Detrital zircon U–Pb reconnaissance of the Franciscan subduction complex in northwestern California

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In northwestern California, the Franciscan subduction complex has been subdivided into seven major tectonostratigraphic units. We report U–Pb ages of ≈2400 detrital zircon grains from 26 sandstone samples from 5 of these units. Here, we tabulate each unit’s interpreted predominant sediment source areas and depositional age range, ordered from the oldest to the youngest unit. (1) Yolla Bolly terrane: nearby Sierra Nevada batholith (SNB); ca. 118 to 98 Ma. Rare fossils had indicated that this unit was mostly 151–137 Ma, but it is mostly much younger. (2) Central Belt: SNB; ca. 103 to 53 Ma (but poorly constrained), again mostly younger than previously thought. (3) Yager terrane: distant Idaho batholith (IB); ca. 52 to 50 Ma. Much of the Yager’s detritus was shed during major core complex extension and erosion in Idaho that started 53 Ma. An Eocene Princeton River–Princeton submarine canyon system transported this detritus to the Great Valley forearc basin and thence to the Franciscan trench. (4) Coastal terrane: mostly IB, ±SNB, ±nearby Cascade arc, ±Nevada Cenozoic ignimbrite belt; 52 to <32 Ma. (5) King Range terrane: dominated by IB and SNB zircons; parts 16–14 Ma based on microfossils. Overall, some Franciscan units are younger than previously thought, making them more compatible with models for the growth of subduction complexes by progressive accretion. From ca. 118 to 70 Ma, Franciscan sediments were sourced mainly from the nearby Sierra Nevada region and were isolated from southwestern US and Mexican sources. From 53 to 49 Ma, the Franciscan was sourced from both Idaho and the Sierra Nevada. By 37–32 Ma, input from Idaho had ceased. The influx from Idaho probably reflects major tectonism in Idaho, Oregon, and Washington, plus development of a through-going Princeton River to California, rather than radical changes in the subduction system at the Franciscan trench itself.

Keywords: Franciscan; subduction; zircon; Sierra Nevada; Idaho batholith; Challis; Great Valley; Tyee

Introduction

Since the advent of plate tectonic theory, the Franciscan Complex of California (Figure 1) has been regarded as a classic example of a subduction-generated accretionary prism (e.g. Hamilton 1969; Dickinson 1970; Ernst 1970; Wakabayashi 2015). From Middle(?)-Jurassic time up until the present, more than 10,000 km of the oceanic Farallon ± Kula plates have subducted eastward beneath the North American continental edge at the Franciscan trench (Engebretson et al. 1985). This convergence generated the Sierra Nevada–Klamath magmatic belt, a massive Andean-style arc built along the western edge of the continent, which supplied abundant detritus to both the Great Valley forearc basin and the Franciscan trench (Dickinson 1970, 2008; Williams and Graham 2013). These clastic sediments, plus lesser volumes of far-travelled oceanic basalt and traces of pelagie sediments, accreted above the subduction zone to build the accretionary prism (e.g. Blake et al. 1988; Wakabayashi 2015). The subduction process accounts for the Franciscan’s highly tectonized structure, high-pressure/low-temperature metamorphism (Ernst 1971), and juxtaposition of sparse oceanic rocks with voluminous arc-derived clastic sedimentary rocks. This tectonic environment also accounts for the striking contrast between the intensely deformed Franciscan rocks that accumulated near the trench and the weakly deformed, coeval forearc basin sandstone and mudstone strata laid down atop the more stable forearc of the North American continent.

The Franciscan geology of the northern part of northern California consists of a series of W-vergent, dominantly metasedimentary thrust sheets. Jayko et al. (1989), Blake et al. (1992, 1999), and McLaughlin et al. (2000) presented 1:100,000-scale geologic maps and data from the Covelo, Willows, Red Bluff, Garberville, Cape Mendocino, and Eureka 30° × 60′ quadrangles that document these thrust sheets in northwestern California. Built on previous work of Blake et al. (1988), McLaughlin et al. (1994), and many others, these maps provide a valuable framework for further studies. Franciscan structural relationships in this region have been less overprinted and obscured than those farther south by strike-slip faulting.
associated with the late Cenozoic, post-subduction San Andreas transform plate boundary (e.g. Wakabayashi 2015).

Figure 2 presents a simplified map of the Cape Mendocino, part of the Garberville, and part of the Covelo quadrangles within this representative tract of the Franciscan Complex. Working in this area, Ernst and McLaughlin (2012) reported detailed petrographic and analytical investigations of 88 Franciscan metasandstone
introduced U-Pb age data from 6 of these samples to reconstruct Eocene sediment source areas and transport pathways. In this new article, we present zircon data from 20 additional samples (Figure 2), with interpreted depositional ages between ca. 118 and 15 Ma, and employ these data to constrain the depositional ages and sediment provenance of Franciscan clastic rocks in the area.

Franciscan geology of northwestern California

Franciscan rocks are chiefly strongly deformed mudstones and sandstones sourced from the continent, with lesser amounts of oceanic basalt and pelagic sedimentary rocks, all subjected to very low to moderate grades of metamorphism (Figure 3). In most Franciscan sedimentary rocks, megafossils are very rare or absent and microfossils have been underutilized, so knowledge of depositional

![Figure 2. Generalized geology of the Cape Mendocino, part of the Garberville, and part of the Covelo 1:100,000 quadrangles, simplified after Jayko et al. (1989) and McLaughlin et al. (2000) (see also Blake et al. 1988). Major terrane boundaries are chiefly interpreted as gently E-dipping thrust faults. Locations of 21 detrital zircon samples are indicated; symbols distinguish the 5 zircon age distribution patterns ('Types') exhibited by different samples, as discussed later (e.g. Figure 6J–N). Also shown are locations of six previous detrital zircon fission track samples in the Westport-Laytonville area (Tagami and Dumitru 1996). Fission track data are much less precise, so, for these samples, it is not possible to distinguish type A1 from A2 or type C from D zircon age distributions.](image-url)
ages is generally quite incomplete (e.g. Evitt and Pierce 1975; Blake et al. 1988; Murchey and Blake 1993; McLaughlin et al. 1994, 2000; Lucas-Clark 2007). The oceanic basalts in the Franciscan cannot be dated using current isotopic methods. Intrusive rocks are very rare within the Franciscan and only four small intrusions have been dated (Meyer and Naeser 1970; Mattinson and Echeverria 1980; Mertz et al. 2001).

In northern California, the Franciscan Complex has been subdivided into three major, fault-bounded belts, the Eastern, Central, and Coastal Belts (Figure 1; Irwin 1960; Bailey et al. 1964; Berkland et al. 1972). In the Red Bluff to Cape Mendocino area (Figure 2), the Eastern Belt was subsequently divided into two terranes and the Coastal Belt into four terranes (e.g. Blake et al. 1985b, 1988, 1999; Jayko et al. 1989; McLaughlin et al. 1994, 2000). The Central Belt has been subdivided into terranes in some parts of California, but Central Belt terranes are not formally recognized in our study area (McLaughlin et al. 2000). Therefore, in our study area, we assume that the Franciscan Complex consists of seven major tectonostratigraphic units, six terranes plus the Central Belt. This article presents zircon data from five of these units, omitting the Pickett Peak terrane on the far east (see Dumitru et al. 2010) and the small False Cape terrane on the far west. To the northeast, the Klamath Mountains–Franciscan boundary is an E-rooting thrust fault, as are tectonic contacts juxtaposing the belts and terranes within the Franciscan Complex itself. Each mega-allochthon appears to have been assembled as an imbricate stack of indistinctly defined thrust sheets (Ernst and McLaughlin 2012) that are interpreted as roof thrusts in a series of eastwardly underthrust structural wedges (McLaughlin et al. 2000, sheet 6). Petrofacies analyses and bulk-rock geochemistry indicate that Franciscan sedimentary rocks were derived chiefly from landward volcanic-plutonic arc(s) (Dickinson et al. 1982; Seinders 1983; Underwood and Bachman 1986; Ghatak et al. 2013), similar to the sources of the coeval strata of the directly onshore Great Valley forearc basin (Ingersoll 1979, 1983; Linn et al. 1992; Surpless 2015). The simple interpretation that Franciscan sedimentary rocks were derived from nearby Sierra Nevada–Klamath arc rocks was questioned and re-interpreted to reflect far-travelled, northward translation for some Franciscan units (e.g. Blake et al. 1984; Jayko and Blake 1984; Hagstrum and Murchey 1993). Both interpretations may be partially correct, inasmuch as former magmatic arc sources extended for a considerable distance further south along the North American margin than the realm of the Sierra Nevada arc.

In a simple progressive accretion model for the long-term growth of accretionary prisms (Seeley et al. 1974; Ernst 1975; Cowan and Silling 1978), packets of rocks generally young seaward (away from the arc) and structurally downward, as successively younger trench sediments underthrust older sediments previously incorporated into the wedge. Below, we briefly introduce the seven Franciscan units in the study area from east to west, presumably equivalent to older to younger and structurally higher to structurally lower. Here, we include only a brief synopsis of previous age information obtained from fossils and detrital zircons, then add details after presenting our new zircon data.

Blake et al. (1988, 1999) subdivided the Eastern Belt into two packages, the Pickett Peak and the Yolla Bolly terranes. Pickett Peak rocks were probably subducted, accreted, and metamorphosed to blueschist facies at about 123–117 Ma (Lanphere et al. 1978; Dumitru et al. 2010). The structurally lower Yolla Bolly terrane, which crops out over about 6000 km², has yielded megafossils from roughly 20 localities with ages from Tithonian to Cenomanian (ca. 151–94 Ma; e.g. Blake et al. 1988; Jayko et al. 1989; Ohlin et al. 2010).
Recent zircon work, discussed later, generally favours a younger depositional age range, probably ca. 117–98 Ma (Ernst et al. 2009; Dumitru et al. 2010; Dumitru 2012; this article). Ernst and McLaughlin (2012) estimated that the area of the Yolla Bolly terrane they studied (samples 40 and 43 in Figure 2) was metamorphosed at peak temperatures of 200–250°C and at peak pressures of ~8 kbar (Figure 3). However, McLaughlin et al. (2000) tentatively included that area within the higher-grade Taliaferro metamorphic complex (Suppe 1973), and T and P may have been somewhat lower in the remainder of the Yolla Bolly (e.g. Dumitru et al. 2010, Figure 4).

Blake et al. (1988) suggested that the Yolla Bolly terrane was metamorphosed at ca. 92 Ma, based on a U-Pb isochron age for metamorphic titanite and plagioclase from the Diablo Range in central California (Mattinson and Echeverria 1980). Dumitru et al. (2010) preferred a metamorphic age of ca. 110 Ma for the Pickett Peak–North Yolla Bolly area in northern California, based on five argon whole-rock ages of Lanphere et al. (1978), and speculated that different parts of the terrane may have been metamorphosed at different times.

As mentioned, rocks of Central Belt affinity have been subdivided into terranes in parts of California (Blake et al. 1984, 2000; see also Wakabayashi 2015), but terranes have not been formally designated in our study area (McLaughlin et al. 2000). The belt is characterized by abundant, feebly recrystallized, strongly sheared mélangé. The matrix of the mélangé is argillite with variable amounts of sandy and mafic tuffaceous material. Peak metamorphic conditions were ~150–250°C and up to ~4.5 kbar (Figure 3; Cloos 1981, 1983; Blake et al. 1988; Terabayashi and Maruyama 1998; Ernst and McLaughlin 2012). The matrix encases blocks, boudins, and slabs of relatively coherent metasandstone, argillite, metachert, greenstone, and rare higher-grade blueschist, eclogite, and amphibolite that range in size from centimetres to a few kilometres (e.g. Cloos, 1982, 1983; 1986; Terabayashi and Maruyama 1998; McLaughlin et al. 2000). A few of the blocks (basalts, pelagic cherts, pelagic limestones) have been identified as fragments of oceanic terranes that are formally defined in the San Francisco Bay region, based on correlative foraminiferal and/or radiolarian faunas (Sliter 1984; Murchey and Jones 1984; McLaughlin et al. 1994, 2000, Table 1 on Sheet 5). Mélange matrix also surrounds several informally named slabs of relatively coherent metasandstone and argillite up to about 5 km in width (Figure 2; Gucwa 1974, 1975; Jayko et al. 1989; McLaughlin et al. 2000), similar to formally named sandstone–mudstone terranes in the Bay region. Some of the blocks and slabs have been identified as fragments of older Eastern Belt units or the Coast Range Ophiolite. The northeastern part of the belt is transected by steep, NNW-trending, dextral, strike-slip faults (some presently active) that have displaced Central Belt rock bodies (Herd 1978; McLaughlin et al. 2000; Ohlin et al. 2010).

Age relations in the Central Belt are very complex. Focusing near our study area, roughly 12 samples from probable mélange matrix have yielded microfossils with essentially Cretaceous ages (145–65 Ma), with a possible clustering in the Albian (112–98 Ma; Gucwa 1974, 1975; O’Day 1974; Lucas-Clark 1986, 2007; Larue 1986; McLaughlin et al. 2000). However, essentially all megafossils in the mélange are Tithonian to Valanginian (151–137 Ma; Suppe 1969, 1973; Irwin et al. 1974; Blake and Jones 1974; Lehman 1974; Jordan 1978). Two samples from the far western side of the Central Belt probably yield Palaeocene–Eocene microfossils (but see Table DR1; see http://dx.doi.org/10.1080/00206814.2015.1008060). Some of the sandstone slabs yield limited Early Cretaceous, mid-Cretaceous, and Late Cretaceous microfossils (Berkland 1972, 1973, 1978; Gucwa 1974, 1975; O’Day 1974; Evitt and Pierce 1975; Larue 1986), whereas one slab probably yields Palaeocene–Eocene microfossils (but see Table DR1). Various blocks of pelagic radiolarian chert range in age from ca. 172 (McLaughlin et al. 2000) and blocks of pelagic foraminiferal limestone from Albian (112–103 Ma) to early Coniacian (88–85 Ma) (Alvarez et al. 1980; Sliter 1984; Tarduno et al. 1986, 1990; McLaughlin et al. 2000).

The Coastal Belt has been subdivided into the Yager, Coastal, King Range, and False Cape terranes (Underwood and Bachman 1986; McLaughlin et al. 1994, 2000; Langenheim et al. 2013). Peak metamorphic conditions reported by Ernst and McLaughlin (2012) were 50–135°C and 1–2 kbar (Figure 3). The Yager terrane is predominantly massive, coherent, weakly metamorphosed feldspathic sandstone and mudstone. Dinoflagellates, pollen, and spores suggest that the unit is mainly Eocene (O’Day 1974; Evitt and Pierce 1975; Bachman 1978; Damassa 1979a, 1979b; Frederiksen 1989; McLaughlin et al. 1994, 2000; summary in Dumitru et al. 2013, Table DR1). The Coastal terrane is mainly weakly recrystallized broken formation and minor mélange made up of medium–fine-grained arkosic to lithic sandstone and mudstone, including sparse blocks and slabs of pillow basalt associated with foraminiferal limestone. Microfossils from the clastic section suggest that its depositional age is mostly Eocene (O’Day 1974; Evitt and Pierce 1975; Bachman 1978; Berry 1982; Sliter et al. 1986; Frederiksen 1989; Jayko et al. 1989; McLaughlin et al. 1994, 2000; summary in Dumitru et al. 2013, Table DR1). One locality south of our report area (McLaughlin et al. 2013) exposes basaltic rocks with intercalated and overlying pelagic foraminiferal limestone of Campanian–Maastrichtian age (84–65 Ma), with the uppermost part of the limestone section as young as late middle Eocene (ca. 45 ± 4 Ma).
The King Range terrane has been divided into the Point Delgada and King Peak subterrane (McLaughlin et al. 1982, 1985, 1994, 2000). The Point Delgada subterrane includes pillow basalt associated with minor limestone and with dark red argillite containing Campanian to Coniacian radiolarians (71–89 Ma). These oceanic rocks apparently accumulated well offshore, isolated from sediment input from the continental margin. The basalts are overlain by a thin, undated mélange, which in turn is structurally overlain by undated turbiditic arkosic metasandstone and argillite. Hydrothermal veining dated at 13.8 Ma overprints the entire Point Delgada subterrane. The King Peak subterrane comprises mainly highly folded arkosic sandstone and mudstone turbidites. Vitritine reflectance data indicate thermal overprinting by the 13.8 Ma event (Underwood et al. 1999), indicating that the Point Delgada and King Peak subterrane amalgamated to form the King Range terrane by 13.8 Ma. Radiolarians and foraminifera from the King Peak subterrane suggest a middle Miocene depositional age (ca. 16–14 Ma, predating veining). King Peak subterrane strata may have been deposited roughly 500–700 km SSE of their current location, then may have been translated NNW and then eastward within the San Andreas system, as the King Range terrane accreted to older parts of the Franciscan after 13.8 Ma.

The False Cape terrane comprises complexly folded, hard, black diatomaceous mudstone, sandstone, and minor pink pelagic limestone. Diatoms and foraminifera suggest late Oligocene to early Miocene ages (Aalto et al. 1995, 1998; McLaughlin et al. 2000).

Zircon U-Pb age data

Methods and data presentation

Ernst and McLaughlin (2012) reported detailed petrographic and electron microprobe analyses of 88 Franciscan metasandstone samples from our study area. We separated detrital zircons from 19 of those samples, plus 3 samples collected in late 2012 and 1 collected in 1989, using standard methods (e.g. Dumitru 2000, Table 2). G.E. Gehrels also generously shared previously unpublished zircon data from three additional samples. Analyses of 6 samples (557 total grains) were completed at the University of Arizona LaserChron Center (ALC) and reported by Dumitru et al. (2013). U and Pb isotopic measurements were performed using a Nu HR multicollector inductively coupled plasma mass spectrometer (ICPMS) coupled to a New Wave UP193HE excimer laser ablation system (e.g. Gehrels et al. 2006, 2008; Gehrels 2012). Zircon grains were selected for dating at random. For ALC zircon grains younger than 1000 Ma, we employ Pb/238U ages that have been corrected for common Pb using Pb measurements. For the relatively small number of grains older than 1000 Ma, we employ Pb-corrected Pb/206Pb ages (e.g. Gehrels et al. 2006, 2008; Gehrels 2012).

Seventeen samples (~1600 total grains) were analysed at the University of California, Santa Cruz (UCSC), using a Thermo Scientific Element XR single-collector, high-resolution, magnetic-sector ICPMS coupled to a Photon Machines Analyte H 193 nm ArF excimer laser system (e.g. Sharman et al. 2013). Grains were selected at random. Data were reduced and ages were calculated using the Iolite software package (Paton et al. 2010), followed by the spreadsheet of Sharman et al. (2013). For UCSC grains younger than 1000 Ma, we employ Pb-corrected Pb/206Pb ages. For the relatively small number of grains older than 1000 Ma, we employ uncorrected Pb/204Pb ages, calculated after first trimming out data from any high-Pb domains within the grains using Iolite.

G.E. Gehrels (written communications, 2014) provided data from three samples (300 total grains) analysed ca. 2003. These data were collected at the University of Arizona using a Micromass Isoprobe multicollector ICPMS coupled to a New Wave DUV193 excimer laser. Grains were selected at random. All grains were younger than 371 Ma and we employ Pb-corrected Pb/206Pb/238U ages. These samples were also analysed by single-grain thermal ionization mass spectrometry (TIMS); the TIMS data are included in the data repository but are omitted from our plots.

Data tables and additional method details are available in the data repositories of Dumitru et al. (2013) and this article (Appendix DR1; Tables DR3–DR5). Our sample-numbering scheme is explained in Figure 4A. Sample coordinates for eight samples are listed in Table DR2. Coordinates for the 18 remaining samples are listed in Ernst and McLaughlin (2012), except that, to correct minor errors, the longitude of CR-75 should read 124°03.0252' W and the latitude of CR-90 should read 39°49.964' N. Ernst and McLaughlin (2012) reported modal composition data for 88 metasandstone samples. Table DR6 presents modal data for 91 total samples and corrects minor typographical errors discovered in the previously published table.

Data for all 26 samples are presented in Figure 4. Three types of data are presented for each sample: (1) a probability density plot (PDP) of single grains ages (Ludwig 2008; Gehrels 2012) with a change in the age scale at 300 Ma that has proven to be helpful in displaying data from the western USA (Grove et al. 2008; Sharman et al. 2014); (2) a plot of the youngest 40 grains (Ludwig 2008), useful for estimating the youngest zircon population (YZP) in the sample, and thus the maximum depositional age of the rock; and (3) a synopsis interpretation of the data. We rely on these synopses to avoid a tedious sample-by-sample interpretation within the text. The ages of the YZP in each sample were estimated by visual
methods similar to those of Dumitru et al. (2010, Table S2; see also; Dickinson and Gehrels 2009b).

Table DR2 includes additional information for eight of the samples that would not figure in Figure 4. Examination of the data shows that samples tend to cluster into five main types with similar age patterns. The ternary plot in Figure 5 shows these types and Figure 6J–N shows composite PDP and synopses for each type. For comparison, we also show composite PDP for forearc basin sandstones in California and Oregon, PDP for Jurassic erg sandstones on the Colorado plateau, and age ranges of potential source regions (Figure 6A–I).

Age patterns in sediment source regions

The synopses in Figure 4 show that the interpreted depositional ages of our samples fall between 118 and 14 Ma. Numerous potential source regions exist for Franciscan and Great Valley sandstones of such ages. Locations of these potential sources are shown in Figure 7 and their age ranges are shown in Figure 6A–C. To anticipate our results, in this section, we first describe sources that we actually recognize in at least one of our samples. The next section briefly describes sources that apparently did not supply detritus to our samples.

Sierra Nevada batholith (SNB). The major Sierra Nevada batholith lies east of the Great Valley forearc basin and the Franciscan Complex. Intrusion rates varied greatly over time, with relatively high rates ca. 170–145 Ma and yet higher rates ca. 120–87 Ma (e.g. Barton et al. 1988; Saleeby et al. 1989; Irwin and Wooden 2001; DeGraaff-Surpless et al. 2002; Chapman et al. 2012). Figure 6A includes a histogram of magmatic ages (Sharman et al. 2014), which illustrates these variations. Periods of more intense plutonism probably generate a stronger zircon signal in derived sediments. Important shifts in the location of the arc also took place over time. The ‘Jurassic arc’ (more precisely, 200 to 136 Ma) trended NNW–SSE, whereas the ‘Cretaceous arc’ (136 to 85 Ma) trended closer to N–S and crosscut the Jurassic arc trend. In the northern Sierra Nevada, Jurassic arc rocks lie west of the Cretaceous arc rocks,
whereas the reverse prevails in the southern Sierra Nevada, the Mohave region, and the Peninsula Ranges batholith (e.g. Sharman et al. 2014). This has important effects on the ages of zircons shed to the west towards the Franciscan and Great Valley. Wall rocks of the batholith and associated accreted Palaeozoic terranes host smaller numbers of Palaeozoic to Archaean zircons (Sharman et al. 2014; Surpless 2015).

Klamath Mountains (included within SNB source). Circa 170 to 136 Ma plutons are common in the Klamath Mountains and represent the northern continuation of the ‘Jurassic’ Sierra Nevada arc, so we include Klamath rocks as part of our SNB source area. The younger ‘Cretaceous arc’ intruded far to the east, bypassing the Klamath salient. Most pre-170 Ma rocks in the Klamaths are zircon-poor ophiolite-chert-argillite sequences; some of the pre-170 Ma sequences correlate with units in the foothills belt of the western Sierra Nevada (e.g. Irwin and Wooden 1999; Allen and Barnes 2006; Ernst et al. 2008; Van Buer et al. 2009; Ernst 2013; Sharman et al. 2014).

Idaho batholith (IB) and adjacent areas. Magmatism in the SNB peaked at 88 Ma, then waned and ceased at

![Figure 4](https://example.com/figure4.png)

Figure 4. (Continued).
80 Ma. However, magmatism continued in the Idaho batholith. The IB has an overall age range from 98 to 53 Ma and was emplaced in a contractional setting during Sevier–Laramide orogenic thrusting (Gaschnig et al. 2010, 2011, 2013). Moderate volumes of 98–85 Ma granitic rocks are present and may have been more extensive before later assimilation into younger plutons. The Atlanta lobe produced the south part of the batholith from 82 to 67 Ma, followed by the Bitterroot lobe on the north at 65 to 53 Ma. The general IB region also shed older zircons (Figure 6B), although extensive cover by Miocene volcanic rocks obscures characterization of pre-Miocene surface exposures. Small plutons associated with the west Idaho shear zone probably range from ca. 125–97 Ma (Gaschnig et al. 2010, 2011). The Eocene Tyee Formation in coastal Oregon contains about 10% 125–100 Ma detrital zircons.
Figure 5. Ternary plot of single grain zircon ages from our Franciscan samples, plus samples from the Eocene Tyee forearc basin in Oregon. End member poles are proportions of zircon grains with ages of 45–80 Ma, 85–120 Ma, and 130–210 Ma. These end member age ranges were selected to distinguish zircons from the Idaho batholith + Challis, Cretaceous Sierra Nevada arc, and Jurassic Sierra Nevada arc. Other ages were discarded because they are less diagnostic (e.g. 120–130 Ma, >210 Ma). Note that the Idaho region also shed considerable 85–120 Ma zircon, so a sample sourced 100% from Idaho is not expected to plot at the 45–80 Ma pole. Different symbols are used for the YZP of each sample. The ternary plot plus YPZ define five types of zircon age distributions (plus a sixth type for Tyee strata). More details on these types are shown in Figure 6J–N. 

Eocene core complexes and Challis volcanic-plutonic complex. IB magmatism in a contractional setting ceased at 53 Ma and was abruptly superseded by major extensional exhumation in the Bitterroot, Anaconda, Clearwater, and Priest River metamorphic core complexes (53–40 Ma; Foster et al. 2007). This extensional deformation apparently uplifted a broad region (Figure 7) and shed large volumes of sediments to the west and southeast (Heller et al. 1985; Chetel et al. 2011; Dumitru et al. 2013). The widespread Challis volcanic-plutonic complex was associated with this extension; it intrudes and overlies large parts of the IB, although it tends to concentrate east of the main Atlanta lobe. Magmatism started at 51 Ma, peaked at 47 Ma, and ended at 43 Ma (Gaschnig et al. 2010, 2011; Chetel et al. 2011). Challis volcanic rocks were probably initially much more extensive than shown in Figure 7, but have since been eroded away.

Cascade magmatic arc. After the Challis event, arc magmatism was established in the new Cascade arc to the west, becoming moderately voluminous by ca. 35 Ma and continuing to the present day (Du Bray and John 2011; see also Colgan et al. 2011). Rocks of 35 to 26 Ma age were mainly basalt, basaltic andesite, and andesite. Such rock types generally have much lower zircon fertility than granitic rocks.

Nevada ignimbrite flare-up. Major intermediate to silicic eruptions occurred in Nevada from ca. 45 to 8 Ma, with particularly intense activity between 37 and 22 Ma (Henry and John 2013). Ash flow tuffs from some of these eruptions travelled far to the west and remnants of tuffs dated from 32 to 25 Ma are preserved in palaeocanyons on the western flank of the northern Sierra Nevada. For at least four tuffs, eruptive volumes exceeded 1000 km³. Cassel et al. (2012) dated 17 zircon grains with ages of ca. 44 to 33 Ma in one Eocene sandstone sample from the northern Sierra.

Redeposition from older Franciscan and Great Valley rocks. Mass wasting deposits (i.e. olistostromes) are common in the Franciscan and most probably represent erosion and redeposition of older Franciscan and/or Great Valley rocks (e.g. Dickinson et al. 1982; Aalto 2014; Platt 2015; Wakabayashi 2015). The overall proportion of Franciscan rocks that are redeposited is not known. Some Franciscan sedimentary rocks presumably consist almost entirely of redeposited materials whereas others are mixtures of redeposited and primary materials. Such rocks may yield zircon data that are confusing to interpret.

Patterns in regions that were not major sources for our samples

Southwest US Proterozoic basement. Basement in the southwestern USA belongs mostly to the 1.6–1.8 Ga Yavapai and Mazatzal provinces, intruded by 1.30–1.53 Ga anorogenic granitoids (Figure 6C). Parts of
Figure 6. Comparison of zircon age distributions in potential sediment source areas (A–C), in our Franciscan samples (J–N), and in other basins in the western USA (D-I). Colours highlight interpreted major source areas for zircons of various ages. Note 10× change in the age scale at 300 Ma.
this basement in Arizona and southwestern California were exposed to erosion over the interval 95–14 Ma (e.g. Dickinson and Gehrels 2008; Jacobson et al. 2011; Dickinson et al. 2012; Sharman et al. 2014). Some of our samples contain limited numbers of zircons with such ages. However, zircons of such ages are also present in metasedimentary basement terranes of the Sierra Nevada and Klamath Mountains, as well as in Idaho (Figure 6), and we tentatively infer that those areas were their likely sources (Surpless 2015, p. 755–757).

Jurassic erg deposits. An aeolian sand sea covered much of the present-day Colorado Plateau during Jurassic time. Redeposition of these sands was a major source for Cretaceous and Palaeocene strata in the foreland basin of the Cordilleran orogen to the east of the Sierra Nevada arc and the Sevier foreland thrust belt (Dickinson and Gehrels 2008, 2009a; Dickinson et al. 2012). The erg sands were originally sourced mainly from the eastern USA via transcontinental rivers and contain prominent age peaks (Figure 6I) at ca. 420 and 615 Ma (Appalachian sources) and 1055 and 1160 Ma (Grenville), plus lesser peaks from central and western US sources at 1465 Ma (anorogenic granitoids), 1675–1855 Ma (Yavapai and Mazatzal; suture belts), and 2760 Ma (Archaean Craton).

Nevada back arc region. Before extensive burial by the ignimbrite flare-up and other Cenozoic volcanic rocks, sedimentary rocks exposed in Nevada included those making up the Roberts Mountain allochthon, the Golconda allochthon, the Cordilleran miogeocline, and the western Nevada Triassic back arc basin (Van Buer et al. 2009; Long 2012). Considered together, the first three of these units are dominated by 1.7–2.0 Ga detrital zircons, with lesser amounts of older and younger zircons (Gehrels et al. 1995, 2000; Gehrels and Dickinson 2000; Riley et al. 2000; Harding et al. 2000; Spurlin et al. 2000; see also Cecil et al. 2010, Figure 5). The back arc basin (Wyld 2000) contains zircon ages similar to the Jurassic erg deposits and apparently was similarly sourced (Dickinson and Gehrels 2008). In the latest Cretaceous–early Palaeocene, the N-trending palaeodivide between W- and E-draining rivers may have run near the California–Nevada border, then migrated to central Nevada by the mid-Eocene (Figure 7; DeGraaff-Surpless et al. 2002; Van Buer et al. 2009; Henry et al. 2012; Sharman et al. 2014).

85 to 45 Ma volcanic air fall. Non-basaltic tuffs are extremely rare in Franciscan and Great Valley strata. This suggests minimal potential for plumes from 85 to 43 Ma volcanoes (Laramide, etc.) to supply zircons directly to the Franciscan.

Archaean cratonal core. Basement in Montana, Wyoming, and adjacent areas is chiefly older than 2.5 Ga. From 95 to 14 Ma, most areas did not shed detritus towards the west, due to cover by younger strata, subsidence beneath the Cretaceous interior seaway, and/or known drainage pathways in other directions.

Age patterns in nearby basins

Franciscan zircon age patterns need to be compared to previous data from coeval basins in California and Oregon. Sharman et al. (2014) presented an extensive compilation of zircon data from 100 to 38 Ma forearc basin strata from southern Oregon to northern Baja California. Figure 6E–G shows composite data from three time slices of the northern Great Valley forearc basin, where detrital zircons were mainly derived from the adjacent northern Sierra Nevada and Klamath block (Ingersoll 1979, 1983; DeGraaff-Surpless et al. 2002). Samples deposited 100–85 Ma are dominated by zircons from the Jurassic arc, with little input from the Cretaceous arc. Over time, the Jurassic arc signal waned and the Cretaceous arc signal increased, such that 71–38 Ma samples are strongly dominated by the Cretaceous arc. Examination of the original data shows that peak ages vary considerably from sample to sample, presumably reflecting the specific plutons sourcing each sample. Few grains are older than 210 Ma, indicating that the Late Cretaceous Great Valley basin was isolated from major input from the erg sandstones, the Nevada backarc region, and basement terranes of the continental interior.

Figure 7. Map of the western USA at about 35–43 Ma, roughly the time we infer for the transition between type D and type A1 zircon age patterns. The map emphasizes relevant source-to-sink sediment transport pathways, including major sediment source areas near arcs and core complexes, depocentres, palaeevallyes, palaeeorivers, and submarine canyons. Post-45 Ma translation of Salinian and Nacimiento blocks along the San Andreas and related faults are restored using the models of Jacobson et al. (2011) and Sharman et al. (2013, 2014). We have not attempted to restore other Cenozoic translations, rotations, and extension that occurred east of the San Andreas. This is due mainly to the lack of data needed to confidently restore post-45 Ma extension and major block rotations in Oregon and Washington and to a lesser extent uncertainties about restoring Basin-and-Range extension farther south. If extensions were restored, the Idaho batholith would be considerably closer to the Tyee and Great Valley depocentres. The Franciscan Coastal Belt may have been translated north a moderate distance after accretion. Based mainly on Dickinson et al. (1979, 1988, 2012), Heller et al. (1985, 1987), Underwood and Bachman (1986), Renne et al. (1990), Wells et al. (2000, 2014), Wyld (2000), Stewart et al. (2003), Reed (2004), Foster et al. (2007), Van Buer et al. (2009), Davis et al. (2010), Lewis et al. (2010), Gaschnig et al. (2010, 2011), Chetel et al. (2011), Jacobson et al. (2011), Henry et al. (2012), Cassel et al. (2012), Long (2012), McCrory and Wilson (2013), Dumitrut et al. (2010, 2013), Sharman et al. (2014), Digital Geology of Idaho (undated), and our data. The basemap and map projection follow Reed (2004).
Figure 6H shows the distinctive data pattern exhibited by mid-Maastrichtian to Palaeocene strata from one specific portion of the unnamed forearc basin of southernmost California (Jacobson et al. 2011; see also Sharman et al. 2014). Key populations are ca. 80–120 Ma zircons from the nearby Cordilleran arc (Sierra Nevada, Mohave region, and Peninsula Ranges batholith), plus zircons from rocks in the southwestern US interior, including Laramide-age magmatic rocks, Yavapai–Mazatzal basement, and Cretaceous grains, with only minor Jurassic and do not match any of our Franciscan samples, except perhaps 7–002 (Point Delgada). This suggests that our samples are older than 71 Ma (discussed later). DeGraaff-Surpless et al. (2002, Figure 12C) concluded that, before 87 Ma, the northern Great Valley was sourced mainly from short rivers that tapped mostly the Jurassic arc rocks on the western flank of the northern Sierra Nevada. Later, the rivers eroded headward and lengthened somewhat to also tap Cretaceous arc rocks that lay farther east and the Cretaceous relative signal strengthened with time (see also Sharman et al. 2014, Figure 10). This interpretation works well for the type C and D samples. Differences between the Franciscan and northern Great Valley patterns are easily attributable to minor differences in specific source areas and depositional ages as the sediment transport networks evolved. The dominance of Jurassic grains in the type C and D samples is incommensurate with sources mainly in the southernmost Sierra Nevada, Mohave region, or Peninsula Ranges batholith, where Jurassic arc rocks lie east of voluminous Cretaceous rocks. Forearc basin sediments sourced from those areas between ca. 100 and 56 Ma contain little 200–135 Ma zircon (Sharman et al. 2014). There also do not appear to be magmatic belts in Mexico that could provide the combination of zircon ages seen in the type C and D samples (e.g. Dickinson and Lawton 2001; Reed 2004).

Type D. This pattern is exhibited by five samples from the Central Belt and one from Point Delgada (Figure 4A and C). Features that distinguish it from type C are that fewer grains have ages characteristic of the Jurassic arc (36–78%), more grains have ages characteristic of the Cretaceous arc (22–64%), and YZP are younger (86 to 100 Ma). Because there are few plutons younger than ca. 87 Ma in the Sierra Nevada to supply zircons, some samples are probably considerably younger than required by their YZP.

Figure 8 compares zircon age distributions in our type C and D samples with distributions from three depositional age intervals in the Great Valley forearc basin. The 98–85 Ma patterns from 10 Great Valley sandstones show moderate variability from sample to sample (Figure 8A). For example, some samples are dominated by ca. 165 Ma Jurassic zircon whereas others by ca. 145 Ma ‘Jurassic’ zircon. This presumably reflects variability in the specific plutons that sourced each sample. Our type C samples are quite similar to some of the northern Great Valley 100–85 Ma samples. Our type D samples are quite similar to 85–71 Ma Great Valley sandstones, with subequal numbers of Jurassic and Cretaceous grains (Figure 8B). Great Valley strata of 71–38 Ma age are strongly dominated by Cretaceous grains, with only minor Jurassic (Figure 8C), and do not match any of our Franciscan samples, except perhaps 7–002 (Point Delgada). This suggests that our samples are older than 71 Ma (discussed later).

New zircon U-Pb results

The 26 samples in Figure 4 show 5 main types of age patterns (Figure 5). Composite plots for these five types are shown in Figure 6J–N. The rock localities for the five types are coded by symbols on the maps in Figures 1 and 2.

Type C. This pattern is exhibited by all three samples from the Yolla Bolly terrane and three of eight samples from the Central Belt (Figure 4B and C). Distinguishing features are that 81–95% of diagnostic grains (Figure 5) have ages characteristic of the Jurassic arc (135–210 Ma), 5–19% of grains have ages characteristic of the older part of the Central Belt (Figure 4B and C). Features that distinguish it from type C are that fewer grains have ages characteristic of the Jurassic arc (36–78%), more grains have ages characteristic of the Cretaceous arc (22–64%), and YZP are younger (86 to 100 Ma). Because there are few plutons younger than ca. 87 Ma in the Sierra Nevada to supply zircons, some samples are probably considerably younger than required by their YZP.

Figure 8 compares zircon age distributions in our type C and D samples with distributions from three depositional age intervals in the Great Valley forearc basin. The 98–85 Ma patterns from 10 Great Valley sandstones show moderate variability from sample to sample (Figure 8A). For example, some samples are dominated by ca. 165 Ma Jurassic zircon whereas others by ca. 145 Ma ‘Jurassic’ zircon. This presumably reflects variability in the specific plutons that sourced each sample. Our type C samples are quite similar to some of the northern Great Valley 100–85 Ma samples. Our type D samples are quite similar to

Type A2. This pattern is exhibited by 2 of 9 samples from the Coastal terrane and is characterized by abundant
IB-age zircons plus ≈8% Challis-age zircons. These samples were probably deposited slightly later than type A1, after Challis magmatism intensified. One sample from the Eocene Great Valley contains a particularly strong Challis peak, indicating that Challis material did indeed reach northern California (Dumitru et al. 2013, locality p).

Figure 8. Comparison of cumulative probability distributions (CPD) of zircon ages from type C and D Franciscan Complex (FC) samples with distributions from Great Valley (GV) samples from three different depositional age intervals. In CPD plots, large numbers of zircons in a short age interval are expressed as high slopes in the line. Age scale changes by a factor of 40× at 300 Ma. Great Valley plots slightly modified from Sharman et al. (2014), based on their data plus data of DeGraaff-Surpless et al. (2002).
**Type B.** This pattern is exhibited by 3 of 10 samples from the Coastal terrane and 1 of 3 from the King Range terrane. It is similar to type D, with the addition of small to moderate populations of 35–32 Ma zircons that were probably derived from the Nevada ignimbrite flare-up (discussed later). Two of the Coastal terrane samples are strongly volcanolithic, with 33–35% volcanic rock fragments and only 12–18% quartz (Table DR6). Such volcanolithic rocks are fairly rare in the terrane. For example, they make up only 3 of 33 total samples from the terrane in Table DR6 and even this may overstate their abundance. We specifically selected these two samples to date because of their unusual compositions. The third type B Coastal terrane sample was collected 50 m from the tuff of Fish Rock Road. This ≈1 m-thick, undated tuff is the only distinct tuff layer yet found in the Coastal terrane (Figure 1; description in Table DR2).

**Zircon fission track samples.** Tagami and Dumitru (1996) reported detrital zircon fission track ages from six samples within the area of Figure 2. These data are much less precise than the new U-Pb ages, because only 7 to 26 grains were dated per sample and uncertainties on single grain ages averaged roughly ±15% (±1σ). The data are good enough, however, to say that the three Coastal terrane fission track samples are either type A1 or A2 and the three Central Belt samples are either type C or D, slightly expanding the areas where those types are documented to occur.

**Depositional ages of Franciscan units in the study area**

The depositional ages of many Franciscan sedimentary units are poorly known, mainly because of the rarity or total absence of megafossils and the limited amount of microfossil work. Zircon data provide helpful constraints, because the age of sedimentation must be equal to or younger than the youngest zircon population that a sample contains. However, there are limitations in using zircon data for this purpose in the Franciscan. Under favourable geologic circumstances, sandstones may contain significant numbers of zircons derived from volcanism essentially contemporaneous with deposition. Such zircons may be transported by air fall from volcanic plumes, or via rivers or turbidity flows. However, data from coeval basins to the east suggest that Cretaceous Franciscan samples should contain little contemporaneous zircon. Upper Jurassic to Campanian sandstones from the Cordilleran foreland basin have yielded virtually no coeval zircons (Dickinson and Gehrels 2008), despite the location of this basin downwind (east) of the Sierra Nevada arc (Figure 7) and despite the basin containing numerous tuffs within its mudstone units (Christiansen et al. 1994). Apparently, the total volume of air fall zircon was small and was overwhelmed by zircons reworked from older strata. Upwind (west) of the arc, Degraaff-Surpless et al. (2002) reported data for 15 samples from the Upper Cretaceous part of the Great Valley forearc basin. In 12 samples, YZP were about 2–19 million years older than depositional ages indicated by fossils, whereas in three they were about 45 million years older. Therefore, in general the YZP in Franciscan samples sourced from the Sierra Nevada should pre-date the true depositional ages, reflecting the time interval required for exhumation and erosion of the youngest zircon-bearing plutons present in their source areas. In some cases, the age difference could be tens of millions of years.

In contrast, the Eocene Tyee Formation in coastal Oregon was partially derived from the active Challis magmatic belt and contains abundant zircon with ages essentially coeval with deposition (Dumitru et al. 2013). Therefore, Franciscan sandstones partially sourced from the Challis complex while the Challis was active would probably contain contemporaneous zircons, but younger Franciscan sandstones sourced after Challis extinction would not.

A further point is that there were intervals of time when plutonism in the Sierra Nevada was subdued (e.g. 140–125 Ma, 85–80 Ma) or extinct (after 80 Ma), so few zircons were generated. So, for example, a 60 Ma Franciscan sandstone sourced solely from the Sierra Nevada would probably have ca. 87 Ma youngest zircons.

**Yolla Bolly terrane**

The Yolla Bolly terrane crops out over a very large area, roughly 6000 km². The continent-derived, elastic rocks of the terrane had long been regarded as having a largely Tithonian to Valanginian depositional age (151–137 Ma), based on about 20 *Buchia* (bivalve) fossil localities (Figure 9; e.g. Blake et al. 1988; Jayko et al. 1989). An *Inoceramus* (bivalve) of Cenomanian (?) age (100–94 Ma) was found at a single locality near Hull Mountain, suggesting deposition there was younger (Blake and Jones 1974; Ohlin et al. 2010). Pelagic radiolarian cherts are mostly late Kimmeridgian to Valanginian (Worrall 1981; Blake et al. 1988; Jayko et al. 1989), but two localities near Clear Lake and Hull Mountain probably range up as young as Aptian–Albian (124–100 Ma; McLaughlin and Ohlin 1984, p. 226; Ohlin et al. 2010, p. 9).

Dumitru et al. (2010) reported ca. 111 Ma YZP in three samples from the classic Yolla Bolly outcrop areas in northern California, an area where all fossil age calls had been late Kimmeridgian to Valanginian. Dumitru (2012) then resampled exactly at three of the *Buchia* sites, determined YZP of 104, 106, and 110 Ma, discovered a *Buchia* inside a conglomerate clast, and argued that essentially all of the extremely rare *Buchia* in the Yolla Bolly terrane are redeposited. In addition, Ernst et al. (2009) found 88 to 105 Ma YZP (recalculated here) in four samples far to the south in the Diablo Range that some workers correlate with the Yolla Bolly terrane (for
review, see Raymond 2015). The current study includes three Yolla Bolly samples from two areas. Two samples from a western outlier of the Yolla Bolly near Zenia have YZP of ca. 108 and 110 Ma. McLaughlin et al. (2000) tentatively identified the Zenia outlier as a fragment of the Taliaferro metamorphic complex, a subunit of the Yolla Bolly notable for a slightly higher metamorphic grade than the remainder of the terrane (Suppe 1973; see also Ernst and McLaughlin 2012). We recently determined a preliminary YZP of ca. 112 Ma from one sample from the main Taliaferro outcrop area (Buck Rock Creek), so this relationship is reasonable. One sample (‘Clear Lake’ in Figure 1) from the Yolla Bolly terrane (Table DR2) near Clear Lake and has a YZP of 118 Ma. McLaughlin and Ohlin (1984) reported pelagic radiolarian chert ages of five samples 4 to 17 km away as Tithonian (n = 1), Tithonian–Valanginian (n = 3), and Aptian–Albian(?) (n = 1). The YZP confirms that Aptian or younger sandstones are present. We are currently working on additional samples from the Yolla Bolly terrane for a future paper.

In summary, the depositional age of the bulk of Yolla Bolly clastic rocks is apparently much younger than traditionally thought. There are several constraints on its age range (Figure 9): (1) near North Yolla Bolly peak, some rocks yield ca. 111 Ma YZP, whereas rocks several kilometres away yield ca. 110 Ma average whole-rock total-gas argon ages, which have been interpreted as dating post-depositional metamorphic recrystallization (Lanphere et al. 1978; Dumitru et al. 2010). This suggests that some Yolla Bolly rocks were deposited ca. 111 Ma in the trench, then immediately subducted and metamorphosed, a sequence of events that should be common in a subduction
zone. (2) Seven other Yolla Bolly samples in northern California yield YZP of ca. 104, 106, 110, 111, 112, 114, and 118 Ma. (3) A Cenomanian (?) fossil (94–100 Ma) was discovered at Hull Mountain and possibly correlated radiolarian chert-bearing units have yielded probable Aptian–Albian radiolarians (124–100 Ma). (4) In a progressive accretion model, structurally higher packets of strata are generally older than the structurally lower packets. In the northern Coast Ranges, the Yolla Bolly terrane structurally underlies the Valentine Spring Formation of the Pickett Peak terrane, which yields ca. 120 Ma YZP and ca. 117 whole-rock total gas $^{40}$Ar/$^{39}$Ar ages (Lanphere et al. 1978; Dumitra et al. 2010). Thus, the Yolla Bolly is likely younger than ca. 120 Ma. (5) For the most part, the Yolla Bolly terrane appears to be older than the underlying Central Belt mélange, which we tentatively interpret to be ca. 103 to 53 Ma (see below). However, limited overlap in Yolla Bolly and Central Belt ages is probably reasonable.

Therefore, our best current estimate for the longest plausible time span for the bulk of the clastic strata in the Yolla Bolly is from 120 Ma (post-Valentine Spring) to ca. 53 Ma (pre-Yager terrane, see below), our best estimate for the shortest plausible time span is from ca. 110 Ma ($^{40}$Ar/$^{39}$Ar ages) to ca. 100 Ma (Inoceramus), and our judgement of the actual age range is ca. 117 Ma (post-Valentine Spring) to ca. 98 Ma (Inoceramus). The 98 Ma younger bracket remains poorly constrained. In this interpretation, Yolla Bolly rocks far to the south in the Diablo Range would represent a partly younger area of the terrane, or perhaps are not actually part of the terrane (see Raymond 2015).

Central Belt

As noted, age relations in the Central Belt are complex. The depositional age of the mud and sand that make up the mélange matrix itself obviously will differ from place to place. Blocks encased in mélange may be exotic or native (derived from shearing of the same strata as the matrix). Exotic blocks and slabs may have depositional ages similar to, older than, or possibly younger than the associated matrix. In many cases, the categorization of a specific rock sample as matrix or block is ambiguous or undocumented, including many fossil-bearing samples or samples dated by zircon U-Pb analysis. Units and contact locations shown on geologic maps may differ, reflecting contrasting interpretations (e.g. Gucwa 1974, 1975; versus Jayko et al. 1989). The block-in-matrix character of some mélanges probably results from mixing of older materials during submarine mass wasting (olistostromes), so some mélanges consist of redeposited material. Finally, parts of the belt have been displaced northward, juxtaposing rock bodies of different ages.

Figure 10 compiles published Central Belt fossil ages in the map area of Figure 2. Only about 30 megafossil localities are known, all from the eastern half of the belt. About 26 are Tithonian to Valanginian Buchia, one is an unspecified Neocomian fossil, and three did not yield useful ages.

Laboratory recovery of microfossils has been more productive. Experience has shown that, by focusing on favourable lithologies such as carbonate concretions, limestone blocks, and radiolarian cherts, roughly 50% of samples may be useful (e.g. Evitt and Pierce 1975; Murchey and Jones 1984; Sliter 1984; Lucas-Clark 1986, 2007; McLaughlin et al. 2000). Nevertheless, the number of published microfossil localities in the area of Figure 2 is only about 50, and many samples (especially mudstones) yield only imprecise age calls (e.g. ‘Cretaceous’). All of these localities are from the western half of the belt and do not overlap with the Buchia localities.

One class of exotic blocks comprises microfossil-bearing pelagic cherts and limestones, which were presumably deposited atop the oceanic Farallon (or Kula) plate, drifted landward with the plate, then accreted into the Central Belt. Radiolarian chert blocks in Figure 10 range from ca. 172 to 94 Ma and a few blocks preserve condensed sections with a protracted range in age. Blocks of the Laytonville foraminiferal limestone preserve an

Figure 10. Published ages of fossils from localities in the Central Belt. Age calls are grouped by subunits within the belt (subunit assignments of some samples are not certain). Age calls within each subunit are ordered by age for simple visual clarity. Most localities are spot samples that must have a single depositional age and lengths of bars show age ranges permitted by the fossils (which are in some cases quite broad, e.g. ‘Cretaceous’). Note however that five pelagic chert and limestone blocks expose stratigraphic sequences with extended age ranges; for these, lengths of bars show full age duration. Fatness and dashing of bars express our impressions of the precision and reliability of the age calls. Also shown are ages of the youngest zircon populations (YZP) of our 8 Central Belt zircon samples, plotted next to the nearest fossil locality, with distance to fossils in kilometres. This figure only includes fossil localities from these map areas in northern California: Covelo, Garberville, Hayfork, and Willows 1:100000 quadrangles; Hull Mountain; Van Arsdale; Geyers chert section. Fossil localities farther south in the San Francisco Bay region are not included. Fossil sources: Alvarez et al. (1980); Berkland (1972, 1973, 1978); Blake and Jones (1974); Clark (1940); Evitt and Pierce (1975); Gucwa (1974, 1975); Hagstrum and Murchey (1993); Irwin et al. (1974); Jayko (1984); Jayko et al. (1989); Jordan (1978); Kaplan (1983, 1984); Koski et al. (1993); Larue (1986); Lehman (1974); Lucas-Clark (1986, 2007); Mankinen et al. (1991); McLaughlin and Pessagno (1978); McLaughlin and Ohlin (1984); McLaughlin et al. (2000); O’Day (1974); Murchey and Blake (1993); Ohlin et al. (2010); Orchard (1979); Sliter (1984); Suppe (1969, 1973); Tarduno et al. (1986, 1990). Note that the extensive fossil compilation table in Jayko et al. (1989) contains numerous errors, which we have corrected from original sources as shown in Table DR7. The 2009 GSA geologic time scale is used (Walker and Geissman 2009).
≈50 m-thick stratigraphic succession that ranges from late Albian (104 Ma) to early Coniacian (88 Ma) and was deposited at 14 ± 5° south palaeolatitude, followed by remarkably rapid northward drift and eventual accretion into the Central Belt (Alvarez et al. 1980; Debiche et al. 1987; Tarduno et al. 1990). Southeast of our study area near Middle Mountain, Bathanitic(?) through Tithonian (?) Van Arsdale terrane basalts and overlying Maastrichtian (ca. 71–65 Ma) basalt-clast conglomerate, calcadonitic sandstone, and shale apparently represent a seamount and its capping sediments accreted into the Central Belt during Palaeocene time (Berkland 1972, 1978; Murchey and Blake 1993; see also Mankinen et al. 1991). The Van Arsdale rocks especially indicate that accretion in the Central Belt continued at least until the Palaeocene.

Figure 10 also includes microfossil ages from dinoflagellate cysts, spores, and pollen preserved in clastic rocks sourced from the continental margin. Detailed work on carbonate nodules and lenses at two localities recovered well-preserved dinoflagellates that strongly indicate a late Albian age (104–100 Ma; Gucwa 1974, 1975; Lucas-Clark 1986, 2007). Of the remaining 30 localities, 26 permit wider age ranges within the Jurassic–Cretaceous, such as ‘Late Jurassic–Early Cretaceous’, ‘Cretaceous’, or ‘Albian–Cenomanian’. It may be significant that none of these samples actually require ages older than late Albian. These include samples from mélangé areas (presumably mostly from matrix but some probably from blocks) and from four slabs(?) of more coherent strata as shown in the figure. Cenozoic age calls are limited to two localities from the westernmost edge of the Central Belt that are Palaeocene, Eocene or, less likely, Oligocene, plus two localities from a slab(?) along Blue Rock Creek (Figure 2), which are probably Eocene–Palaeocene (it is alternately possible that some of these four localities are actually from the Coastal Belt, see Table DR1). These data are limited, but indicate that Albian rocks are present and possibly abundant, and that some rocks towards the west are probably early Cenozoic in depositional age. In southwestern Oregon, exhumed rocks of Central Belt affinity are overlapped by little-deformed strata that are probably ca. 54 Ma (early Eocene; Blake et al. 1985a; age updated from Wells et al. 2000, Figure 4), suggesting that Central Belt deposition had ended before 54 Ma in Oregon.

Figure 10 also shows one sample of an impure cherty hemipelagic lens interbedded with argillite that probably represents mélange matrix and that yielded Hauterivian–Barremian radiolarians (136–125 Ma) (McLaughlin et al. 2000). In the San Francisco Bay region, Blake et al. (2000, p. 5) highlighted rocks at a quarry in Greenbrae as a rare preserved slab of interbedded sandstone, mudstone, chert, and tuffaceous volcanic rocks that probably represent the original sedimentary accumulation that was subsequently sheared to form the mélange matrix. The cherts yielded a Late Jurassic (probably Tithonian) radiolarian assemblage (locality 47 of Murchey and Jones 1984). Also, in the Bay region, the relatively coherent sandstone-mudstone Novato Quarry terrane of the Central Belt has yielded rare early Campanian fossils (ca. 82–79 Ma; Blake et al. 1984, 2000; Elder and Miller 1993). Snow et al. (2010) presented detrital zircon U-Pb data from the Bay region.

Our eight Central Belt zircon samples are all from the western half of the belt and yield youngest zircon populations as follows:

- CR-17 (matrix) 86 Ma;
- CR-30 (broken formation) 90 Ma;
- CR-65 (Cummings Cr. slab) 90 Ma;
- CR-10 (matrix) 90 Ma;
- CR-81 (broken formation) 95 Ma(?);
- CR-19 (Poonkinney slab) 100 Ma(?);
- Ukiah (Lookout Peak slab) 106 Ma (plus 96 Ma TIMS grain);
- Geysers (sandstone) 113 Ma (but overlies 95 Ma chert).

All samples have overall distributions indicative of Sierra Nevada sources. The first four samples have YZP of 86 to 90 Ma, with roughly 15% of grains with ages of 90 Ma within uncertainty. These samples are no older than ca. 86–90 Ma, but are likely younger, given that little of the Sierra Nevada batholith is younger than ca. 87 Ma. On the other hand they are probably not younger than ca. 71 Ma, because they have much stronger Jurassic components than seen in post-71 Ma strata from the Great Valley (Figure 6G and 8C). The Geysers sample is a reconnaissance sample dated by G.E. Gehrels from outside our study area and it is beyond the scope of this article to discuss the intriguing and complex relations in that area (Table DR2; McLaughlin and Pessagno 1978; McLaughlin 1978; McLaughlin and Ohlin 1984; McLaughlin et al. 1988; Hagstrum and Murchey 1993).

The age data from the clastic rocks of the Central Belt in our study area present a perplexing picture, with the megafossils, microfossils, and zircon data each suggesting different predominant ages. The megafossils yield roughly half Tithonian and half Valanginian ages, but nothing younger. The microfossils yield virtually no Tithonian ages, allow but do not require Valanginian ages, perhaps tend to cluster in the Albian, actually require no ages older than late Albian (except for the cherty hemipelagic lens), give only weak indications of Late Cretaceous ages from mélange samples, and include only a few early Cenozoic samples. All of the zircon samples suggest ca. 100–71 Ma depositional ages.
The contrasts between the eastern and western parts of the belt may be important. Aside from the contrasts in fossil types and ages, the eastern part contains relatively sparse high-grade blocks, whereas much of the western part is noted for numerous blocks (Bailey et al. 1964, Figure 18; Coleman and Lanphere 1971; Cloos 1986, Figure 1). Perhaps the western part is younger than the eastern and records a period of more intense incorporation of high-grade blocks into mélange units?

Several lines of evidence suggest that many or all of the Buchia megafossils might possibly be redeposited. At one locality, Jordan (1978, p. 54) found Buchia only inside limestone cobbles within a conglomerate body within a mélange unit, unambiguous evidence for redeposition (Figure 10). Irwin et al. (1974) reported two localities (PP3, PP12) with both Tithonian (151–145 Ma) and Valanginian (141–136 Ma) Buchia species, which, if reliable, suggests redeposition of both species during post-Valanginian time. Almost all of the published Buchia were found in four areas where intensive, ≈1:63000-scale mapping shows mélange with abundant blocks of volcanic rock, sandstone, and chert. Three of these areas also expose sparse high-grade blocks (Irwin et al. 1974; Lehman 1974; Jordan 1978), whereas one lacks such blocks (Suppe 1973). If these mixed rocks are olistostromal, redeposition of fossils would be plausible. None of the microfossil or zircon age calls lend support to the megafossil ages, although they are all from the opposite side of the belt.

Based on these data, our tentative estimate for the longest plausible time span for the bulk of the clastic sedimentary rocks in the Central Belt in northern California is from 151 Ma (Tithonian Buchia) to 53 Ma (pre-Yager). Our estimate for the shortest plausible time span is from 102 Ma (late Albian microfossils) to 65 Ma (Van Arsdale seamount). As a tentative working hypothesis, we suggest that the true age range might be ca. 103 to 53 Ma. For the 103 Ma older age bracket, this presumes that all the Buchia are redeposited and allows the older age limit of the Central Belt clastic rocks to overlap only slightly with the younger age limit of the Yolla Bolly terrane, as permissible in a simplistic application of a slightly time-transgressive progressive accretion model. For the younger age limit, this presumes that Central Belt deposition continued until superseded by deposition in the Coastal Belt. However, there is little clear evidence for Central Belt deposition after 71Ma and the period from 71Ma until 53 Ma could have been a period of nonaccretion or only minor accretion. About 75% of modern subduction zones are in a state of nonaccretion or subduction erosion, in which virtually all trench sediments are subducted towards the mantle, so periods of nonaccretion in the Franciscan should be expected (e.g. Clift and Vannucci 2004; Scholl and Von Huene 2007; Dumitru et al. 2010). The Van Arsdale seamount apparently arrived within this time interval, but partial decapitation and accretion of a tall seamount during a period of nonaccretion of a relatively thin section of trench sediments are predicted in some models of accretionary processes (e.g. Cloos 1993).

Clearly more work is needed on the Central Belt and we are continuing our zircon studies. It also appears that genuine potential exists for future dinoflagellate work. In two areas, Lucas-Clark (1986, 2007) was able to determine late Albian ages within a tight 4-million-year range, using samples from lenses of carbonate nodules and limestone, where preservation is generally better. Larue (1986) reported relatively imprecise age calls (e.g. ‘Cretaceous’) determined by Lucas-Clark from 12 less optimal mudstone samples originally collected for petrology and geochemistry. This suggests that at least marginal preservation of dinoflagellates is common in mudstones, despite petromorphism of the belt, so careful sampling of carbonate nodules, concretions, and lenses might well be fruitful. Very little California dinoflagellate work has been published since the 1980s, when age calibrations were less sophisticated. Table DR8 includes details of Larue’s (1986) work, in hopes of inspiring future efforts.

Yager terrane

Depositional ages for the Yager terrane have been based entirely on dinoflagellates, pollen, and spores, inasmuch as the terrane has proven to be barren of diagnostic megafossils, foraminifera, and radiolarians (O’Day 1974; Evitt and Pierce 1975; Bachman 1978, 1979; Damassa 1979a, 1979b; Frederiksen 1989; McLaughlin et al. 1994, 2000). Dumitru et al. (2013, Table DR1) reviewed the data in detail and concluded that the microfossil data indicate that the Yager terrane contains rocks of Eocene and possibly also Palaeocene and Late Cretaceous ages. Reworked microfossils are fairly common in the terrane (Evitt and Pierce 1975; Frederiksen 1989), so the pre-Eocene ages could be spurious. Our three Yager samples are all type A1 with YZP of ca. 53, 53, and 58; younger Challis-age grains are absent. The Tyee Formation in Oregon has a depositional age range of 49.3–46.5 Ma and contains abundant zircon of both Challis and Idaho batholith age, indicating that the Challis complex was a fertile source for zircon at 49.3–46.5 Ma (Figure 6D). Our preferred interpretation is that our Yager samples were deposited before Challis magmatism became voluminous. If so, these samples are early Eocene, possibly extending into the earliest middle Eocene (ca. 48 Ma). A less likely alternative is that these samples were sourced from an area with few Challis rocks, in which case the samples are early Eocene and/or younger (<58 Ma).

Therefore, our estimate for the longest plausible depositional time span for the bulk of the clastic rocks in the Yager terrane is from 58 Ma (YZP) to 48 Ma
(pre-Challis), our estimate for the shortest plausible time span is ca. 52 to 50 Ma, and our judgement of the actual age range is the latter from 52 Ma to 50 Ma. The very rapid deposition of the Yager probably reflects prodigious supply of sediments from the 53–40 Ma extensional terranes in Idaho (Foster et al. 2007; Dumitru et al. 2013).

Coastal terrane

Dumitru et al. (2013, Table DR1) compiled the palaeontological data from the Coastal terrane. McLaughlin et al. (1994, 2000) concluded that the collective fossil evidence indicates that the elastic strata of the northwestern quarter of the Coastal terrane (north of about latitude 40° N) are dominantly middle to late Eocene in age. Some strata of early Eocene and/or Palaeocene age crop out farther south. Several Late Cretaceous palynomorph ages have also been reported (A. Traverse, unpublished; compiled by Bachman 1979, p. 95), but may be unreliable (see Kleist 1975) or possibly reworked. Evitt and Pierce (1975) and McLaughlin et al. (1994, p. 23 and 25) noted reworking of older microfossils in some samples.

Our four type A1 samples from the Coastal terrane each yields 52–55 Ma YZP and lacks Challis zircons. We interpret their depositional ages as ca. 52–50 Ma (early Eocene). Both type A2 Coastal terrane samples yield abundant IB-age plus roughly 8% Challis-age zircons, with YZP of ca. 49 Ma. Their depositional ages are probably 49 Ma. Our three type B Coastal terrane samples have YZP of ca. 35 to 32 Ma, setting their maximum depositional ages and suggesting that some earliest Oligocene or younger rocks are present. One of these was collected near the tuff of Fish Rock Road (Table DR2 and Figure 1) and is arkosic (Table DR6). The other two are highly volcanolithic, with 33–35% volcanic rock fragments and only 12–18% quartz. As noted previously, such high Lv sandstones are rare in the terrane and are overrepresented in our data set because we specifically targeted them for analysis. Kleist (1974), Underwood and Bachman (1986), and Langenheim et al. (2013) described distinctive high-Lv sandstones within the Coastal terrane, but this petrofacies did not form discrete, mappable bodies.

Therefore, our estimate for the longest plausible time span for the bulk of the elastic strata in the Coastal terrane is from 55 Ma (YZP; pre-Challis) to post-32 Ma (YZP), our estimate for the shortest plausible time span is from 52 to 32 Ma, and our judgement of the actual age range is from 52 Ma to post-32 Ma. Following Underwood and Bachman (1986), we consider it reasonable that the Yager and Coastal terranes partially overlap in age, with some Coastal terrane strata possibly representing deposition in outboard, bathyal-abyssal settings in or near the trench, and with coeval Yager strata deposited farther up-slope in middle to inner bathyal settings.

King Peak subterrane of the King Range terrane

Samples 77 and 75 from the King Peak subterrane yielded contrasting data patterns. Sample 77 yielded abundant SNB-age zircons, 4 grains of IB-age zircon, and 2 grains of young (32 Ma) zircon. Sample 75 is strongly dominated by IB-age zircon with no younger grains. McLaughlin et al. (1982, 2000) recovered middle Miocene radiolarians (16–12 Ma) on two traverses nearby, one from 1 to 1.5 km south of sample 77 and one ≈2 km SSW of sample 75. The few young zircons suggest a 32 Ma depositional age, but easily allow a middle Miocene age and that is our preferred interpretation. The limited zircon data are insufficient to permit refinement of that age. The entire King Range terrane was overprinted by a hydrothermal event dated at 13.8 Ma (McLaughlin et al. 1985), so the best estimate of the depositional age is 16–14 Ma. Other, unsampled parts of the King Peak subterrane could include older strata.

The strong Idaho batholith signal in sample 75 is surprising for a rock purportedly deposited at 16–14 Ma. We speculate that sample 75 may have been sourced from older Coastal Belt strata, rather than from Idaho. McLaughlin et al. (1982) found redeposited early Cenozoic foraminifera mixed with middle Miocene foraminifera and radiolarians, suggesting partial redeposition of older sediments.

Point Delgada subterrane of the King Range terrane

The Point Delgada subterrane includes pillow basalt interbedded with minor dark red argillite that contains Campanian to Coniacian (71–89 Ma) radiolarians (McLaughlin et al. 1982, 2000). The basalts are overlain by a thin, undated argillite-matrix mélange, which in turn is structurally overlain by undated turbiditic arkosic metasandstone and argillite that is locally interbedded with basalt flows and flow breccia (McLaughlin et al. 1985). Sample 7-002 was collected in 1989 for fission track studies (Dumitru 1991), presumably from the metasandstone unit (McLaughlin et al. 1985, Figure 2). It contains SNB-age zircon with a YZP of 86 Ma, plus a single grain of 49 Ma zircon with well-behaved U, Th, and Pb isotopic patterns. It is unclear whether the 49 Ma grain is reliable. The sample was collected from a beach exposure and wave action might have forced traces of contaminating modern beach sand into the fractures that are abundant in most Franciscan rocks. Therefore, we do not attempt to interpret the depositional age of this sample beyond the observation that its YZP is either 86 or 49 Ma.

The Point Delgada subterrane might appear to violate a progressive accretion model for the Franciscan, in that it lies outboard of the younger Coastal terrane. However, McLaughlin et al. (1982, 1994, 2000) argued that the King Range terrane was translated northward parallel to
the plate margin and then accreted during post-middle Miocene time. Moreover, the fault associated with the 1906 San Francisco earthquake surface rupture partly bounds the northeast side of the Point Delgada subterrane. This implies a more complex variant of a progressive accretion model that involves transitioning from subduction to transform faulting.

Sources of Franciscan clastic sediments in the study area

Type C and D samples from the Yolla Bolly terrane and Central Belt

Our 11 type C and D samples from the Yolla Bolly terrane and the Central Belt appear to be highly compatible with sources in the Sierra Nevada and Klamath Mountains, particularly the northern SNB where abundant Jurassic arc rocks crop out west of Cretaceous arc rocks. As noted early, the abundance of Jurassic grains in these samples is incompatible with sources in the southernmost Sierra Nevada, Mohave region, or Peninsula Ranges batholith, where Jurassic arc rocks lie east of voluminous Cretaceous rocks (Figure 7). Forearc basin sediments sourced from those southern areas between ca. 100 and 56 Ma contain little 200–135 Ma zircon (Sharman et al. 2014). There also do not appear to be magmatic belts in Mexico capable of providing the combination of zircon ages seen in type C and D samples (e.g. Dickinson and Lawton 2001; Reed 2004).

Type A1 and A2 samples from the Yager and Coastal terranes

The type A1 samples from the Yager and Coastal terranes (n = 7) were apparently derived from a mixture of IB and SNB sources. They contain the 53–98 Ma zircons characteristic of the Idaho batholith and the 140–180 Ma zircons that are abundant in the Sierra Nevada and Klamaths.

Based on four of these type A1 samples, Dumitru et al. (2013) argued that much of the Yager and Coastal terranes were sourced from the Idaho batholith region at about the time of initiation of major regional extension in the core complexes in that area (Foster et al. 2007). That extensional tectonism also shed clastic sediment to the Tyee forearc basin in Oregon, part of the Great Valley forearc basin in California, and an interval within the Green River Basin. Two-mica granitoids in Idaho apparently sourced relatively abundant detrital muscovite in some of the Yager, Coastal terrane, Tyee, and Great Valley strata (Heller et al. 1985; Dumitru et al. 2013).

The two type A2 samples from the Coastal terrane were apparently derived chiefly from the Idaho batholith and the Challis complex and represent continued input from that region until at least ca. 49 Ma.

Dumitru et al. (2013) reconstructed the sediment transport pathway from extensional areas in Idaho to the Franciscan trench. Detritus apparently left Idaho via a hypothetical Eocene Princeton River which may have transited the poorly known Montgomery Creek basin (Renne et al. 1990) then likely debouched into the northern end of the Great Valley near the head of the well-known Princeton submarine canyon (Redwine 1972, 1984), which now lies mostly buried beneath the northern Great Valley (Figure 7). Detritus then coursed down the canyon to reach the Delta depocentre in the central part of the Great Valley and then supplied the outboard Franciscan trench. A key piece of evidence for this pathway is the presence of significant detrital muscovite in Capay Formation strata that backfilled the canyon. The relative amounts of Idaho and Sierra Nevada zircon reaching the Franciscan probably fluctuated over time, tracking fluctuations in erosion rates in the two source areas (cf. Dickinson et al. 2012). In the Great Valley forearc basin, Markley Formation sandstones as young as 37 Ma contain significant detrital muscovite, suggesting that a continuous Princeton River–Princeton Canyon system persisted until about that time. However, it probably delivered less sediment later in time, because much of the sediments sourced from Idaho apparently were diverted via a newly formed Tyee River to the new Tyee forearc basin in coastal Oregon, where the 49.3–46.5 Ma Tyee Formation records very rapid sedimentation. Many Eocene marine sandstones from the northern Great Valley contain no Idaho batholith-age zircon (Sharman et al. 2014). Such strata were probably deposited upslope of the Princeton Canyon and hence received no IB detritus.

Type B samples from the Coastal terrane

One of the three type B samples from the Coastal terrane was collected near the tuff of Fish Rock Road (Figure 1; R. J. McLaughlin, unpublished data). Based on the light colour of the tuff, the arkosic composition of our sandstone sample, and the 32 Ma YZP of the sandstone, the tuff was likely sourced from intense eruptions in Nevada at roughly 34 Ma (Henry and John 2013). Zircon from various Nevada tuffs was then likely mixed with Sierra Nevada dominated detritus to constitute our sample.

Two of the type B samples have YZP of 35 and 32 Ma and are very rich in volcanic rock fragments (high-Lv, Table 1). They contain abundant SNB age zircons but few or no IB zircons, suggesting primary zircon sources in the Sierra Nevada and the demise of fluvial transport pathways from the Idaho batholith. Curiously, they do contain about 10% Challis-age zircons. We speculate that 51 to 43 Ma-age Challis zircons may have been
transported by volcanic ash plumes and deposited in uplands in the Sierra Nevada, Nevada, and/or the Klamaths, then redeposited after 35 Ma.

The volcanic rock fragments and young zircons in these two samples could have been sourced from either the basaltic to andesitic Oligocene Cascade arc (Du Bray and John 2011; see also Colgan et al. 2011) or from the felsic Nevada ignimbrite flare-up (Henry and John 2013). Petrographic examination of our five Coastal terrane samples with the highest-Lv indicates that their volcanic rock fragments are basalts to andesites (Table 1), hence likely sourced from the Cascades. Langenheim et al. (2013, p. 5) described the magnetic and petrographic properties of a few Coastal terrane high-Lv sandstone samples and inferred that they were arc-derived. Underwood and Bachman (1986) previously defined a high-Lv petrofacies in the Coastal Belt (15 samples) and noted that volcanic rock clasts were chiefly andesitic with lathwork textures (see also Kleist 1974). We infer that most of these high-Lv sandstones were sourced from the Oligocene Cascades (although some high-Lv sandstones could be Eocene and sourced from Jurassic-Cretaceous arc rocks). However, we infer that the young zircons were instead sourced from Nevada ignimbrites, because they match the age of particularly vigorous volcanism there, because the tuff of Fish Rock Road suggests that some ignimbrite material did reach the Coastal Belt, and because basaltic to andesitic rocks commonly contain little zircon.

**Samples from the King Range terrane**

The type D sample from the Point Delgada subterrane was apparently derived from the Sierra Nevada. It also contains about 25% old grains with age peaks of ca. 1300 and 1600–1750 Ma. Numerous possible sources exist for the older grains (Figures 6A–H and 8B; Surpless 2015, p. 755–757).

The type A1 sample from the King Range terrane is surprising, inasmuch as it suggests that a sediment transport pathway from Idaho continued to reach California during Miocene time. We speculated that this sample may instead have been sourced from older Coastal Belt sandstones with type A1 zircon distributions.

The type B sample from the King Peak subterrane contains SNB, IB, and two 32-Ma zircon grains, but lacks Challis zircons. This sample was probably derived mainly from the Sierra Nevada. We again speculate that the IB-age zircons could have been redeposited from older Coastal Belt strata.

**Discussion and conclusions**

Lack of good age control has been a serious hindrance in understanding the Franciscan Complex. The new data presented here suggest that the bulk of the Yolla Bolly terrane, much of the Central Belt, and limited parts of the Coastal Belt are younger than traditionally thought. Figure 9 presents our current best estimates of the times and durations of sedimentation of the various Franciscan units studied in this article.

The tectonic histories of the Yolla Bolly terrane and the Central Belt have been somewhat perplexing, because their depositional lifespans seemed to be surprisingly long. The Yolla Bolly seemed to represent some form of marine basin that received clastic sediments from a continental margin for perhaps 60 million years, from ca. 151 Ma until it was subducted at perhaps ca. 92 Ma (e.g.

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**Table 1.** Average modal percentages of phases inside of trachytic volcanic clasts.

| Coastal terrane sample no. | CR-50  | CR-52  | CR-73  | 2F-519 | 2F-525 |
|---------------------------|--------|--------|--------|--------|--------|
| Quartz:plagioclase:Lv (trachytic rock fragments) of bulk rock (%) | 6:22:35 | 12:18:34 | 21:22:21 | 31:27:21 | 18:22:33 |
| Youngest zircon population (Ma) | Not dated | 35 | Not dated | 52 | 33 |
| Zircon age pattern | – | Type B | – | Type A1 | Type B |
| Number of volcanic clasts examined | 7 | 8 | 7 | 7 | 8 |
| Mineral percentages inside trachytic clasts (%) | | | | | |
| Quartz | 1 | 1 | Tr | Tr | Tr |
| Altered plagioclase | 48 | 48 | 59 | 55 | 59 |
| Opaques | 2 | 1 | 1 | 2 | 2 |
| Titanite | Tr | 1 | 2 | 1 | 1 |
| Limonite | 3 | Tr | Tr | Tr | Tr |
| Layer silicates (incl clays + calcite) | 44 | 48 | 37 | 41 | 37 |
| Altered clinopyroxene | 1 | 1 | 1 | Tr | 1 |

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Notes: These five samples are anomalously rich in volcanic rock fragments, which are highly altered. See Table DR6 for full modal compositional data. These data were collected to determine the original compositions of the volcanic clasts. The data indicate that the highly-altered volcanic clasts in these samples are compatible with derivation mainly from basalts, basaltic andesites, and andesites such as those of the Oligocene Cascade arc, but incompatible with derivation from silicic volcanic rocks of the Nevada ignimbrite flare-up. Note, however, that we do interpret 2F-519 as Eocene rather than Oligocene and it does have a similar volcanic clast composition, although its bulk composition contains considerably more quartz and fewer volcanic rock fragments than the other samples. The source of its volcanic clasts is unclear, but may be maﬁc-to-intermediate volcanic rocks of the Sierra Nevada arc.
Blake and Jones 1974, p. 353, 1981, p. 328; Mattinson and Echeverria 1980, p. 592; Blake et al. 1985b, p. 169; Isozaki and Blake 1994, p. 294; Dumitru et al. 2010, Paragraph 74). If so, it somehow eluded the subduction, metamorphism, and intense deformation inflicted on the now overlying Pickett Peak terrane at roughly 123–117 Ma (Lanphere et al. 1978; Dumitru et al. 2010). Some parts of the Central Belt seemingly accumulated clastic sediments for even longer, from ca. 151 Ma until they were subducted as late as the Palaeogene, meanwhile eluding both Pickett Peak and Yolla Bolly subduction, metamorphism, and deformation. This seemed to suggest that the Yolla Bolly terrane and parts of the Central Belt were allochthonous basins transported intact to the Franciscan trench, then accreted. However, this model has never really been elaborated on in detail (see the above references), probably reflecting the paucity of information and the difficulties of making it work. A simple partial solution, that the Yolla Bolly terrane and Central Belt are composite units comprised of multiple bodies of rock that were deposited, subducted, and accreted at much different times, has not really been articulated and seems contrary to the concept of the Yolla Bolly as a distinct terrane.

It now appears that the lifespan of the Yolla Bolly terrane is actually considerably shorter than formerly thought, ameliorating although certainly not fully resolving these issues. Ages of very rare Buchia had suggested that the Yolla Bolly terrane was mainly Tithonian–Valanginian (e.g. Blake et al. 1988). However, zircon data from three Buchia localities in northern California and one in the Diablo Range yield much younger YZP, indicating that those Buchia are redeposited (Dumitru 2012). There are also several reports from the Yolla Bolly terrane and Central Belt of Buchia found inside clasts in conglomerates (Suppe 1973, Table 13; Jordan 1978, p. 54; Dumitru 2012), loose in conglomerates (Ghent 1963; see also Blake et al. 1984, p. 9), or loose in pebbly mudstones (Blake and Jones 1974, p. 346), strong evidence of redeposition. We suspect that all Buchia in the Yolla Bolly terrane are probably redeposited. Although uncertainties remain, most of the clastic rocks of the Yolla Bolly appear to be younger than the age of Pickett Peak accretion, as expected with a progressive accretion model. In such a model, most of the clastic strata in the Yolla Bolly terrane were deposited in or near the trench in submarine fans or related environments, then subducted and accreted shortly thereafter. Different packets of rocks were deposited and accreted at different times, from about 118 Ma until about 98 Ma. Some packets of rocks may have experienced more complex histories, such as deposition on the trench slope, before being redeposited. Later faulting has complicated stratigraphic and structural relations in many areas. Many oceanic basalts and pelagic cherts are considerably older and drifted to the trench atop the subducting oceanic plate, before detaching and being juxtaposed with the arc-derived clastic rocks.

Various interpretations have been applied to the Central Belt. Many workers attribute mélange formation to mixing during submarine mass wasting (olistostromes), which would interlayer mélange deposits with bodies of more coherent sediments (e.g. Gucwa 1975; Phipps 1984; Wakabayashi 2011, 2012, 2015; Aalto 2014; Platt 2015). Cloos (1982) argued that subduction-driven tectonic return flow of mélange was a key process, subducting materials to depths of 6 to >25 km, mixing and shearing various types of blocks with the matrix, and returning the mélange to shallower depths (see also Cloos and Shreve 1988a, 1988b; Ukar et al. 2012). In the San Francisco Bay region, Blake et al. (1984) speculated that the Central terrane was a more distal volcanic-and chert-rich facies of the Great Valley forearc basin into which other Franciscan terranes were tectonically emplaced, including those comprised of fragments of oceanic crust, oceanic islands or plateaus, and continental margins. McLaughlin and Ohlin (1984) argued that the Central Belt represents a pervasively sheared zone of right-lateral translational tectonism between the oceanic plate and the North American margin, into which allochthonous oceanic terranes docked. Based on faunal and palaeomagnetic data, Hagstrum and Murckey (1993) inferred that pelagic chert sequences at Marin Headlands and The Geysers were deposited at near-equatorial latitudes, then accreted into the Franciscan at low latitudes (e.g. 11° N), and that parts of the Franciscan were subsequently translated north >4000 km along strike-slip faults.

Macrofossil and microfossil data suggest that the bulk of the clastic rocks in the Central Belt range in age from Tithonian to Eocene, perhaps with clusterings in the Tithonian and Albian. Our eight new zircon samples can make only a limited contribution to unravelling the complexities of the belt. All were probably deposited between ca. 100 and 71 Ma (although some samples are less certain), suggesting that Late Cretaceous sediments may be more common than suggested by the microfossil data. It appears plausible but hardly certain that many or perhaps all the Buchia in the Central Belt matrix are redeposited and that the matrix might contain little pre-Albian clastic rock. There appear to be few Franciscan rocks with depositional ages between 71 and 53 Ma, so this might be a period of nonaccretion or only sparse accretion. It is beyond the scope of this article to go further into interpreting the Central Belt.

Beginning at 53 Ma, the general Idaho batholith region became a major source for Franciscan sediments (Dumitru et al. 2013). This apparently reflects the initiation of major extension in the Bitterroot, Anaconda,
Clearwater, and Priest River metamorphic core complexes (53–40 Ma; Foster et al. 2007) and major volcanism in the Challis volcanic field (51–43 Ma; Gaschnig et al. 2010, 2011). This extensional tectonics apparently deformed and uplifted a broad region, shedding voluminous sediments towards depocentres in the Franciscan Coastal Belt, the Great Valley forearc basin, and the Tyee forearc basin (Heller et al. 1985), as well as an interval within the greater Green River lake basin (Chetel et al. 2011). These sediments reached the Great Valley and the Franciscan trench via the Princeton River and the Princeton submarine canyon (Figure 7). In modern subduction zones, the highly variable volume of sediments supplied to the trench appears to be the strongest control on rates of sediment accretion and accretionary prism growth (e.g. Cloos and Shreve 1988a, 1988b; Clift and Vannucchi 2004; Scholl and Von Huene 2007; see also Dumitru et al. 2010). The Coastal Belt crops out over ~3500 km², ~15% of the total outcrop area of the Franciscan Complex, and much of it apparently reflects a relatively brief period of unusually voluminous sediment input from the major core complex extensional episode in Idaho. As such, it illustrates the importance of considering sediment transport pathways and sedimentation rates in reconstructing the growth of ancient accretionary prisms. The Coastal Belt may be an instructive example of a major accretion event driven by continental tectonics far within the interior of the overriding plate. On the other hand, various speculative models suggest that Eocene extension and magmatism in Idaho may have been caused by Eocene subduction-zone plate boundary events off the coast of Oregon, Washington, and British Columbia, such as (1) accretion of the Eocene Siletzia oceanic plateau (Figure 7; McCrory and Wilson 2013; Wells et al. 2014), (2) slab-window tectonics under Idaho in response to Siletzia accretion and breakoff of the subducting Farallon slab (Schmandt and Humphreys 2011), or (3) a wider-scale plate reorganization affecting areas from Oregon and Idaho on northward to Alaska (Foster and Vogl 2011). If any of these models are correct, Coastal Belt accretion was driven indirectly by plate boundary changes farther north, via tectonic events in the continental interior. Still, sediment input rate remains a key parameter in prism growth.

Many workers have inferred that almost all Franciscan clastic sediments were derived from the Sierra Nevada and adjacent regions of the North American continent (e.g. Dickinson et al. 1982; Seiders 1983; Ghatak et al. 2013). However, others have speculated that some Franciscan clastics, particularly sandstones with anomalous compositions (e.g. on quartz-feldspar-lithics ternary plots), could have been sourced and deposited elsewhere and then transported tectonically to their current locations (e.g. Blake and Jones 1981; Blake et al. 1984; Jayko and Blake 1984). The sandstones most likely to be far-travelled would be basal sandstones deposited atop exotic oceanic basalts, pelagic cherts, or pelagic limestones when they were first accreted to North America, possibly in southern California or Mexico (e.g. Hagstrum and Murchey 1993). We have not attempted to sample such sandstones (however, see G.E. Gehrels’ ‘Geysers’ sample in Table DR2, which is such a sample). For the ca. 118 Ma and younger Franciscan units we have studied here, the data are consistent with sources in the Sierra Nevada and Klamath Mountains and in Idaho. The type C and D samples represent a good match with Jurassic arc rocks exposed on the western flank of the northern Sierra Nevada and in the Klamaths, plus a lesser contribution from Cretaceous arc rocks exposed both farther east and south. The Idaho sediments were apparently transported across a gap through the extinct Sierra Nevada–Klamath arc by the Eocene Princeton River, and so also tie the Franciscan to northern California (Dumitru et al. 2013). This conclusion is consistent with only limited post-depositional displacement of these particular Franciscan units.

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Supplemental data

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