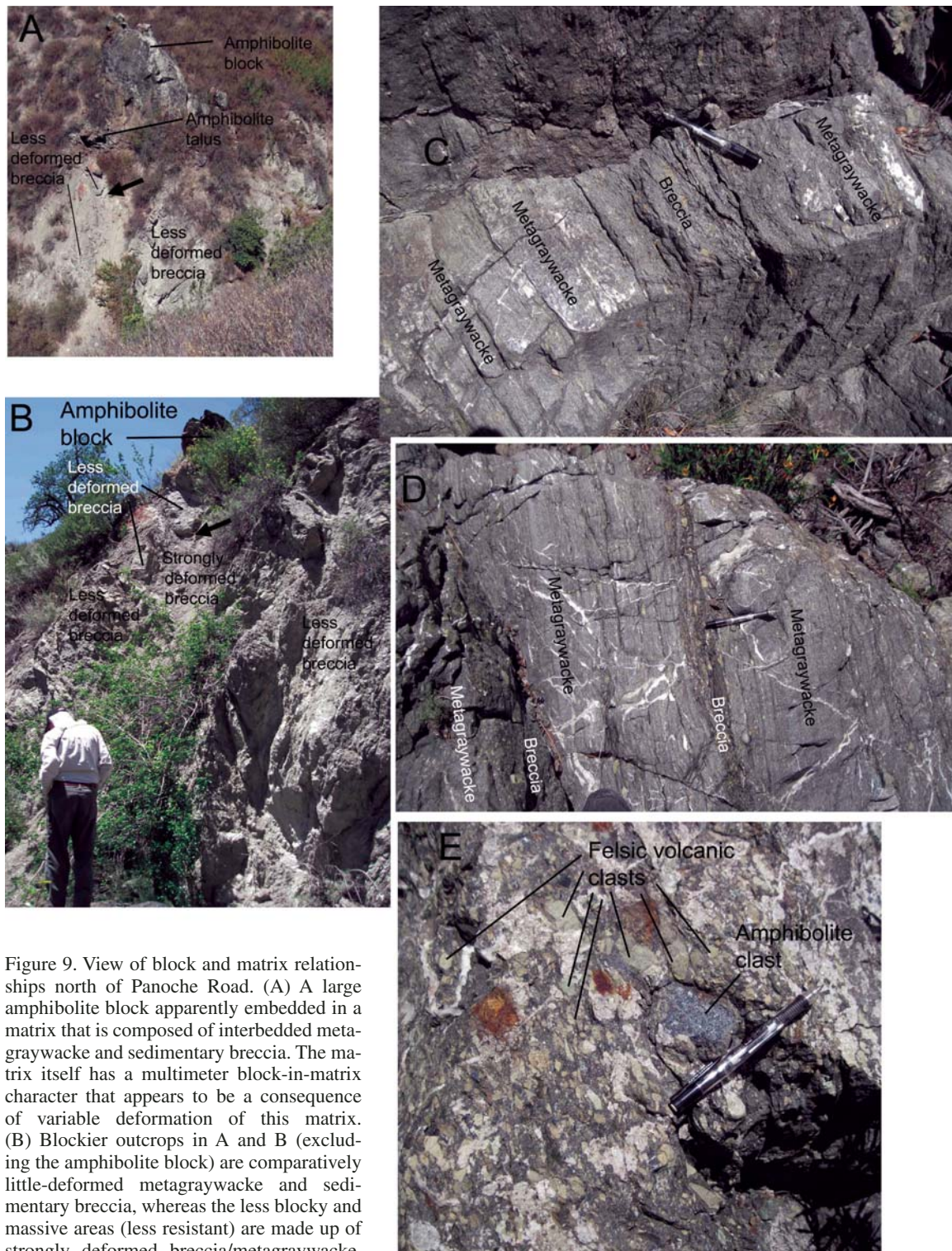


Figure 9. View of block and matrix relationships north of Panoche Road. (A) A large amphibolite block apparently embedded in a matrix that is composed of interbedded metagraywacke and sedimentary breccia. The matrix itself has a multimeter block-in-matrix character that appears to be a consequence of variable deformation of this matrix. (B) Blockier outcrops in A and B (excluding the amphibolite block) are comparatively little-deformed metagraywacke and sedimentary breccia, whereas the less blocky and massive areas (less resistant) are made up of strongly deformed breccia/metagraywacke. The thick black arrow in A and B points to a common reference point. (C–D) Interbedding of breccia and metagraywacke. The light-colored clasts in both are felsic volcanic rocks. In D, the darker breccia zones have a black shale matrix. The breccia varies from clast supported, as shown in Figures 8B and 8C, to having a shale to sandstone matrix, as seen in the examples here in photos C, D, and E. Note the gradation between the metagraywacke beds and breccia beds as a function of an increasing proportion of gravel-sized clasts in the latter. (E) An amphibolite clast in the sedimentary breccia. Note the abundance of felsic volcanic clasts, only a small fraction of which are labeled. C, D, and E are taken at the bottom of the gully (below bottom left corner of B and beneath the line of sight in A). See Figure 7 for location.

with predepositional blueschist (or higher-grade) metamorphism suggests derivation of most of the clasts from exhumed parts of the Franciscan complex.

The photos of Figures 9A and 9B and, to a lesser extent, the nature of exposures along the road cut also show a multiscale block-in-matrix fabric developed in the interbedded metagraywacke and breccia. The “blocks” or more resistant and massive outcrops have the aspect of exotic blocks from a distance, but they are instead composed of breccia that is less deformed than the less-resistant zones that surround them. The variability of physical characteristics of the *mélange*/breccia units is reflected in the geomorphology seen at this stop on larger (tens to hundreds of meters) scales. Some of the breccia exhibits a relatively weakly developed foliation and holds steep slopes, whereas the more strongly foliated breccia/*mélange* matrix underlies lower-angle slopes with the typical “plum pudding” topography commonly associated with *mélanges*. This contrast is especially apparent comparing the slopes north and south of the road.

Sandstone matrix for breccia/*mélanges* appears at least as common or perhaps more common than the better known shaly matrix. Recently, sandstone matrix olistostromes or *mélanges* have been identified in a number of localities throughout the Franciscan complex (Erickson, 2011; Wakabayashi, 2011a; Prohoroff et al., 2012). Sandstone matrix commonly does not exhibit the intense foliation typical of shale matrix. This results in sandstone matrix *mélanges* holding steeper slopes than shale matrix *mélanges*, as well as smaller strength and erosional resistance contrasts between block and matrix, muting the usual lumpy appearance commonly associated with *mélanges*. The variability

in geomorphic expression of Franciscan breccia/*mélange*, and the gradation between graywacke and sedimentary breccia (Figs. 9C and 9D) make distinguishing *mélange* from coherent units difficult in the field and blur the definition between *mélange* and coherent units when considering their place in the architecture of accretionary prisms (see also stop 3).

Further complicating this issue, there is the variety of settings of *mélanges*. Wakabayashi (2011a) delimited three different *mélange* settings within the Franciscan complex: at the structurally highest level of the accretionary complex, between coherent nappes, and within coherent nappes. The *mélanges* and sedimentary breccia units viewed at this stop appear to represent all three settings, and at this locality, all three types show evidence of sedimentary origins (see following paragraph for structurally highest horizon) (Wakabayashi, 2012a).

Much of the serpentinite body lacks a pronounced foliation and appears to be a sedimentary serpentinite breccia (Fig. 10A). Antigorite overgrows clasts boundaries of the serpentinite breccia (Wakabayashi, 2012a). Eclogite blocks occur within the serpentinite body, including some large ones (see block outlined on Fig. 7). As noted already, this serpentinite, or rather serpentinite matrix *mélange*, represents the structurally highest horizon in this area. This position within the Franciscan nappe stack is equivalent to the serpentinite *mélange* we will see at stop 4 (Ring Mountain, Tiburon Peninsula) and is similar to the serpentinite body in Sunol Regional Wilderness of the northern Diablo Range (Wakabayashi, 2012a). Although once considered a possible subduction shear zone, involving the hanging-wall mantle, that formed during and shortly after subduction initiation (Wakabayashi, 1990,

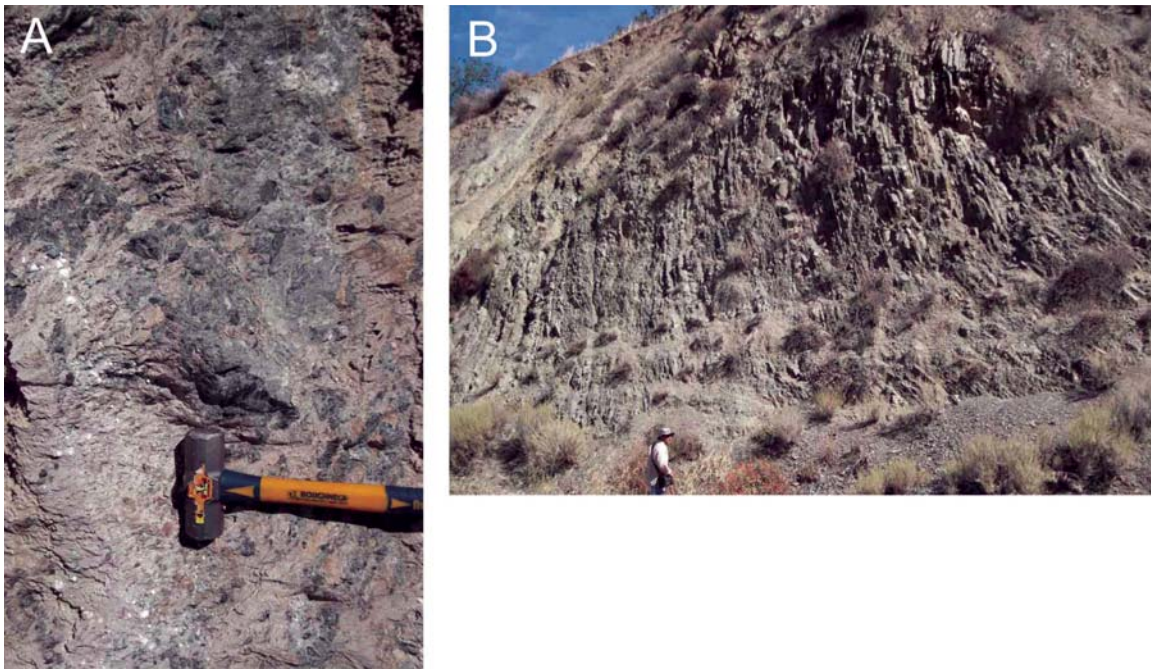


Figure 10. Outcrops along Panoche Road. (A) Serpentinite breccia. (B) Thinly bedded unfoliated metagraywacke that may be correlative of the Burnt Hills terrane. Locations are on Figure 7.

2011a), contact, matrix, and block-matrix relationships show that these serpentinite bodies have a sedimentary origin, and geochronology shows that they were deposited and accreted tens of millions of years after subduction initiation (Wakabayashi, 2012a). I will further discuss of the origins of Franciscan serpentinite bodies of this sort in the descriptions of stops 2 and 4.

East of the serpentinite, there is an exposure of thinly bedded sandstone and shale, possibly correlative to the Burnt Hills terrane as noted already (Fig. 10B). This particular outcrop is part of relatively small body of coherent sandstone (Ernst, 1965), but much larger exposures of this unit can be observed in road cuts to the east on the drive to this stop.

Drive from Stop 1 on the Panoche Road to Stop 2 in the Hayward Hills (~100 Miles; 161 Kilometers)

From Stop 1: Continue the drive westward on the Panoche Road; 0.5 miles (0.8 km) from stop 1, some strata can be seen in the cliffs to the right (north of road) with subhorizontal bedding. These are the Pliocene–Pleistocene San Benito gravels. They give an indication of the vertical component of late Cenozoic deformation in the Diablo Range. We come to the west end of Panoche Road 13 miles (21 km) from stop 1 and turn right (north) onto State Highway 25. Drive 24.4 miles (39.3 km) northward on SR25 through the town of Hollister to the junction with U.S. Highway 101, where we will curve right onto the on-ramp for northbound US101. Drive northward on US101 36.1 miles (58.1 km) to the city of San Jose, and then exit right onto northbound Interstate 880. Drive ~23.1 miles (37 km) northward on I-880 to the city of Hayward, and then exit right (northeast) onto Jackson Street. After ~0.2 miles (0.32 km) on Jackson Street, make a right (southeast) at the first stoplight onto Santa Clara Street, which soon curves leftward (east) and becomes Harder Road. Continue on Harder Road for ~2.2 miles (3.5 km) from Jackson, passing through an intersection with Mission Boulevard at the base of a hill and then climbing steeply to the California State University East Bay (formerly Hayward) campus. After the angle eases, pass one stop sign and then turn right (southward) into a driveway to a dormitory complex on the south side of the campus. This driveway skirts the right (western) edge of a parking lot and then curves left (northerly) to a T-intersection, where you will turn right into a parking lot (37.65126°N, 122.05026°W, depending on where you park in the lot).

Note: The locations of stops 2–5 and their regional geologic context are shown on [Figure 11](#).

Stop 2. Southern Margin of California State University East Bay (Formerly California State University Hayward) Campus (37.65126°N, 122.05026°W)

Here, we see some of the upper plate of the Jurassic–Cretaceous subduction system (Fig. 12). Extensive artificial cuts behind dormitory buildings expose gabbro of the Coast Range ophiolite, part of an elongate belt of exposures that extends for

more than 20 km along strike (Fig. 11). These gabbros exhibit many shear zones (Figs. 13A and 13B), associated with complex crosscutting relationships and multiple episodes of syntectonic mineral growth and veining. The syntectonic minerals include hornblende (Fig. 13C) indicative of amphibolites-grade metamorphism. The Great Valley Group clastic strata that depositionally overlie the Coast Range ophiolite lack burial metamorphism higher than zeolite grade (Dickinson et al., 1969). Accordingly, the shear zones in the gabbro record deformation associated with seafloor spreading.

Near the gabbro, we find outcrops of the Leona Rhyolite, a quartz keratophyre unit associated with the Coast Range ophiolite (Bailey et al., 1970), as well as outcrops of basal Great Valley Group shale and sandstone over Leona Rhyolite. The base of the Great Valley Group appears gradational with the top of the Leona Rhyolite, for the top of the Leona has interbeds of shale, similar to other exposures of this contact in the eastern San Francisco Bay area. The contact is poorly exposed at this location, but it is beautifully exposed in north Oakland (figures 9 and 10 of Wakabayashi, 2005). The basal Great Valley Group was deposited at ca. 145 Ma, based on detrital zircon and fossil ages (Surpless et al., 2006; Wright and Wyld, 2007). The age of the basal Great Valley Group strata indicates that the Leona Rhyolite is significantly younger than the 165–172 Ma age of the main components of the Coast Range ophiolite, such as the gabbroic and basaltic parts (Shervais et al., 2005; Hopson et al., 2008). This suggests that the Leona Rhyolite and similar units in the Coast Range ophiolite formed from a later and different magmatic event than that which formed the main part of the ophiolite (Hopson et al., 1981, 2008).

A block of amphibolite with blueschist overprint crops out on a saddle (see Fig. 12). This block, and similar ones in this area, appears to be associated with a serpentinite and shale matrix olistostromal unit within the basal Great Valley Group. The olistostromal unit and its field relationships were described and discussed in detail by Hitz and Wakabayashi (2012), and this guide merely summarizes some of the main points given therein. This location lacks matrix exposures, but it and its surroundings are more suitable for a large field trip, owing to the accessibility. The locations of good matrix exposures are given in Hitz and Wakabayashi (2012). The blocks in the unit include Great Valley Group sandstone, Coast Range ophiolite gabbro, high-pressure amphibolite, coarse- and fine-grained blueschist, and rare eclogite (one block found), and the matrix is composed of shale and serpentinite.

The olistostrome unit is similar to that described along the eastern margin of the northern Coast Ranges (western margin of northern Central Valley) (Moiseyev, 1970; Lockwood, 1971; Carlson, 1981; Phipps, 1984). The presence of this unit in the Hayward Hills suggests that such olistostromes are more widespread than previously thought. Recent geologic mapping in Jimtown and Healdsburg 7.5' quadrangles of Sonoma County, ~100 km northwest of Hayward Hills, has identified a unit of serpentinite matrix mélangé with high-grade blocks between

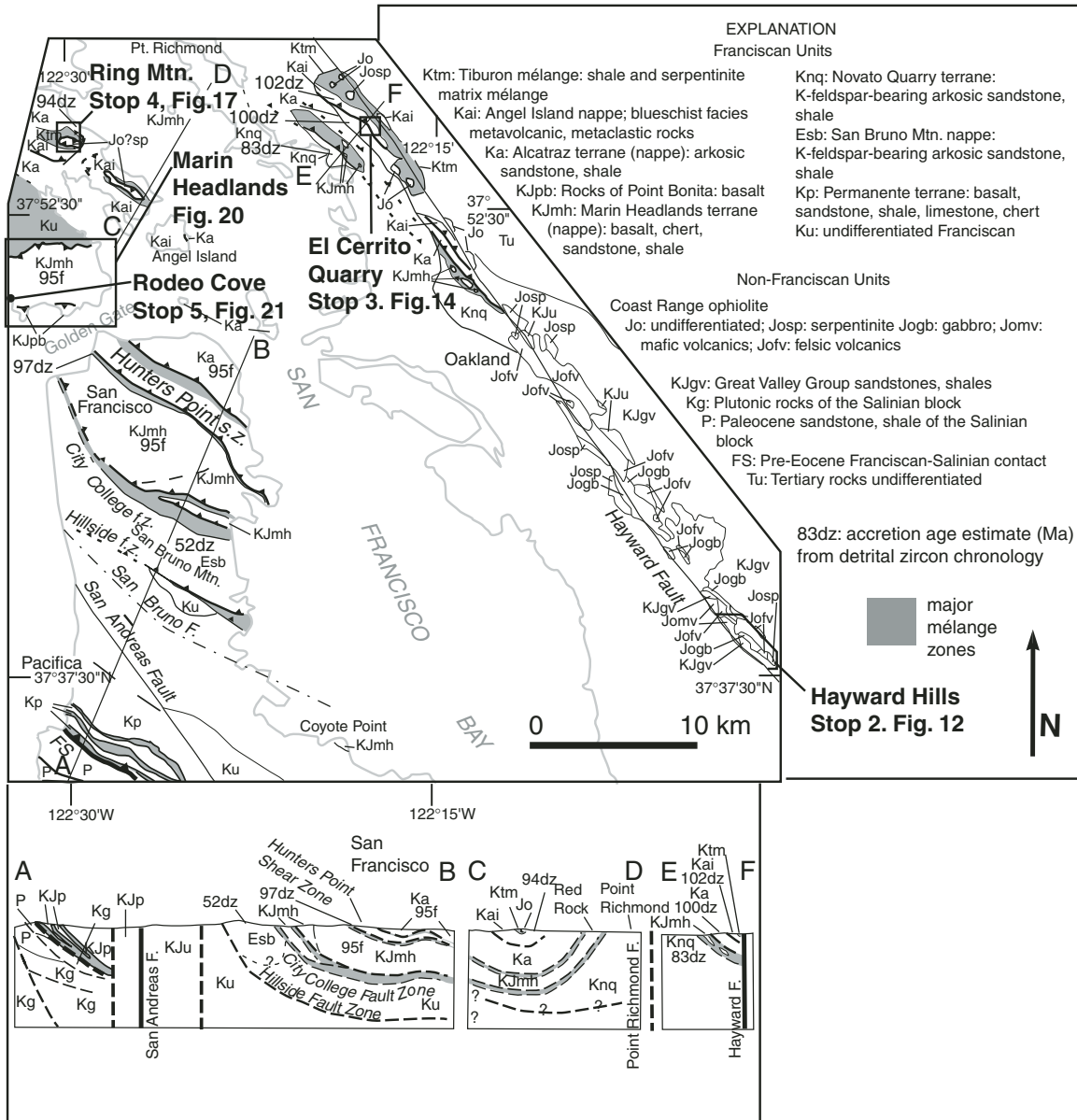


Figure 11. Franciscan complex geology of the San Francisco Bay area with major mélangé zones highlighted and field trip stops shown (modified from Wakabayashi, 2012a). Abbreviations: f.z.—fault zone; s.z.—shear zone.

block-free serpentinite of the Coast Range ophiolite and clastic sedimentary rocks of the Great Valley Group (Delattre, 2011; Delattre and McLaughlin, 2010). The structural and sedimentary position of the serpentinite mélangé unit in Sonoma County suggests that it may also be an olistostromal unit of the basal Great Valley Group.

Fryer et al. (2000) suggested that this olistostromal unit represented forearc serpentinite mud volcano deposits, similar to those found in the Mariana forearc. High-grade blocks are not present in younger Great Valley Group strata, suggesting that the forearc serpentinite mud volcanism was limited to the early part of the development of the trench-forearc system (ca. 145 Ma).

In contrast, deposition of sedimentary serpentinite with high-grade blocks in the Franciscan trench and subduction accretion of such deposits did not take place (or such deposits, if accreted, were not preserved) until 30 Ma or so, after the deposition of such material in the forearc basin, and deposition of Franciscan sedimentary serpentinites with high-grade blocks may have continued until at least 85 Ma (Wakabayashi, 2012a). This temporal relationship and the common occurrence of felsic volcanic rocks in Franciscan mélanges led Wakabayashi (2012a) to speculate that Franciscan sedimentary serpentinites (examples seen at stops 1 and 4) were derived by submarine sliding from basal Great Valley Group olistostrome deposits (Fig. 6).

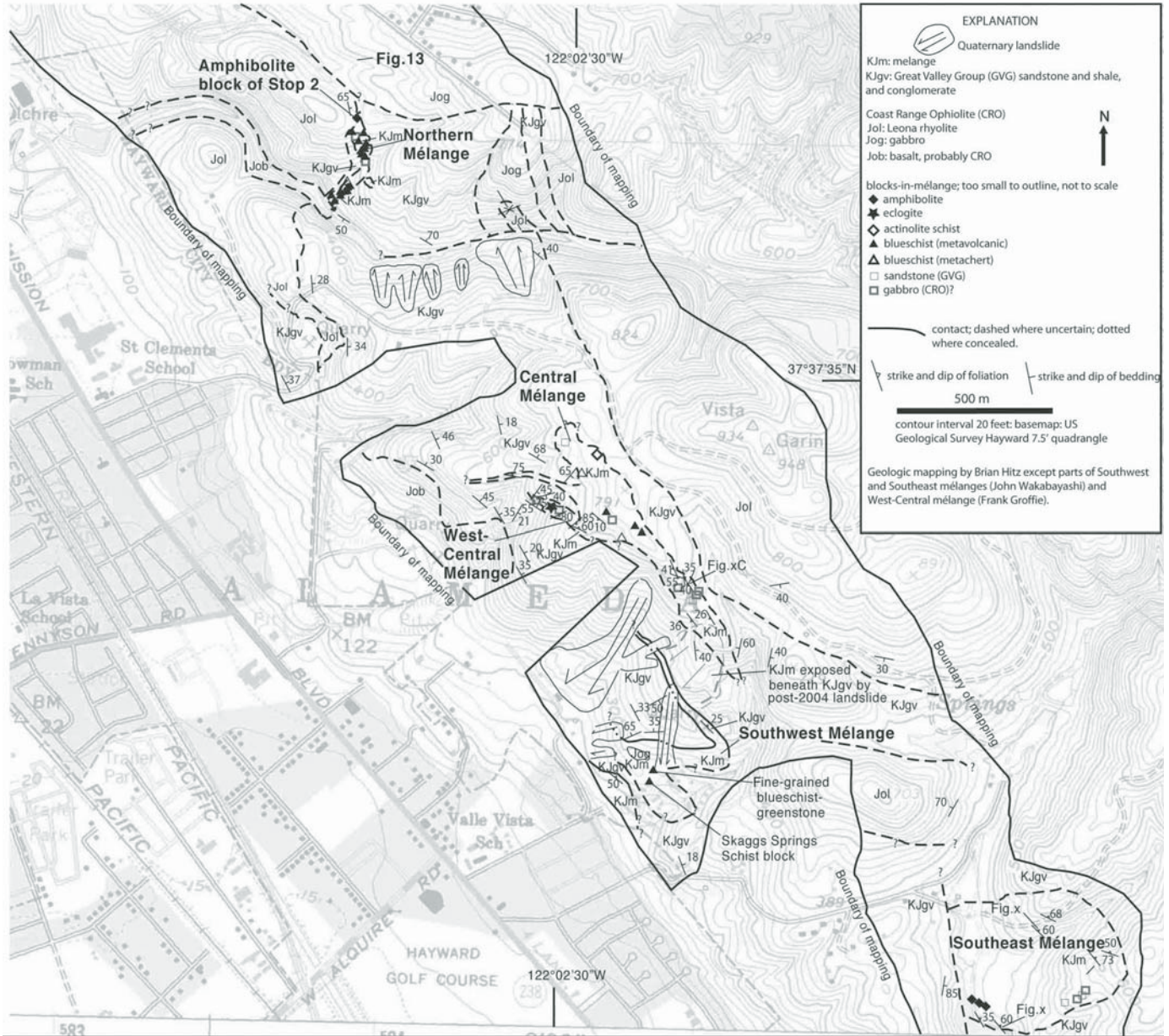


Figure 12. Geologic map of the Hayward Hills showing the location of stop 2. Note that there has been considerable local modification to the topography since the publication of the topographic map base, including the artificial exposures behind the dormitory buildings that are viewed in the photo of Figure 13. Map is adapted from Hitz and Wakabayashi (2012).

Drive from Stop 2 Hayward Hills to Stop 3 El Cerrito Quarry

From Stop 2: Retrace route (~2.5 miles; 4 km) (left onto Harder Road out of dormitory complex, left onto Jackson, right onto northbound I-880 ramp) back to I-880 and head northward. Head northward on I-880 to the city of Oakland. Take the connector freeway to I-80 northbound (right exit) and continue northward on I-80 past Emeryville and Berkeley, and then exit right (northeast) at the Central Avenue, 24 miles (38.6 km) after

entering I-880 from Jackson Street in Hayward. At the bottom of the Central Avenue ramp, turn right (eastward). Drive on Central Avenue ~0.7 miles (1.1 km) eastward, getting into the left lane and turning left (northerly) onto San Pablo Avenue. Drive 0.7 miles (1.1 km) northward on San Pablo Avenue, and then turn right (easterly) on Moeser Lane. Drive 0.2 miles (0.3 km) on Moeser Lane, and then turn left (northerly) onto Richmond Street. Drive 0.3 miles (0.48 km) northward on Richmond Street, and then turn right (easterly) onto Schmidt Avenue and drive ~0.3 miles (0.48 km) to the end of the road outside of the recycling center,

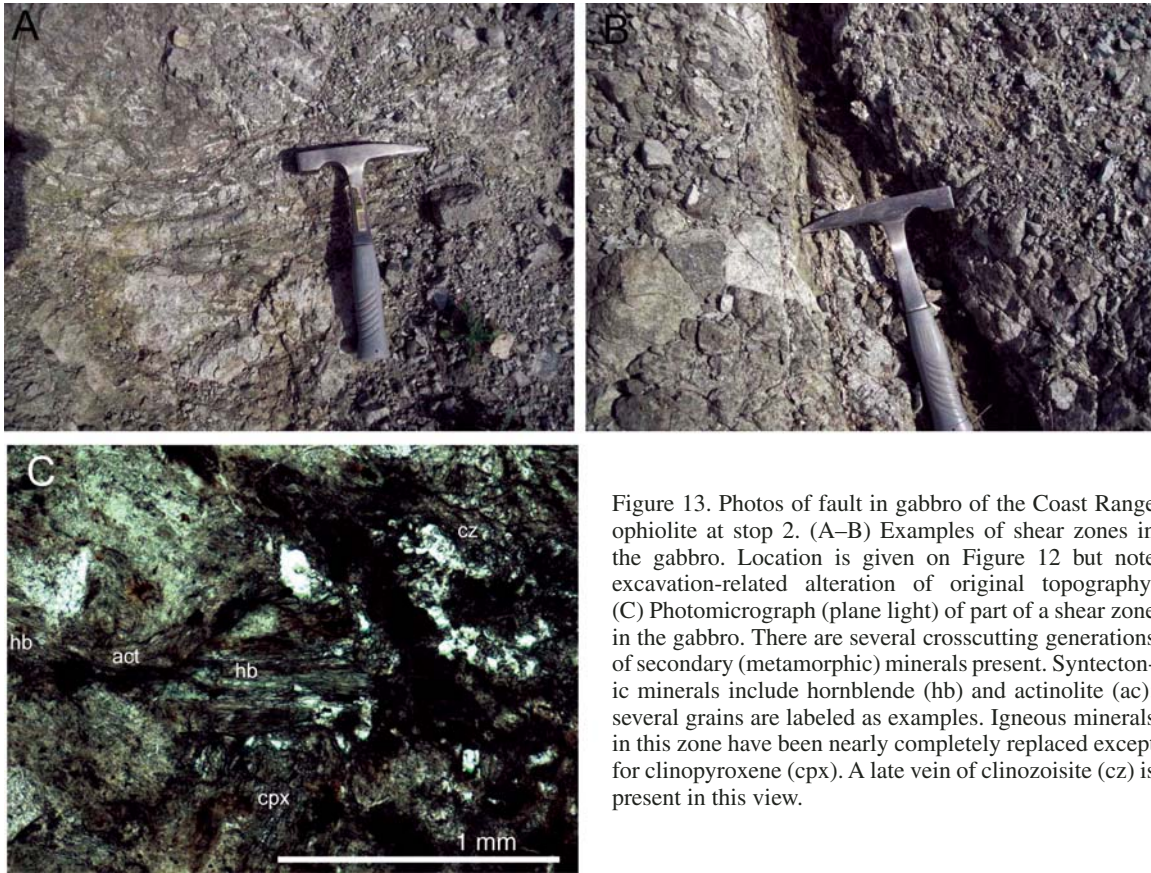


Figure 13. Photos of fault in gabbro of the Coast Range ophiolite at stop 2. (A–B) Examples of shear zones in the gabbro. Location is given on Figure 12 but note excavation-related alteration of original topography. (C) Photomicrograph (plane light) of part of a shear zone in the gabbro. There are several crosscutting generations of secondary (metamorphic) minerals present. Syntectonic minerals include hornblende (hb) and actinolite (ac); several grains are labeled as examples. Igneous minerals in this zone have been nearly completely replaced except for clinopyroxene (cpx). A late vein of clinozoisite (cz) is present in this view.

where we will park to the right (37.91878°N, 122.30013°W, depending on exact location of where car is parked).

Stop 3. El Cerrito Quarry, East End of Schmidt Avenue, El Cerrito (37.91878°N, 122.30013°W)

This abandoned quarry exposes the contact between two coherent nappes of the Franciscan complex in the San Francisco Bay area (Fig. 11). The contact places structurally higher jadeite-bearing metagraywacke of the Angel Island nappe over the prehnite-pumpellyite facies graywacke of the Alcatraz nappe (Figs. 14, 15, and 16). A *mélange* zone separates these two graywacke units. The difference in metamorphic grade and the regional nature of this contact suggest that it represents a paleosubduction megathrust between units accreted at different times and exhumed from different depths (Wakabayashi, 2011a, 2011b). Although a *mélange* zone separates the two coherent units, most of the displacement may not be accommodated within the *mélange*. The block-in-matrix fabric of the *mélange* appears to have a sedimentary origin based on well-preserved sedimentary breccia in little-deformed domains exposed on the quarry walls (Fig. 16D), and the *mélange* or sedimentary breccia unit appears to rest in depositional contact on the Alcatraz nappe graywacke (Fig. 16C). The metamorphic

contrast between prehnite-pumpellyite-grade and blueschist-facies rocks is located within the upper 20 m of the *mélange* zone (Fig. 16B).

At the top of the *mélange*/sedimentary breccia unit, there is an intensely deformed zone of shale and sandstone, ~20 m thick (Fig. 16A). Exotic blocks such as volcanic, chert, and ultramafic rocks that are common within the *mélange*/breccia below are rare or absent in this deformed zone, shown as the “transition” zone on Figure 15. Fault rocks consisting of ultracataclasite and frictional melts, resembling the black fault rocks of the Kodiak Island subduction complex (Rowe et al., 2005; Meneghini et al., 2010), are common in this highly deformed zone (C. Rowe, 2012, personal commun.). The sense of shear is top-to-the-southwest (Figs. 14 and 15A), consistent with the transport required to create the metamorphic contrast of higher-pressure (blueschist facies) over lower-pressure metamorphic (prehnite-pumpellyite facies) rocks.

The *mélange* zone appears to have originated as an olistostromal unit deposited on the lower graywacke and should be considered a part of the lower nappe. Whereas zones of moderately well-developed foliation, with a top-to-the-southwest sense of shear, occur within the *mélange*, most of the displacement, and the likely location of the megathrust during most of the accretionary history of this contact, was along the upper part

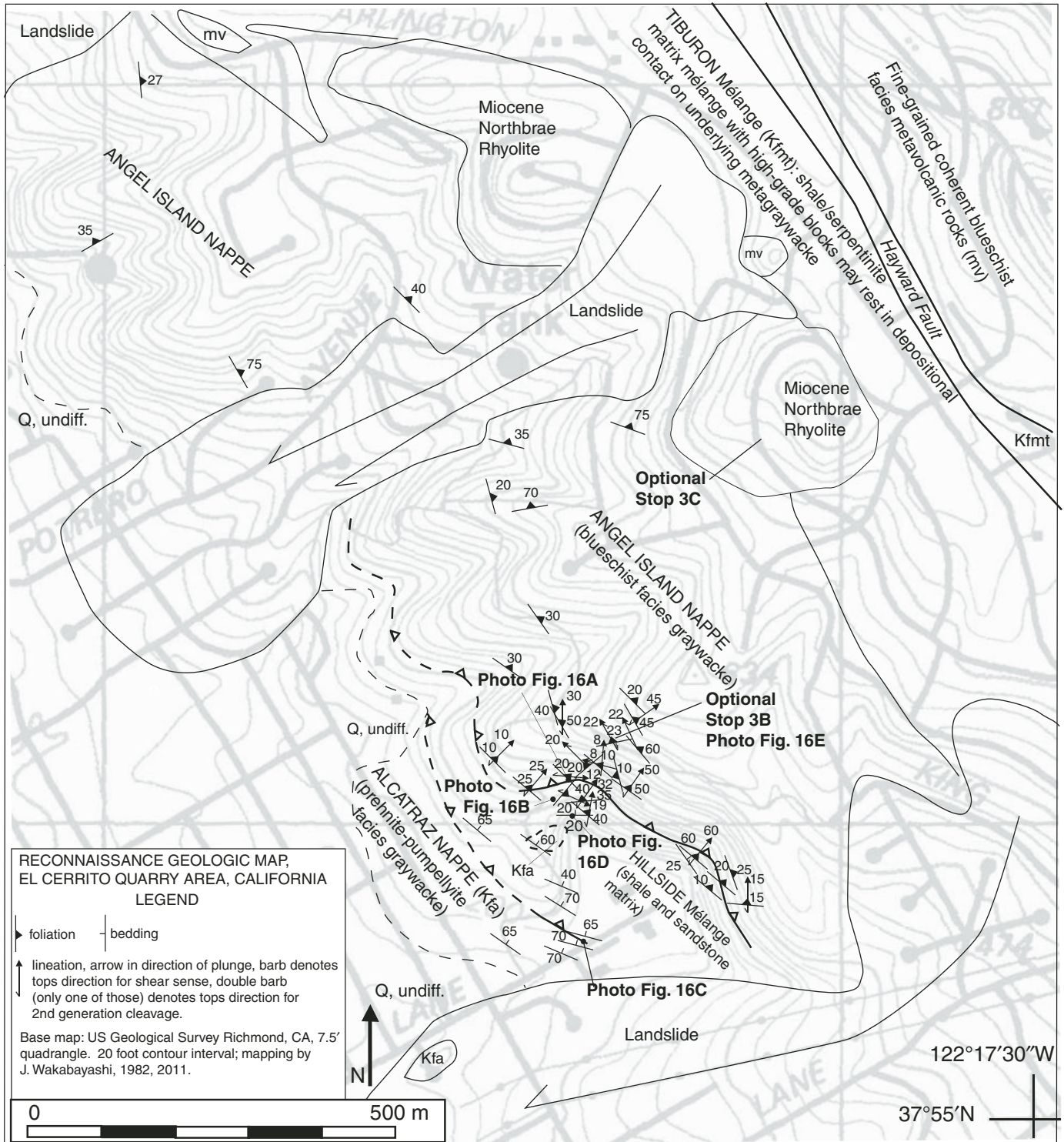


Figure 14. Geologic map of El Cerrito quarry and vicinity. Locations of various outcrop photos are shown.



Figure 15. Photograph of El Cerrito quarry showing locations of some of the outcrop photographs. Photo was taken from near the east end of Schmidt Avenue looking NW.

of the *mélange*, as schematically shown in Figure 5E. Whether this relationship is typical of this level of exposure of subduction complexes is unclear at this time.

If we have time, we will view Miocene rhyolites (Fig. 14) that overlie Franciscan units above the top of the quarry walls.

Optional Stops 3B and 3C above Quarry Driving and Walking Instructions and Geology

From Stop 3: Drive westward on Schmidt Avenue for ~0.1 mile (0.16 km) to the first intersection, where you will turn left (southward) onto Navellier Street. Drive 0.2 miles (0.32 km) southward on Navellier to Moeser Lane and turn left (eastward) and proceed up a steep hill. Drive 0.4 miles (0.64 km) up Moeser Lane, and then turn left (northward) on Shevlin Drive. Proceed ~0.4 miles (0.64 km) north on Shevlin Drive, and then turn left (northward) onto King Street, and then take an immediate left (westerly) into King Court (dead end) and park. One optional stop (3B) will examine some outcrops of blueschist-facies graywacke that appear to host an “anti-thetic” shear zone that may be related to exhumation. At the end of the road, take the hiking trail that is the continuation of King Court. This trail curves rightward (northward). This trail traverses above the rim of the quarry. After ~400 m of walking, we come to an outcrop on the right (N) that has unusual light-colored clasts in it (37.92115°N, 122.29908°W) (stop 3B).



Figure 16. Outcrop photographs spanning a range of exposures from structurally high to low within the Hillside *mélange*. Locations of outcrops are shown on Figures 14 and 15. (A) Uppermost blueschist-facies part of the *mélange* with a strong shear fabric, from ~4 m below the top of the *mélange*. This rock consists of thin sandy interbeds, metashale, quartz veins, and black fault rock. (B) The base of the blueschist-facies part of the *mélange* (right), ~20 m below the top of the *mélange*, against the prehnite-pumpellyite-facies part of the *mélange* (left). The blueschist-facies part of the *mélange* exhibits a strong shear fabric and is laced with black fault rocks, whereas the structurally lower prehnite-pumpellyite part displays less strain. (C) Basal contact of the *mélange* exposed by a temporary (late 2011) excavation along Schmidt Avenue on the eastern margin of exposures along the road. Long dimension of photo is ~3 m. The base of the *mélange* (right) rests on graywacke of the Alcatraz nappe to the west (left) along a steeply dipping (~70° to the northeast) contact. Excavator teeth have smeared the texture of the *mélange* here, but undisturbed (unsmeared) domains can be observed in direct contact with the graywacke along the upper part of this exposure (foliation “blades” are smeared along the lower part). The sedimentary (little deformed) character of the *mélange* directly above the contact shows the depositional nature of the basal contact. (D) Silty to sandy matrix sedimentary breccia with some rounded clasts and no fabric. This is best seen by zooming in on these photos. Although most clasts are angular, rounded clasts are present as well (some pointed out arrows). Pyroxenite clast (px) is labeled, and the region above the clast is very rich in both gravel- and sand-sized volcanic clasts. (E) Optional stop 3B. Brittle top-to-the-right (SSE) C-S fabric overprints earlier top-to-the-left (NNW) fabric seen by tail on left of light-colored clast.



Stop 3B

This metagraywacke is part of the blueschist-facies Angel Island nappe that makes up the upper quarry walls. The light-colored clasts resemble some of the serpentinite sandstone clasts of stop 1 in outcrop appearance, but thin sections I have examined so far show that these clasts have a significant amount of quartz and neoblastic lawsonite, so the origin of these clasts is unclear. The fabric in this outcrop is unusual because the brittle C-S fabric appears antithetic to an earlier weak shear fabric in the metagraywacke observed in much of the area (Fig. 16E; see Fig. 14 geologic map for shear sense data). Whereas the earlier fabric shows a top-to-the-NNW shear sense (shown asymmetric tail off of left side of light-colored clast at center of Fig. 16E photo), owing to late folding, the later fabric is top-to-the-SSE (C-S fabric, including deformation of light-colored clast). The later fabric may have been associated with normal faulting during extensional exhumation of the blueschist-facies graywacke (e.g., Platt, 1986), or related to the roof fault accommodating cross-sectional extrusion (Maruyama et al., 1996), but the significance of this outcrop is unclear, because (1) it occurs within the blueschist-facies slab rather than marking its upper boundary against (presumably) lower-pressure metamorphic rocks; and (2) this fabric has not been identified in any other outcrop in this area, so a number of scenarios may explain its origin. This outcrop, however, illustrates the difficulty in identifying potential exhumation fabrics in high-pressure metamorphic rocks such as these. If this outcrop did not have the rare indicators of the earlier fabric, the later fabric may have been mistaken for the earlier one.

From this optional stop, we may retrace our hike on the trail back to the vehicles and then walk ~700 m to see outcrops of Miocene rhyolite (the Northbrae Rhyolite). To do so: Walk out of King Court, turn left (westward) on King Drive, and walk until it becomes Contra Costa Drive (which has merged from the right or SE). Turn right (NE) onto Buckingham Drive, then left onto Brewster Drive, and then walk until you see light-colored outcrops on your right (E) (stop 3C).

Stop 3C

These unmetamorphosed Miocene rhyolites unconformably overlie the Franciscan complex and are part of the postsubduction magmatic suite associated with the slab window that formed during the transition from a convergent to transform plate boundary (e.g., Johnson and O'Neil, 1984; Cole and Basu, 1995). This is a good example of magmatism associated with subduction termination that is not associated with terminal collision.

These rocks were originally mismapped by many (including me) as equivalents of the Leona Rhyolite of the Coast Range ophiolite before this error was corrected by fieldwork and geochronology (Murphy, 2001; Murphy et al., 2002). As a result of the erroneous assignment of these rocks to the Coast Range ophiolite, in earlier field trip guides, I had considered their basal contact to be a major exhumation fault (Wakabayashi, 1999c, 2005). The identification of these rocks as Miocene volcanic

rocks shows that interpretation to be incorrect, for the basal contact of the volcanic rocks is likely depositional.

Driving Directions from Stop 3 or 3B/C to Oakland (Lodging) (~12 Miles; 19.3 Kilometers)

From either stop 3 or the optional stop 3B/C, retrace the route to San Pablo Avenue, Central Avenue. This road log reflects the route of the 2013 Geological Society of America (GSA) Cordilleran Section trip, wherein stop 3 concludes day 1, and the group drives to downtown Oakland for lodging. However, it is likely that geologists running this trip for themselves will have a different itinerary, so the second-day driving directions will include sufficient information on various road junctions that the route can be followed from a number of locations, or, if desired, one can drive from stop 3 to stop 4. The latter task can be accomplished by heading west on Central Avenue, passing beneath I-80, and then exiting right (northbound) onto I-580 toward the Richmond–San Rafael Bridge.

For the return to Oakland, from westbound Central Avenue, keep left as you approach I-80. Pass beneath the overpass, and then turn left (southward) onto the southbound I-80 on-ramp. Drive southward on I-80, keeping in the middle lanes designated for southbound I-880 as one reaches Emeryville. Once on I-880 and entering Oakland, take the exit for Broadway (right exit) and follow the signs to Broadway. The Marriott Hotel Oakland City Center can be reached by turning left (northward) on Broadway, passing under the freeway, and driving several blocks into the downtown area. The total driving distance from stop 3 to the Marriott Hotel Oakland City Center is ~12 miles (19.3 km).

Driving Directions—Downtown Oakland to Stop 4, Ring Mountain, Tiburon Peninsula (~25 Miles [40 Kilometers]; Stop 3–Stop 4 Direct Distance ~19.5 Miles [31.4 Kilometers])

From downtown Oakland, drive southward toward the I-880 freeway on Broadway. Look for signs directing you to northbound I-880 or San Francisco. Once on I-880, retrace part of the drive of the previous day, heading toward northbound I-80, but then exiting right onto I-580 bound for the Richmond–San Rafael Bridge. The mileage log for the last two stops begins where the on-ramp from Central Avenue joins I-580, ~8 miles (12.8 km) from Oakland. From the junction with Central Avenue, continue northwestward, first to the toll plaza, and then cross the San Francisco Bay on the bridge. On the west side of the bridge, exit right onto Sir Francis Drake Boulevard, 10.7 miles (17.2 km) on I-580 past the Central Avenue on-ramp. This takes you on an overpass over the freeway and then westward to Larkspur. Keep to the left lane. After 2 miles (3.2 km) of driving, pass under the US101 overpass and turn left onto a ramp for southbound US101 toward San Francisco. Follow US101 southward for 1.4 miles (2.3 km) and then exit right (westward) on Tamalpais/Paradise Drive and keep to the left lane of the

off-ramp toward its end. Make a left (eastbound), which takes you over the freeway, and continue for 0.2 miles (0.32 km); then turn right at the T-intersection (curving) onto San Clemente Drive. This will eventually curve into Paradise Drive. After 2.2 miles (3.5 km) on San Clemente/Paradise Drive, make a right onto Taylor Avenue, which ascends steeply. Drive 0.5 miles (0.8 km) to near the end of Taylor Avenue, where you will find parking on the right (37.9133°N, 122.48696°W).

Stop 4. Ring Mountain, Tiburon Peninsula, Taylor Avenue Parking (37.9133°N, 122.48696°W)

Note: This is a nature preserve where collecting of rocks is not permitted. Please do not bring your hammers to this stop.

Geologists and nongeologists alike enjoy Ring Mountain for the beauty and geologic significance of the rocks, as well as the stunning views of the San Francisco Bay and surrounding cities. Many studies have investigated rocks at Ring Mountain. Taliaferro (1943) probably published the first geologic map showing block distribution and detailed geology. Dudley (1967, 1969, 1972) conducted studies on the metamorphism of the high-grade (eclogite, amphibolite, coarse blueschist) blocks-in-mélange there. Coleman and Lanphere (1971) investigated metamorphic blocks at Ring Mountain as a part of the first Franciscan-wide study of the high-grade blocks-in-mélanges. Ring Mountain has become a popular field trip stop, as illustrated by various published field trip guides that feature this stop (e.g., Liou, 1989; Wakabayashi, 1992b, 1999c); many other less formal field trip guides exist (e.g., Bero, 2004).

Previous field trips have focused nearly entirely on the beguiling high-grade blocks rather than field relationships, although Liou (1989) presented a geologic map. This field trip and guide will examine the field relationships a bit more closely as a result of recent geologic mapping of Ring Mountain (Wakabayashi, 2012a). The following field descriptions and the geologic map of [Figure 17](#) are derived from that reference, as are the field panoramas of [Figure 18](#), and outcrop photo shown in [Figure 19A](#); more detailed petrographic documentation and additional outcrop photos are given in the original reference. Outcrop photos in [Figures 19B and 19C](#) are sharper photos and show features better than their equivalents in the earlier paper, and the contact relations between the blueschist-facies coherent rocks and the prehnite-pumpellyite-facies rocks in the southeastern portion of the map have been updated on [Figure 17](#).

The geology of Ring Mountain consists of several folded nappes. Variably serpentinized harzburgite without exotic blocks makes up the highest nappe. Serpentine minerals in this nappe appear to be entirely or predominantly lizardite with some chrysotile. Structurally beneath this unit, there is a serpentinite matrix mélange with a variety of metamorphic blocks. The mélange matrix, consisting of serpentinite sandstone, breccia, and conglomerate with exotic sand- and gravel-sized clasts, has itself been recrystallized, with growth of antigorite and possibly talc and/or tremolite. It is not clear whether the latter occur here as

true neoblastic minerals (postdating sedimentation of serpentinite debris); they are common in detrital clasts.

The blocks-in-mélange range in size up to 160 m in the long dimension. Most of them are “high-grade” blocks. The most common high-grade blocks are amphibolite, garnet-amphibolite, eclogite, omphacitic greenstones (commonly with omphacite, epidote, chlorite, phengite), high-grade metacherts (mostly quartz-dominated coarse blueschists), coarse blueschists, and garnet blueschists. The majority of the blocks are metabasites, with subordinate metachert (including metachert layers with metabasite blocks), felsic metavolcanic rocks, and rare metadiorite and metapyroxenite. All garnet blueschists have some relict omphacite and represent eclogites heavily overprinted by blueschist-facies assemblages. Nearly all of the garnet-free blueschists have relics of earlier higher-grade assemblages (most commonly calcic amphibole cores in sodic amphibole and rutile cores in titanite). Amphibolites, garnet-amphibolites, eclogites, and omphacitic greenstones all feature variable overprinting by blueschist assemblages, and many, if not most, garnet amphibolites show evidence of an eclogitic stage of metamorphism that followed the amphibolite crystallization but preceded the late blueschist overprint.

A few lower-grade blocks are present, including lawsonite-jadeite-glaucophane-metagraywacke, feebly recrystallized metachert, and metabasalt with incipient blueschist-facies mineral growth (some sodic amphibole and lawsonite). Most of the high-grade blocks are metamafic, but some blocks of high-grade metachert crop out, and metachert layers are common within the mafic blocks.

The serpentinite matrix mélange overlies apparently coherent blueschist-facies Franciscan rocks consisting of metagraywacke, chert, and metabasalt similar to the lower-grade blocks described earlier ([Fig. 17](#)). The coherent blueschist-facies rocks tectonically overlie prehnite-pumpellyite sandstone and shale that lacks a metamorphic fabric.

Many striking high-grade blocks crop out within a baseball’s throw of the parking lot of area 1 (northeast part of [Fig. 17](#)) where we get out of our vehicles (parking area shown on northeast part of [Fig. 17](#)). Whereas some blocks may be part of landslide complexes ([Fig. 17](#)), many appear in place within serpentinite matrix, although block-matrix contacts are fairly rare.

The high-grade blocks we see are reasonably representative of the thousands of such blocks that occur elsewhere in various mélanges of the Franciscan complex. Study of Ring Mountain high-grade blocks reflects research on these blocks throughout the complex. Because of their older metamorphic age and higher metamorphic grade than intact (“coherent”) Franciscan metamorphic rocks, Coleman and Lanphere (1971) considered the high-grade rocks to be derived from a cryptic, pre-Franciscan terrane, whereas Platt (1975), Cloos (1985), and Wakabayashi (1990) considered the high-grade blocks to represent rocks formed at the initiation of Franciscan subduction. The finding of coherent fault-bounded slices of Franciscan high-grade rocks at the highest structural levels of the complex, first by Ernst et al. (1970)



Figure 17. Geologic map of Ring Mountain. Location is shown on Figures 1 and 11. Figure is revised from Wakabayashi (2012a).

at Goat Mountain in the northern Coast Ranges, and later in the southern Diablo Range, and Sonoma County (Wakabayashi and Dumitru, 2007), appears to support the interpretation that these high-grade rocks formed during Franciscan subduction, but some may disagree. Lu-Hf garnet and ⁴⁰Ar-³⁹Ar hornblende ages from the high-grade rocks suggest initiation of Franciscan subduction at ca. 155–170 Ma (Anczkiewicz et al., 2004; Wakabayashi and Dumitru, 2007; Shervais et al., 2011).

The initial interpretation of counterclockwise pressure-temperature (*P-T*) paths (*P* on positive *y*-axis) in Franciscan high-grade blocks was based on rocks from Ring Mountain (Wakabayashi, 1990) and Jenner (Oh and Liou, 1990; Krogh et al., 1994). The interpretation of this *P-T* path as the metamorphic record of subduction initiation was based on Ring Mountain samples (Wakabayashi, 1990). The rocks of Ring Mountain have figured prominently in interpretation of *P-T* evolution associated

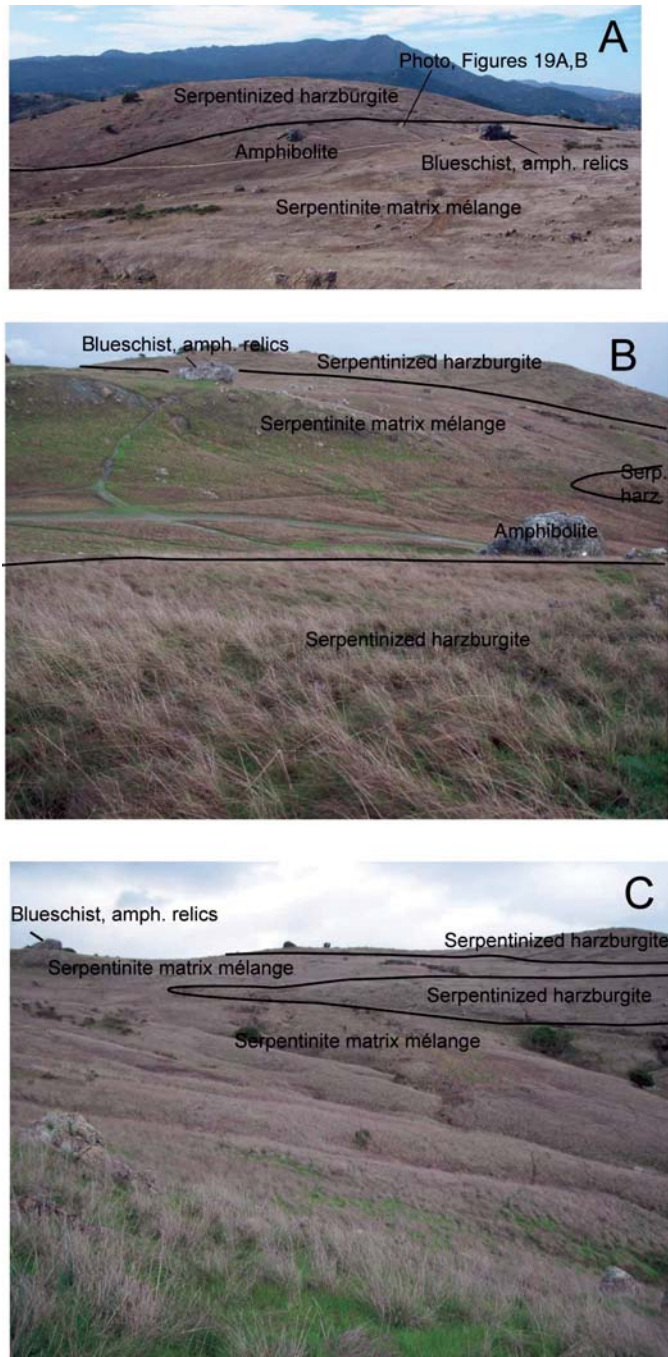


Figure 18. Ring Mountain panoramas. (A) View westward. The position for photos in Figures 19A and 19B can be used for reference on map Figure 17, and the blueschist block with amphibolite relics is common to all three of these photos. “Serpentinized harzburgite” refers to somewhat massive serpentinite lacking blocks, whereas serpentinite matrix *mélange* includes exotic block. (B) View eastward. The two blocks labeled are the same blocks labeled in “A.” Note the block-in-granular-matrix texture of the matrix exposure in middle ground. This contrasts to the foliated matrix seen in Figures 19A and 19B. (C) View northeast.

with the subduction process in general (Ernst, 1988). Massonne (1995), in the first application of garnet-clinopyroxene-phengite (grt-cpx-phen) thermobarometry to Franciscan rocks, used mineral compositions in Wakabayashi (1990) from a Ring Mountain sample and estimated metamorphic conditions of 600 °C and 2.5 GPa for eclogite-facies conditions, approximately doubling earlier pressure estimates and slightly raising the temperature estimates. Tsujimori et al. (2006) applied more recent calibrations of grt-cpx-phen thermobarometry on a hornblende eclogite studied earlier by Wakabayashi (1990) and estimated *P-T* conditions of 550–620 °C at 1.9–2.5 GPa. Page et al. (2007) obtained similar *P-T* estimates from an eclogite block in Sonoma County to the north. The higher pressure of eclogite metamorphism for Franciscan high-grade blocks also demands a somewhat more “hairpin” counterclockwise *P-T* path for high-grade blocks that record eclogite metamorphism (Tsujimori et al., 2006; Page et al., 2007), compared to the more “open” counterclockwise *P-T* loop proposed earlier by Wakabayashi (1990).

It is likely that ranges of peak *P-T* conditions and *P-T* paths are recorded by the blocks at Ring Mountain; studies to date have tended to focus on the eclogites. As noted before, a few fine-grained blocks appear to have experienced only blueschist-facies metamorphism. In addition, there are garnet-free amphibolites (and their equivalents retrograded to blueschist) that do not appear to have experienced eclogite-facies conditions, although the presence of rutile, as well as the high Al and comparatively low Ti content of calcic amphiboles (Wakabayashi, 1990) suggests relatively high pressure (e.g., Ernst and Liu, 1998). Given the field relationships that suggest the blocks were originally exhumed, then deposited in a serpentinite olistostrome, and then resubducted, a range in *P-T* conditions and *P-T* paths in the blocks should be expected (Wakabayashi, 2012a).

Geochronologic studies have come to somewhat different conclusions based on rocks at Ring Mountain. Anczkiewicz et al. (2004) obtained a 153.4 ± 0.8 Ma Lu-Hf garnet age from a block at Ring Mountain. Anczkiewicz et al. (2004) suggested that the relationship between estimated peak metamorphic temperatures and metamorphic age for various high-grade blocks in the Franciscan complex indicated very slow cooling of the subduction zone after subduction inception and slow initial subduction rates. Page et al. (2007) disagreed, concluding that their *P-T* estimates and those of Tsujimori et al. (2006) indicated the establishment of very low geothermal gradients by the time of eclogite metamorphism, including that at Ring Mountain. Catlos and Sorensen (2003) obtained $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 153 ± 2 Ma to 156 ± 6 Ma from texturally late white micas from a block at Ring Mountain. These ages may suggest fast cooling if taken together with the Anczkiewicz et al. date, although these dates apparently came from a different block that may have had a different cooling history. A K/Ar hornblende age of 152.8 ± 7.5 Ma coupled with a K/Ar phengite age of 153.1 ± 3.3 Ma obtained by Tsujimori et al. (2006) on the same block dated by Anczkiewicz et al. (2004) may suggest rapid cooling, but the large uncertainties are also permissive of fairly slow cooling too.

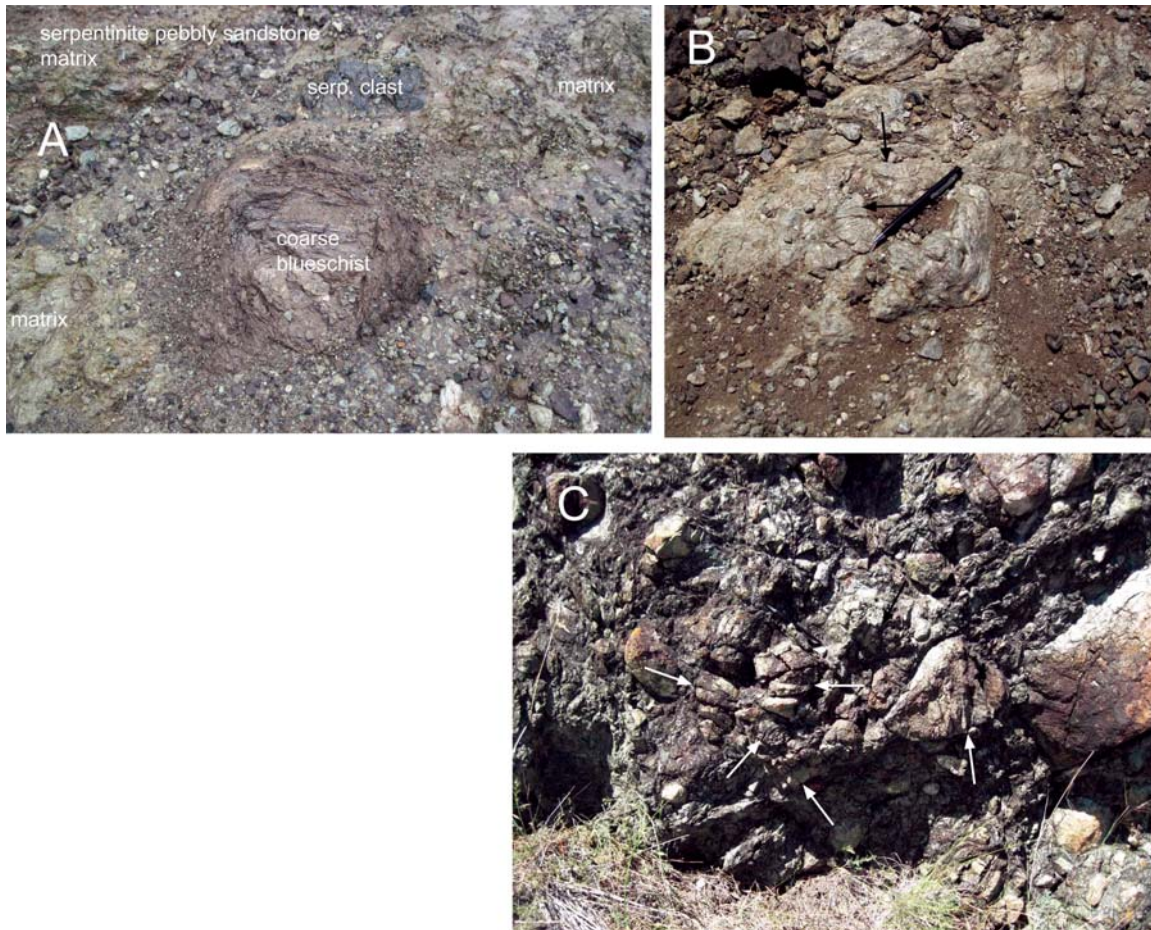


Figure 19. Ring Mountain serpentinite matrix outcrops. (A) Serpentinite conglomerate and breccia surrounding a block of coarse blueschist. Field of view is roughly ~1 m across. (B) Pebbly sandstone near photo shown in A. Note the range in extension crack orientations in the clasts, similar to the nonschistose conglomerate in C. Two examples of crack orientations are noted by arrows, with the arrow oriented parallel to the cracks. The locations of these outcrops are shown on Figure 17, and the location of the outcrops in A and B is also shown on Figure 18. (C) Serpentinite conglomerate that overlies coherent blueschist-facies metagraywacke. Note the different orientations of extension cracks in the rounded clasts, indicating that the cracks did not form in place in this rock. Arrows point to some examples, with the arrows oriented parallel to the cracks. Pencil in middle of photo for scale.

Geochemical studies of blocks at Ring Mountain have spawned widely differing interpretations. Several studies have concluded that the blocks and their rinds record evidence of significant element mobility during subduction metamorphism and chemical exchange between host metabasite blocks and fluids derived from subducted continental sediments (e.g., Nelson, 1991; Sorensen et al., 1997; Catlos and Sorensen, 2003; Penniston-Dorland et al., 2010). In contrast, another group of researchers has concluded that some chemical signatures attributed to metamorphic modification are a product of protolith chemistry and that there is little evidence of chemical exchange between metabasites and fluids derived from subducted sediments (Saha et al., 2005; Wakabayashi et al., 2010; Ghatak et al., 2012).

Many field trip participants may choose to spend their time among the superb high-grade blocks near our vehicles, for these

are the best high-grade blocks on Ring Mountain. Those that wish to see outcrops that illustrate the field relationships may walk a bit further from the vehicles. We will look at two areas that display the field relationships. The first (“area 2” on Fig. 17) will focus on exposures of serpentinite matrix material. We will walk west-northwestward along a walking path that stays high and just north of the crest of the ridge. About 400 m west of our parking area, we will come to the first of two resistant knobs that expose foliated serpentinite matrix. We will continue walking to the second knob a bit more than 100 m further west because exposures are better there.

The exposed matrix consists mainly of schistose serpentinite conglomerate with clasts of primarily serpentinite and meta-ultramafic schist (tremolite-chlorite-antigorite to nearly all tremolite), and some clasts of metamorphosed felsic volcanic rocks. The resistant nature of some of this matrix material results in

resistant outcrops that give the impression of blocks-in-mélange rather than matrix. From this large outcrop, we will continue another 400 m or so to reach an exposure of matrix exposed in the bed of a hiking trail (star for area 2 on Fig. 17; see location for Figs. 19A and 19B on panorama photo Fig. 18A). Here, the matrix exhibits less strain, and the sedimentary character is easily seen. Matrix consists of serpentinite sandstone and conglomerate/breccia, with a block of coarse blueschist embedded in it (Figs. 19A and 19B). This matrix itself has exotic detrital clasts such as glaucophane, hornblende, quartz, rutile, and omphacite in it. There is a similar, but poorer exposure near the top of Ring Mountain (not marked on Fig. 17). Rounding of most of the grains in these sandstones/conglomerates cannot be ascribed to strain, because in many cases these are clast-supported polyolithic rocks.

After viewing the matrix with embedded block, we will take a different route back to the vehicles, staying southwest of the crest of the ridge for much of the distance. During this hike, we will view an exposure of what may be the basal contact of the serpentinite matrix mélange unit (star for area 3 on Fig. 17). In taking this walk, we can see the structural relationships of the various units on Ring Mountain.

The matrix view of Figures 19A and 19B is located at the top of the mélange just beneath the block-free serpentinitized harzburgite cap. The largest continuous exposure of this harzburgite sheet is that capping the hill to our west (Fig. 17). The tectonic stack of units on Ring Mountain has been folded about a sub-horizontal synformal axis oriented parallel to the axis of Tiburon Peninsula (i.e., WNW-ESE). However, this structure does not appear to be symmetric, given the occurrence of several exposures of harzburgite on the southern flank of the Ring Mountain (Fig. 17). It is possible that the complexity of the outcrop pattern is a product of (1) smaller-scale south-vergent overturned folds, with exposure of the structurally lower serpentinite mélange in the cores of antiforms, and/or (2) minor diapiric remobilization of the structurally lower mélange unit upward relative to the harzburgite cap. I am currently conducting further structural studies that I hope will test these two alternatives and give further insight into the kinematics associated with the various nappe contacts.

As we walk east-southeast to the southern side of Ring Mountain, we will reach an area where the stack of nappes appears to be truncated along a NE-striking contact (eastern margin of Fig. 17). This is probably a late fault that postdates subduction accretion of the various tectonic slices. Locally, this places the structurally lowest nappe, comprising prehnite-pumpellyite-facies sandstone and shale, against the serpentinite matrix, and possibly the harzburgite (Fig. 17).

Our hike will curve northward along the southeast slope of Ring Mountain, and we will reach an outcrop where the apparent base of the serpentinite mélange overlies coherent jadeite-bearing metagraywacke (see star for location in area 3, Fig. 19C, on Fig. 17). The serpentinite here appears as a clast-supported serpentinite conglomerate or breccia with both rounded and angular clasts of serpentinitized harzburgite. Extension fractures in clasts show a wide range of orientations and could not have formed by

strain in the present rock; they apparently record deformation in a serpentinite body prior to sedimentation (Fig. 19C).

Collectively, the field relationships at Ring Mountain, as well as regional geologic studies, suggest early metamorphism and exhumation of the high-grade rocks, and then sedimentation of the blocks in a serpentinite-rich olistostrome into the trench where the unit appears to have been deposited on graywacke. The serpentinite matrix mélange and the underlying graywacke were subducted to blueschist-facies depth, with growth of at least antigorite in the matrix, and jadeite, lawsonite, and glaucophane in the graywacke (suggesting reburial to perhaps 30 km or more; e.g., Ernst, 1993). The blueschist-facies graywacke nappe appears to correlate to a similar nappe in El Cerrito, which has a maximum depositional age from detrital zircon U-Pb geochronology of 102 Ma (Snow et al., 2010). This correlation suggests that 65 m.y. of nonaccretion and/or subduction erosion followed initiation of Franciscan subduction for the Franciscan rocks of the San Francisco Bay area. This is even longer than the ~45 m.y. period of nonaccretion that followed the initiation of subduction in the Franciscan units of the northern Coast Ranges (Dumitru et al., 2010). The recognition of periods of nonaccretion or subduction erosion in the Franciscan complex is consistent with observations along modern-day subduction zones, the majority of which show evidence for subduction erosion or lack of accretion (e.g., von Huene and Scholl, 1991; Clift and Vannucchi, 2004).

The next nappe structurally beneath these rocks, consisting of coherent prehnite-pumpellyite-facies rocks, apparently accreted not long afterward, based on a 94 Ma maximum depositional age from a sample along Paradise Drive to the north (T. Dumitru, 2012, personal commun.). Field relationships in the Diablo Range (Sunol Regional Wilderness, and Panoche Road of stop 1) may resemble those in the San Francisco Bay area, with a structurally high sedimentary serpentinite matrix mélange with high-grade blocks (Wakabayashi, 2012a) positionally overlying blueschist-facies graywacke deposited no earlier than ca. 108 Ma, based on published ages of graywackes (Ernst et al., 2009; Unruh et al., 2007), or perhaps a bit earlier (see stop 1 discussion).

At least one unresolved issue is how the early exhumation of the high-grade blocks at this locality and stop 1 took place prior to deposition in the sedimentary serpentinite mélange. Whereas the deposition of the high-grade blocks in the sedimentary serpentinite may have taken place at ca. 100 Ma, as noted earlier herein, encapsulation of the blocks by serpentinite apparently took place at least as early as 149 Ma, based on an Ar-Ar white mica age from one of the rinds of the blocks at Ring Mountain (Catlos and Sorensen, 2003). This rind age is similar to the white Ar-Ar age of ca. 151 Ma obtained by Ukar et al. (2012) from the rind of a lower-grade blueschist block-in-mélange in the San Simeon area along the central California coast exposures of Franciscan complex rocks. Thus, it is unlikely that the blocks in Franciscan serpentinite mélanges were derived from an exhumed coherent high-grade unit. The source of the Franciscan deposits appears to have been an exhumed serpentinite mélange. It is possible that the blocks were originally exhumed in serpentinite diapirs that

fed forearc mud volcanoes, and then they were later reworked from the basal forearc basin strata into the trench by submarine sliding, as noted earlier (Wakabayashi, 2012a).

Driving Directions from Stop 4, Ring Mountain, Tiburon Peninsula, to Stop 5, Rodeo Cove, Marin Headlands (about 14.6 Miles; 23.5 Kilometers)

Note: On the field trip, we will probably make a short detour to Larkspur Landing for lunch, but this road log is written to show distances for the direct route from stop 4 to stop 5. From stop 4, retrace route to US101 and enter US101 southbound toward San Francisco (~2.9 miles [4.7 km] from stop 4). Drive 7.4 miles (12 km) on southbound US101 and exit right (westward) at the Sausalito exit. At the end of this short ramp, make a left (southward), and then almost immediately a right (westward) following signs to Marin Headlands/Fort Cronkhite. Proceed 1.2 miles (1.9 km) along this road, which features great views of the Golden Gate on your left and wonderful road cuts of chert and basalt on your right; then keep right at a roundabout toward Fort Cronkhite/Rodeo Cove. Descend for 0.9 miles (1.4 km), and then make a left at the T-intersection following signs to Fort

Cronkhite/Rodeo Cove. Drive 2.2 miles (3.5 km) westward to the parking area on the south side of the road next to the lagoon (37.83174°N, 122.53594°W, depending on specific parking space). Hike over the lagoon and walk southward to the beach cliffs that we will examine at the stop.

Stop 5. Marin Headlands, Rodeo Cove (~37.83174°N, 122.53594°W)

Similar to Ring Mountain, the rocks of Marin Headlands attract tourists as well as geologists. Most tourists visit the Marin Headlands for its stunning views (when not foggy) of the Golden Gate Bridge and San Francisco, but many enjoy the intricately folded cherts, although they have little idea of the geologic significance of these rocks.

The Marin Headlands (Fig. 20) represent the classic Franciscan locality of ocean plate stratigraphy, consisting of a series of imbricate slices that have repeated a stratigraphic stack of basalt overlain by chert, overlain by graywacke. Steinmann (1905) was the first to recognize the significance of these rocks as part of his ocean crust “Trinity,” decades ahead of geologists in California. Wahrhaftig (1984) demonstrated the imbrication of this ocean

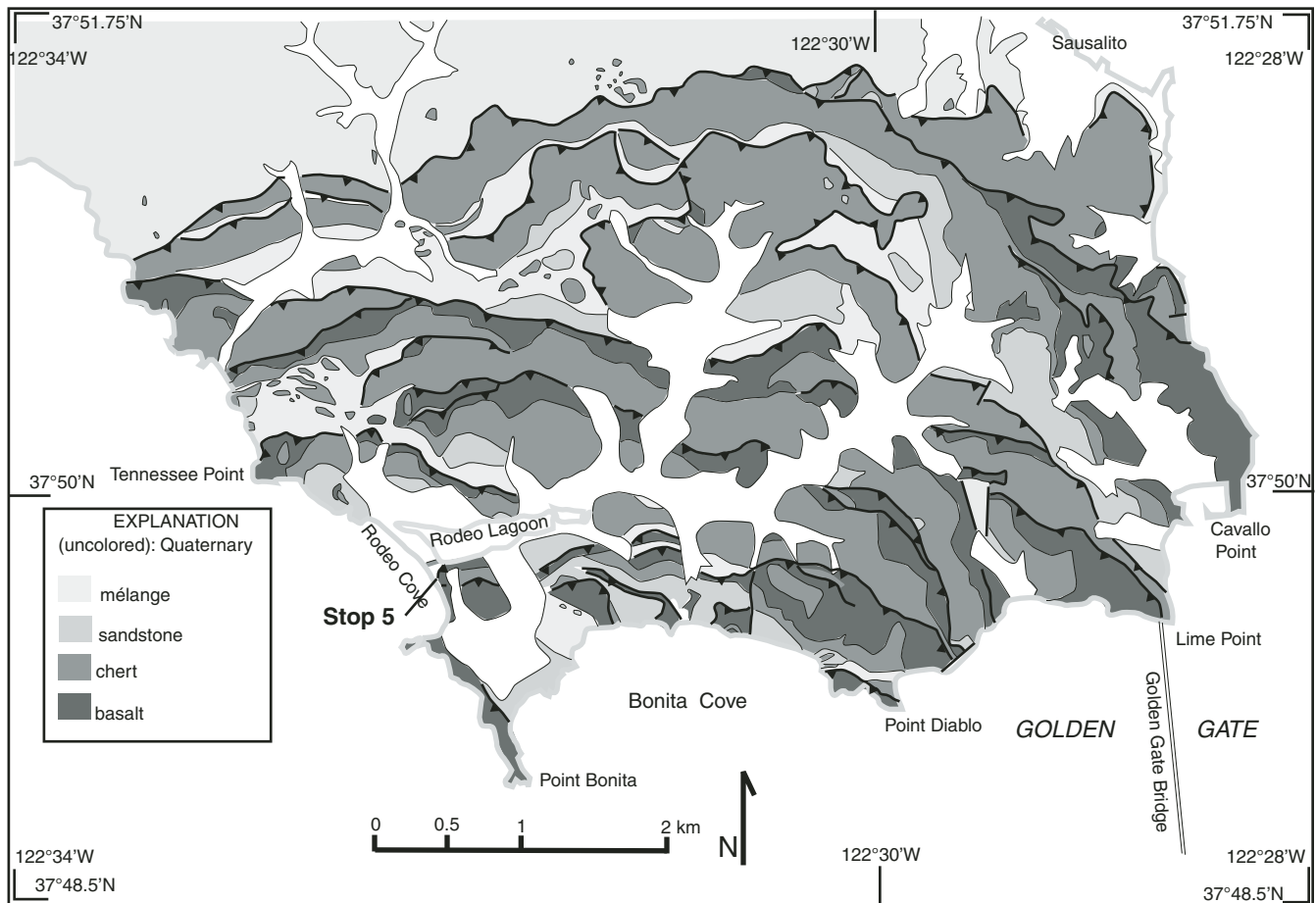


Figure 20. Geologic map of the Marin Headlands area adapted from Kusky et al. (2013). Original map is simplified from Wahrhaftig (1984).

stratigraphy in the subduction complex by showing the structural repetition of chert sequences of equivalent age (biostratigraphy reported in Murchey, 1984). The ocean plate stratigraphy of the Marin Headlands provides an extended rock record of plate tectonics and subduction.

The 80-m-thick chert section spans some 100 m.y. in geologic time from ca. 195 Ma to 95 Ma, and it is overlain by graywacke that has yielded Cenomanian fossils (Murchey, 1984). These cherts rest depositionally upon basalt that exhibits MORB geochemistry (Ghatak et al., 2012). Thus, the Marin Headlands record a history of seafloor spreading at a mid-ocean ridge followed by 100 m.y. of transit across the deep ocean toward the Franciscan subduction zone, a travel distance of ~10,000 km at the estimated plate motions of Engebretson et al. (1985). Upon

arrival at the trench, the graywacke was deposited on the chert, and then the unit was imbricated upon subduction and incorporation into the subduction complex.

Nowhere are the principles of age and structure within an accretionary complex better illustrated than here. The difference between formational age and accretionary age is larger for the Marin Headlands than for any other unit in the Franciscan complex (Fig. 4). In other words, the ocean crust at the time of subduction was oldest when the Marin Headlands accreted than at any other time in Franciscan subduction history.

Whereas most field trips, as well as tourists, visit the overlooks facing the Golden Gate Bridge, we will examine coastal exposures at Rodeo Cove to examine one of the fault zones between imbricate slices (Figs. 20 and 21). This provides a good

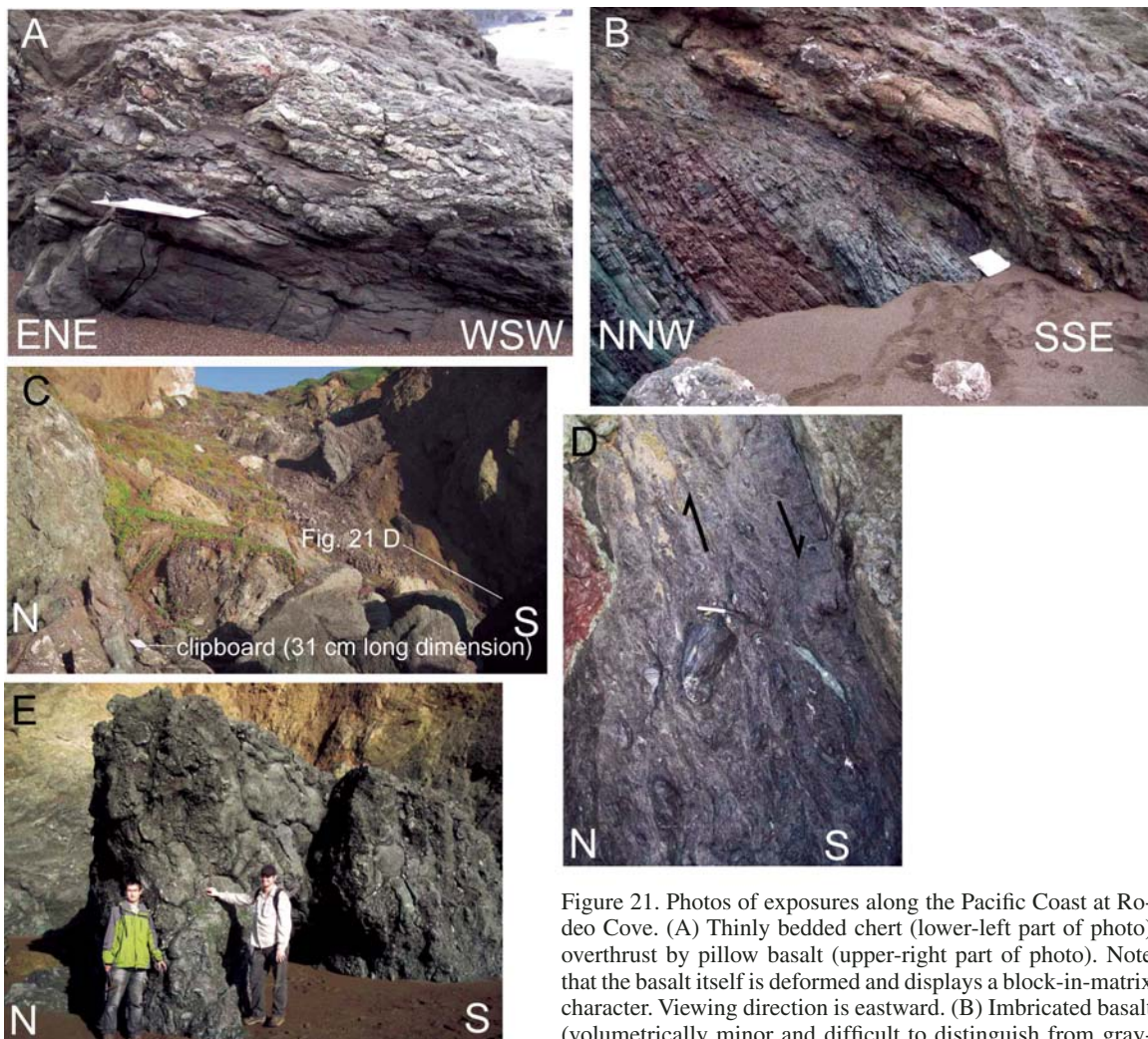


Figure 21. Photos of exposures along the Pacific Coast at Rodeo Cove. (A) Thinly bedded chert (lower-left part of photo) overthrust by pillow basalt (upper-right part of photo). Note that the basalt itself is deformed and displays a block-in-matrix character. Viewing direction is eastward. (B) Imbricated basalt (volumetrically minor and difficult to distinguish from graywacke in this view), graywacke (generally darker shades), and

chert (generally lighter shades). This view shows meter-scale duplex-like structures and the development of a block-in-matrix fabric by fault displacement. The top-to-the-north shear sense seen in A and B is typical of the Marin Headlands block shown in Figure 20. (C) A *mélange* exposure south of A and B with a very different character. Much of this zone is prone to landslides, but shale matrix is exposed along the south border (D), and the fabric in it displays a top-to-the-south shear sense. (E) Pillow basalts exposed south of the *mélange* zone shown in D.

counterexample to the sedimentary mélanges we viewed at the earlier stops. We see duplex structures and block-in-matrix structures formed by faulting rather than submarine sliding (Figs. 21A and 21B). One can see the progression between imbrication and block-in-matrix structures. Most of the block-in-matrix structures exposed along the beach cliffs here consist of chert and basalt blocks in a graywacke matrix, and chert and/or basalt blocks in a basalt matrix (Fig. 21A).

In the Marin Headlands area, the faults repeating the ocean plate stratigraphy strike approximately E-W and dip southward (Fig. 20), in contrast to the regional NW strike of Franciscan structures and (expected) original NE dip. This is a consequence of postaccretionary rotation of the Marin Headlands block with respect to adjacent exposures of Franciscan rocks in the San Francisco Bay area (Curry et al., 1984). Thus, the imbricate faults of the Marin Headlands, including those at Rodeo Cove are north-vergent (Figs. 21A and 21B), rather than the typical southwest vergence of most of the Franciscan complex.

If time and tide permits, we will walk southward along the beach cliff exposures to reach exposures of a different sort of mélangé zone (Figs. 21C and 21D), as well as superb exposures of pillow basalt (Fig. 21E). Much of this mélangé zone is covered by landslide deposits (Fig. 21C), but exposures of the matrix are seen along the southern margin (Fig. 21D). The mélangé strikes approximately E-W and dips steeply south, parallel to the other imbricates. The rare exposures expose shaly matrix with blocks of graywacke, chert, basalt, and serpentinite. The serpentinite marks this mélangé as different from the other block-in-matrix zones observed on the beach because it is not a part of the flanking coherent slabs; the serpentinite blocks can be considered “exotic.”

The sense of shear in the matrix along the southern boundary of the mélangé is top-to-the-south (Fig. 21D), antithetic to the sense of shear associated with the ocean plate imbricates seen throughout the Marin Headlands and seen in Figures 21A and 21B. Unfortunately, matrix was not exposed along the northern margin of the mélangé zone. On the southern edge of the intact rocks on the north border of the mélangé (at about the location of the clipboard in Fig. 21C), shears show a top-to-the-north sense of shear. However, it is not clear whether this sense of movement is related to diapiric movement of the mélangé (which would require the opposite sense of shear on opposite borders of the mélangé) or whether the shears along this margin are part of the top-to-the-north sense of shear that is common throughout the Marin Headlands block as seen in Figures 21A and 21B. Slickenlines and mullions associated with the exposure shown in Figure 21D plunge at 30° to the west, so the movement on the southern margin of the mélangé has a strong dextral component.

Lacking the exposures on the north side of this mélangé zone, there are at least three alternatives for the origin of the deformation seen on its south margin: (1) diapiric movement of the mélangé, which predicts that exposures of matrix along the northern margin would have a top-to-the-north sense of shear; (2) dextral strike-slip movement, which predicts that the sense

of shear should be the same (top-to-the-south or dextral) across the mélangé; or (3) normal faulting, which also predicts a uniform top-to-the-south sense of shear across the zone. Whether or not the block-in-matrix fabric originally formed via sedimentary process (to emplace the exotic serpentinite blocks) or via tectonic or diapiric movement alone cannot be determined; the only exposure of the matrix (shown in Fig. 21D) is too deformed to see any trace of sedimentary structure.

Driving Directions for Return to Fresno

Return by retracing driving route across the Richmond–San Rafael Bridge to southbound I-80, and then eastbound I-580 to leave the San Francisco Bay area. Continue over Altamont Pass. Stay on I-580 (exit right) rather than continue on I-205. Exit SR132 east to Modesto, and then take SR99 southward to Fresno (distance is ~210 miles; 338 km).

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