The Catalina Schist: Evidence for Mid-Cretaceous Subduction Erosion of Southwestern North America

M. Grove\textsuperscript{1} \hspace{1cm} Department of Earth & Space Sciences, University of California, Los Angeles, 3806 Geology, Los Angeles, CA 90095 USA
Tel. (310) 794-5457 FAX: (310) 825-2779 marty@oro.ess.ucla.edu

G.E. Bebout \hspace{1cm} Department of Earth & Environmental Sciences, Lehigh University, 31 Williams Dr, Bethlehem, PA 18015 USA

C.E. Jacobson \hspace{1cm} Department of Geological and Atmospheric Sciences, 253 Science I, Iowa State University, Ames, IA 50011-3212 USA

A.P. Barth \hspace{1cm} Department of Earth Sciences, Indiana/Purdue University, 723 West Michigan Street, Indianapolis, Indiana 46202 USA

D.L. Kimbrough \hspace{1cm} Department of Geological Sciences, San Diego State University, 5500 Campanile Drive, San Diego CA 92182-1020 USA

R.L. King \hspace{1cm} School of Earth and Environmental Sciences, Washington State University, Pullman, WA 99164-2812, USA

Haibo Zou \hspace{1cm} Department of Earth & Space Sciences, University of California, Los Angeles, 3806 Geology, Los Angeles, CA 90095 USA

O.M. Lovera \hspace{1cm} Department of Earth & Space Sciences, University of California, Los Angeles, 3806 Geology, Los Angeles, CA 90095 USA

B.J. Mahoney \hspace{1cm} Department of Geology, University of Wisconsin-Eau Claire, Phillips 157, Eau Claire, WI 54702-4004 USA

G.E. Gehrels \hspace{1cm} Department of Geosciences, University of Arizona, Tucson, Gould-Simpson Bldg. 529, AZ 85721 USA

\textsuperscript{1}Corresponding author

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Formation and applications of the sedimentary record in arc collision zones
Edited by Amy E. Draut, Peter D. Clift, and David W. Scholl
Abstract  The Catalina Schist underlies the inner southern California borderland of southwestern North America.  On Santa Catalina Island, amphibolite facies rocks that recrystallized and partially melted at ca. 115 Ma and 40 km depth occur atop an inverted metamorphic stack that juxtaposes progressively lower-grade, high-P/T rocks across low-angle faults. This inverted metamorphic sequence has been regarded as having formed within a newly initiated subduction zone. However, subduction initiation at ca. 115 Ma has been difficult to reconcile with regional geologic relationships since the Catalina Schist formed well after emplacement of the adjacent Peninsular Ranges batholith (PRB) had began in earnest. New detrital zircon U-Pb age results indicate that the Catalina Schist accreted over an ~20 m.y. interval. The amphibolite unit metasediments formed from Latest Neocomian to early Aptian (122 to 115 Ma) craton-enriched detritus derived mainly from the pre-Cretaceous wallrocks and Early Cretaceous volcanic cover of the PRB. In contrast, lawsonite-blueschist and lower grade rocks derived from Cenomanian sediments dominated by PRB plutonic and volcanic detritus were accreted between 97-95 Ma. Seismic data and geologic relationships indicate that the Catalina Schist structurally underlies the western margin of the northern PRB. We propose that construction of the Catalina Schist complex involved underthrusting of the Early Cretaceous forearc rocks to a subcrustal position beneath the western PRB. The heat for amphibolite facies metamorphism and anatexis observed within the Catalina Schist was supplied by the western PRB while subduction was continuous along the margin. Progressive subduction erosion ultimately juxtaposed the high-grade Catalina Schist with lower grade blueschists accreted above the subduction zone by 95 Ma. This coincided with an eastern relocation of arc magmatism and emplacement of the ca. 95 Ma La Posta tonalite-trondjhemite-granodiorite suite of the eastern PRB. Final assembly of the Catalina Schist marked the initial stage of the Late Cretaceous-early Tertiary craton-ward shift of arc
magmatism and deformation of southwestern North America that culminated in the Laramide orogeny.

**Key Words**: Catalina Schist, U-Pb, zircon, subduction erosion

**INTRODUCTION**

Subduction erosion is a fundamental convergent margin process that tectonically shortens the forearc region by displacing rocks from the overriding plate to sub-crustal positions and/or into the asthenosphere (Scholl et al., 1980; Von Huene and Scholl, 1990, 1992). Roughly half of all present-day convergent margins exhibit geologic and/or geophysical evidence for subduction-related removal of rocks from the forearc regions of the overriding plate to positions beneath the arc and/or into the mantle (e.g., Clift and Vannucchi, 2004). The globally average rate at which continental debris is transported towards the mantle is significant (~2.5 km³/year; Scholl and Von Huene, in press). Unfortunately, its consumptive nature means that direct evidence for subduction erosion in ancient convergent margins is generally ambiguous or lacking. Indirect evidence for subduction erosion such as “missing” or “telescoped” forearc crust is difficult to distinguish from alternative tectonic processes such as margin-parallel strike-slip faulting (e.g., Karig, 1980). A classic illustration of the difficulty in deciding between subduction erosion and margin-parallel strike-slip dispersal of forearc rocks is provided by the Nacimiento fault in west-central California (Page, 1970). There, plutonic and metamorphic rocks of the Salinian block are juxtaposed against high-pressure, low-temperature mélangé of the Sur-Obispo terrane. This anomalous juxtaposition of lithotectonic belts has been explained both by underthrusting (subduction erosion) of intervening rocks (Page, 1981; Hall, 1991) or by lateral transport of hundreds (Hill and Dibblee, 1953; Suppe, 1970; Dickinson, 1983; Dickinson et al., 2005) to
thousands of kilometers (Page, 1982; McWilliams and Howell, 1982; Vedder et al., 1983; Debiche et al., 1987).

Like the Sur-Obispo terrane, seismic data and geologic relationships indicate that the high-pressure/temperature Catalina Schist of southern California is also in direct contact with its coeval magmatic arc. This relationship was revealed by middle Miocene extension which exhumed the Catalina Schist from beneath the northwestern margin of the Peninsular Ranges batholith (PRB) across an east-dipping Miocene detachment fault (Crouch and Suppe, 1993) (Fig. 1). The dramatic extension of the borderland occurred during a middle Miocene microplate capture event (e.g., Nicholson et al., 1994) triggered by Pacific-North American shearing along the margin. During this event, Late Cretaceous forearc strata and underlying Jurassic and earliest Cretaceous basement rocks were ripped from the western margin of the PRB, rotated clockwise, and translated northwestward into the outer California borderland along a detachment system underlain by Catalina Schist (Wright, 1991; Crouch and Suppe, 1993; Bohannon and Geist, 1998; Ingersoll and Rumelhart, 1999).

While the Catalina Schist contains abundant blueschist facies rocks typical of subduction zone environments it also has higher-temperature lithologies which define an inverted metamorphic sequence that includes amphibolite facies rocks at the highest structural levels (see inset in Fig. 1). These partly melted rocks experienced peak grade conditions at ca. 115 Ma (Suppe and Armstrong, 1972; Mattinson, 1986; Sorensen and Barton, 1987; Grove and Bebout, 1995; Anczkiewicz et al., 2004). A longstanding explanation for the unusually high-temperature, inverted metamorphism of the Catalina Schist is that it reflects progressive accretion of early subducted rocks beneath unrefrigerated mantle lithosphere during nascent subduction (e.g., Platt, 1975). Although Platt’s (1975) nascent subduction hypothesis was based upon the Catalina
Schist, he extended it to explain the origin of meter- to kilometer-scale high-grade blocks within the Franciscan Complex and other accretionary complexes along the western margin of North America (e.g., Platt, 1975; Cloos, 1985; Wakabayashi, 1990, 1992, 1999, Anczkiewicz et al., 2004; Wakabayashi and Dumitru, 2007).

The nascent subduction hypothesis makes sense for the Franciscan Complex because the blocks are coeval with a major pulse of Middle Jurassic magmatic activity along the western North American continental margin (Evernden and Kistler, 1970; Saleeby and Sharp, 1980; Stern et al., 1981; Chen and Moore, 1982; Wright and Fahhan, 1988; Barton et al., 1988; Staude and Barton, 2001; Irwin, 2002). The Middle Jurassic orogenesis included formation of the Coast Range ophiolite (Hopson et al., 1981; Shervais et al., 2005) that is considered by some (e.g., Stern and Bloomer, 1992) to be a product of Middle Jurassic subduction initiation (see also Dickinson et al., 1996). The equivalence in age of the Coast Range ophiolite and the high-grade blocks of the Franciscan complex corresponds well with a common relationship exhibited by many of the world’s ophiolites: namely age equivalence of the ophiolite and their metamorphic sole (Jamieson, 1986; Hacker, 1990a,b; Wakabayashi and Dilek, 2000).

While establishment of a new subduction regime during the middle Jurassic may generally explain the occurrence of high-grade rocks of this age within the Franciscan Complex of central and northern California, an equivalent event at ca. 115 Ma off the southern California margin makes little sense with respect to the convergent margin evolution of southwest North America. Emplacement of the 750 km long PRB had begun by Middle Jurassic time (Shaw et al., 2003) and was well underway prior to the proposed subduction initiation event at 115 Ma (Silver and Chappell, 1988; Kistler et al., 2003). Furthermore, no mid-Cretaceous ophiolite exists within the southern California region. Only fragments of older forearc basement are
preserved. The latter is best represented by Triassic and Middle Jurassic ophiolitic rocks that border the southern portion of the PRB (Kimbrough and Moore, 2003). Just as in the case of the Franciscan complex, the Middle Jurassic ophiolite of west-central Baja California is associated with high-pressure/temperature blocks of equivalent age on the Vizcaino Peninsula and Cedros Island (Fig. 2; Baldwin and Harrison, 1989, 1992).

This paper presents new detrital zircon U-Pb results from metagraywackes of the major metamorphic units of the Catalina Schist that clarify its accretion history. We confirm a genetic relationship to forearc and batholith rocks of the adjacent PRB. These new results indicate that the anomalously high-T, high-P/T amphibolite facies rocks of the Catalina Schist had been deposited, accreted, and metamorphosed at peak grade conditions ~15-20 m.y. before the lawsonite-blueschist and lower-temperature, high-P/T rocks of the complex were accreted. We conclude that amphibolite facies, epidote-amphibolite, and possibly epidote blueschist units of the Catalina Schist formed from forearc strata and basement rocks that were underthrust and sheared together with lithospheric mantle beneath the western margin of the PRB in a subduction erosion process. Our model is consistent with recent deep seismic reflection imaging of subparallel megathrusts within modern subduction zones (e.g., Calvert, 2004). Continued subduction erosion ultimately juxtaposed the high-grade rocks of the Catalina Schist with lawsonite-blueschists and lower-grade rocks that had formed within a subduction zone setting.

BACKGROUND

Catalina Schist High-Pressure/Temperature Complex

High-P/T rocks distributed along the western margin of southern and Baja California appear to represent the southern continuation of the better exposed Franciscan Complex of central and northern California (Fig. 1; Woodford, 1924; Suppe and Armstrong, 1972; Kilmer,
Within the southern California region, on-land exposures of a high-P/T complex referred to as the Catalina Schist occur on Santa Catalina Island (Bailey, 1941; Platt, 1976) and the Palos Verdes peninsula (Woodring et al., 1946; Dibblee, 1999). More widespread submarine exposures of Catalina Schist occur throughout the inner southern California borderland and equivalent basement is detected in boreholes within the western and southwestern Los Angeles basin (Schoellhamer and Woodford, 1951; Yerkes et al., 1965; Yeats, 1968, 1973; Sorensen, 1985, 1988; Wright, 1991; Crouch and Suppe, 1993; Bohannon and Geist, 1998; ten Brink et al., 2000).

The Catalina Schist on Santa Catalina Island has been described by Woodford (1924), Bailey (1941), Platt (1975, 1976), Sorensen and Barton (1987), Sorensen (1986, 1988a, 1988b), Sorensen and Grossman (1989), Bebout and Barton (1989, 1993, 2002), and Grove and Bebout (1995) and is only briefly summarized here. It consists of metasedimentary, metavolcanic, and ultramafic protoliths that were metamorphosed and sheared together under amphibolite facies to lawsonite-blueschist facies and lower-grade conditions during the mid-Cretaceous (Suppe and Armstrong, 1972; Platt, 1976; Mattinson, 1986; Sorensen and Barton, 1987; Sorensen, 1988; Grove and Bebout, 1995). Individual tectonic slices (Fig. 1) contain rocks of broadly equivalent metamorphic grade. These are juxtaposed across low-angle faults in an apparent inverted metamorphic sequence (Platt, 1976; Fig. 1). The structurally highest unit is an amphibolite facies shear zone composed primarily of intercalated and metasomatically altered mafic and former harzburgite/dunite protoliths (Sorensen and Barton, 1987; Sorensen, 1988b; Bebout and Barton, 1993, 2002). The proportion of sediment, particularly immature graywacke, increases structurally downward. At the structurally lowest levels, the lawsonite-blueschist and lower grade units are sediment-dominated. Each of the major metamorphic units is characterized by
meter- to kilometer-scale, compositionally heterogeneous shear zones that appear to have facilitated metasomatic fluid infiltration at near peak-grade conditions (Bebout and Barton, 1989, 1993, 2002; King et al., 2006, 2007).

Primary geologic mapping of the Catalina Schist on Santa Catalina Island was performed by Bailey (1941) and Platt (1976). Although we rely heavily upon Platt’s (1976) mapping of the central part of the island, we recognize different tectonometamorphic units than he did. Platt’s (1976) Catalina greenschist unit has been subdivided into epidote-amphibolite and epidote-blueschist units (Grove and Bebout, 1995). Results presented in Grove and Bebout (1995) and in this paper indicate that the epidote-amphibolite and epidote-blueschist rocks had discernable differences in accretion and metamorphic history. In addition, we have delineated lawsonite-albite and actinolite-albite facies units on the west end of the island. These lower grade rocks appear to structurally underlie a lawsonite-blueschist unit (Altheim et al., 1997).

Because the Catalina Schist terrane was significantly extended as it was exhumed during the early middle Miocene formation of the inner continental borderland (Wright, 1991; Crouch and Suppe, 1993; Nicholson et al., 1994; Bohannon and Geist, 1998; Ingersoll and Rumelhart, 1999), original contacts bounding the major units within the schist are likely to have been strongly modified or excised altogether by mid-Tertiary structures. In addition, the extent to which the Santa Catalina Island exposures of Catalina Schist represent the much larger area of high-P/T rocks that underlie the inner continental borderland is difficult to assess. Study of widely distributed deposits of San Onofre Breccia indicates that the Santa Catalina Island exposures are probably broadly representative of the overall terrane (Stuart, 1979). The San Onofre Breccia is an early middle Miocene deposit that accumulated as the Catalina Schist was exhumed (Wright, 1991; Crouch and Suppe, 1993). Clast counts performed by Stuart (1979)
appear to be broadly consistent with an unroofing sequence in which amphibolite and epidote-
amphibolite grade Catalina Schist and low-P/T mafic basement from the hanging wall were
much more abundant at lower stratigraphic levels within the San Onofre Breccia. Our own
observations confirm this relationship for the thick deposits of San Onofre breccia that
accumulated along the western margin of the PRB (Fig. 1). In contrast, stratigraphically younger
San Onofre Breccia, including deposits laid down atop extended borderland crust underlain by
Catalina Schist, tends to be dominated by lawsonite-blueschist and lower-grade detritus. Based
upon these relationships, we are reasonably confident that our sampling of Catalina Schist on
Santa Catalina Island is broadly representative of the Catalina Schist terrane as a whole.

Forearc Basement and Strata

The relatively complete forearc sequence of the Vizcaíno Peninsula at the southern end of
the Peninsular Ranges batholith provides the basis to interpret the forearc of the highly disrupted
southern California borderland (Fig. 1, 2). On the Vizcaíno Peninsula and on Cedros Island,
both Late Triassic (221 Ma) and Middle Jurassic (173 Ma) ophiolite sequences (Moore, 1985;
Kimbrough, 1985; Kimbrough and Moore, 2003) are the basement for Upper Jurassic, Lower
Cretaceous, Upper Cretaceous, and lower Cenozoic arc volcanic and forearc sedimentary rocks
(Boles, 1978; Kilmer, 1979; Kienast and Rangin, 1982; Boles and Landis, 1984; Patterson, 1984;
Smith and Busby, 1993; Busby et al., 1998; Kimbrough et al., 2001; Critelli et al., 2002; Busby,
2004). High-P/T Cretaceous rocks of the Western Baja terrane constitute an accretionary
complex that is juxtaposed beneath the Triassic and Jurassic ophiolitic rocks well outboard of the
batholithic margin (Suppe and Armstrong, 1972; Moore, 1986; Sedlock, 1988a, 1988b; Baldwin
and Harrison, 1989; Baldwin et al., 1992).
Only fragmentary evidence exists for the early Mesozoic forearc basement within the southern California area. Within the outer California continental borderland, offshore drilling and seismic exploration detect mafic basement beneath forearc strata (Bohannon and Geist, 1998; ten Brink et al., 2000). Surface exposures of lower Mesozoic basement rocks occur on Santa Cruz Island (the Willows Complex of Weaver and Nolf, 1969; Hill, 1976; Mattinson and Hill, 1976) are correlated with the Coast Range ophiolite (Jones et al., 1976). The associated Santa Cruz Island Schist and correlative Santa Monica slate of the western Transverse Ranges exhibit distinctly arc-like compositions and could be equivalent to accreted Jurassic terranes of the Sierran Foothills (Sorensen, 1985, 1988a). “Saussurite” gabbro similar to altered zones of the Willows Complex gabbros is in fault contact with the Catalina Schist on Santa Catalina Island (Platt, 1976) and within the subsurface of the southwestern Los Angeles basin (Schoellhamer and Woodford, 1951; Yeats, 1973; Sorensen, 1985, 1988). Saussurite gabbro clasts are a major component within the San Onofre Breccia (Stuart, 1979) and confirm that low-P/T mafic basement was exhumed along with the Catalina Schist during middle Miocene borderland rifting.

Cenomanian and younger forearc strata onlap the western margin of the northern Peninsular Ranges batholith (Fig. 2; Woodring and Popenoe, 1942; Yerkes et al., 1965; Flynn, 1970; Nordstrom, 1970; Peterson and Nordstrom, 1970; Kennedy and Moore, 1971; Sundberg and Cooper, 1978; Schoellhamer et al., 1981; Nilsen and Abbott, 1981; Bottjer et al., 1982; Bottjer and Link, 1984; Fry et al., 1985; Girty, 1987; Bannon et al., 1989). Thick sections of Upper Cretaceous strata also occur throughout the outer borderland (Howell and Vedder, 1981; Vedder, 1987; Bohannon and Geist, 1998). The existence of Lower Cretaceous forearc strata is far less certain however. Sedimentary rocks of this age have not been described anywhere along
the western margin of the northwestern Peninsular Ranges, the Santa Monica Mountains, nor on the offshore islands. Their existence in the subsurface is mostly inferred in all areas that have been drilled (Howell and Vedder, 1981; Vedder, 1987; Bohannon and Geist, 1998; ten Brink et al., 2000). Our model for the formation of the Catalina Schist explicitly accounts for the scarcity of these Lower Cretaceous forearc rocks in the southern California region (see Discussion).

Peninsular Ranges Batholith

The Peninsular Ranges batholith of southern and Baja California constitutes a classic Cordilleran continental margin batholith (Fig. 2; Larsen, 1948; Jahns, 1954; Gastil et al., 1975). The better-studied northern segment of the PRB consists of longitudinal western and eastern zones based on age, petrology, prebatholithic wall rock, geophysical parameters, and depth and style of emplacement (Gastil et al., 1981; Baird and Miesch, 1984; Taylor, 1986; Gromet & Silver, 1987; Silver and Chappell, 1988; Ague & Brimhall, 1988; Todd et al., 1988; Hill & Silver, 1988; Gastil, 1993; Johnson et al., 1999; Lovera et al., 1999; Todd et al., 2003; Kistler et al., 2003; Langenheim and Jachens, 2003). The oldest recognized intrusive rocks are gneissic S-type plutons that occupy the medial zone of the PRB and yield Middle Jurassic emplacement ages (Todd and Shaw, 1985; Thompson and Girty, 1994; Shaw et al., 2003; Kistler et al., 2003). Cretaceous plutons as old as 140 Ma also occur within the PRB (Silver and Chappell, 1988; Alsleben et al., 2005; D.L. Kimbrough, unpublished data). These are cut by a regionally extensive dike swarm emplaced at ca. 130 to 120 Ma (Böhnel et al., 2002; D.L. Kimbrough, unpublished data). The well-developed western zone of the PRB is composed mainly of 125 to 100 Ma gabbro to monzogranite plutons with primitive island arc geochemical affinities. The eastern zone of the batholith is defined by a belt of large-volume, 95±3 Ma tonalite, trondjhemite, and low-K granodiorite (TTG) plutons (Gastil et al., 1975; Silver & Chappell,
1988; Walawender et al., 1990) that comprise the La Posta TTG suite (Tulloch and Kimbrough, 2003; Kimbrough and Grove, 2006). These Na- and Al-rich plutons have a deep, garnet-present, melt source signature, as seen in high Sr, Ba, Sr/Y, and La/Yb. Field and thermobarometric data indicate emplacement at ~2 to 6 kbar pressures (Rothstein and Manning, 2003) followed by rapid Cenomanian-Turonian uplift and denudation at rates of ~1-2 mm/yr (Krummenacher et al., 1975; Lovera et al., 1999; Johnson et al, 1999; Kimbrough et al., 2001; Ortega Rivera, 2003; Grove et al., 2003b). A delayed late Campanian-Maastrichtian phase of uplift (e.g., Krummenacher et al., 1975) appears to be related to Laramide shallow subduction and removal of the deep crustal and lithospheric roots of the La Posta belt (Lovera et al., 1999; Grove et al., 2003a, 2003b).

SAMPLING AND METHODS

Detrital Zircon U-Pb Age Measurements

A total of 645 U-Pb zircon ages were measured from 33 samples of amphibolite through sub-blueschist facies metagraywacke of Catalina Schist collected from Santa Catalina Island. Many of the samples examined were previously studied by Grove and Bebout (1995). We focused upon metasedimentary rocks from each of the major metamorphic units in order to obtain an upper bound upon the depositional age of their sedimentary protolith and to determine sediment provenance. In our U-Pb analysis of zircons from the Catalina Schist, we employed secondary ionization mass spectrometry (SIMS) methods using the UCLA Cameca ims 1270 ion microprobe. The extensive metamorphic recrystallization that affected zircons from the amphibolite and epidote-amphibolite units was significantly ameliorated by the high spatial resolution of the ion microprobe in conjunction with cathodoluminence (CL) imagery (e.g., only ~1 nanogram of sputtered zircon required to yield a U-Pb age). The techniques are described in
Grove et al. (2003b) with additional details included in the data repository\(^1\). We analyzed conventionally sectioned and polished zircons in epoxy mounts. The zircons were hand-selected from heavy mineral concentrates produced from standard crushing, density, and magnetic methods. Further details are included in the data repository\(^1\). Because the number of zircons measured from individual samples was typically small due to the low yields realized during mineral separation, we pooled data from multiple samples to obtain statistically meaningful results for each of the major tectonic units of the Catalina Schist.

As noted above, metamorphic zircon growth was a significant issue for some of the grains we examined, particularly for the metasediments from the amphibolite unit. Anczkiewicz et al. (2004) also describe evidence for metamorphic zircon growth in reconnaissance U-Pb zircon measurements they reported from the Catalina Schist amphibolite unit. While we selected our analysis sites on the basis of morphologic and optical criteria in CL imagery, a significant number of the spot analyses from zircons of the amphibolite and epidote-amphibolite units overlapped regions affected by metamorphic recrystallization. Metamorphic overgrowths on igneous zircons generally have Th/U < 0.1; Kröner et al., 1994; Rubatto, 2002; Williams and Claesson, 1987). We have found a cutoff of Th/U = 0.1 to be empirically well supported by detrital zircon results from the Cenomanian and younger, pluton-derived strata that overlie the PRB. In our sampling of these units (Mahoney et al., 2006), 1506 out of 1527 or 98.6% of the zircon U-Pb analyses yielded Th/U values equal to or greater than 0.1 (Fig. 3A). The average Th/U value measured was 0.48 ± 0.41, with 3.3% of the analyses falling within the Th/U = 0.1 to 0.2 range. Thus a cutoff of Th/U = 0.1 is useful for identifying detrital igneous zircons adversely

\(^1\)GSA Data Repository item 200XXXX, Table DR1 and references for localities, is available online at www.geosociety.org/pubs/ft2006.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.
affected by metamorphic recrystallization. Accordingly, we have excluded all analyses with Th/U < 0.1 from the summary plots or calculations presented below (Fig. 3B).

Complete data tables of U-Pb age measurements, lithologic descriptions, and sample locations are available from the GSA data repository\(^1\). In this paper, quoted uncertainties are ±1σ errors unless otherwise specified. The U-Pb ages are generally \(^{206}\text{Pb}/^{238}\text{U}\) values for < 1 Ga zircons and \(^{207}\text{Pb}/^{206}\text{Pb}\) ages for older grains. Analyses with age uncertainties exceeding 10% have been discarded. The U-Pb ages of zircons that exhibited resolvable \(^{206}\text{Pb}/^{238}\text{U}\) vs. \(^{207}\text{Pb}/^{235}\text{U}\) discordance are \(^{207}\text{Pb}/^{206}\text{Pb}\) values.

**Rb-Sr Measurements**

To clarify the significance of previous \(^{40}\text{Ar}/^{39}\text{Ar}\) phengite step-heating results (Grove and Bebout, 1995) that had indicated a possible Middle Jurassic age for garnet-bearing blueschist blocks within the lawsonite-blueschist mélange, we carried out Rb-Sr measurements with one of the blocks (330-4B) that had been analyzed in this previous study. All column chemistry and mass spectrometer measurements were carried out at UCLA. There, Rb and Sr are separated in cation exchange columns containing AG50W-X8 resin, using 2.5N HCl. The Sr isotopic compositions are measured with a VG54-30 multicollector thermal ionization mass spectrometer. \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios are normalized to \(^{86}\text{Sr}/^{88}\text{Sr}=0.1194\). In this analysis session, the measured \(^{87}\text{Sr}/^{86}\text{Sr}\) value for our Sr standard (NBS 987) value was \(^{87}\text{Sr}/^{86}\text{Sr}=0.710239±16\) (2σ, n=13). The Rb and Sr concentrations are measured by isotope dilution using VG54-30. Model Rb-Sr isochron ages are calculated using ISOPLOT/Ex Version 3 (Ludwig, 2003).

**RESULTS**

**Amphibolite Facies**
Five biotite + muscovite + garnet ± kyanite-bearing metagraywackes from the amphibolite unit yielded abundant subhedral to well-rounded zircon. Alternatively, both metachert samples examined contained only metamorphic zircon. Although external overgrowths <1 to 5 µm or more were common, zircons also exhibited internal areas of patchy diffuse recrystallization in CL imagery. Of the 169 analyses, 48 (28%) overlapped metamorphic zircon based upon Th/U (Fig. 3B). About half of the affected analyses yielded obviously mixed ages. U-Pb ages calculated for the remaining analyses with Th/U < 0.1 fell between 107-126 Ma with a peak at 116 ± 6 Ma. This result is in good agreement with independent estimates for the timing of amphibolite unit recrystallization (ca. 115 Ma; Suppe and Armstrong, 1972; Mattinson, 1986; Grove and Bebout, 1995; Anczkiewicz et al., 2004).

The data for zircons with Th/U > 0.1 clearly indicate that the sedimentary protolith of the amphibolite facies metagraywackes contained a large proportion of craton-derived detritus (Fig. 4A). Roughly 50% of the analyses were from Middle Proterozoic zircon (Fig. 4G). Unfortunately, many of these grains exhibited variable Pb loss and high degrees of discordance. This condition complicates characterization of the Middle Proterozoic age distribution. Although $^{207}$Pb/$^{206}$Pb ages more accurately approximate the crystallization age than the highly discordant U-Pb ages, they are insufficiently precise to reveal strong clustering of ages. Nevertheless, the subdued maxima that occur at 1.15, 1.40, and 1.65-1.80 Ga are compatible with a southwestern North American provenance. Age peaks also occur at 126, 145, and 162 Ma. The five youngest zircons measured from the amphibolite unit metagraywackes yield a weighted mean age of 122 ± 3 Ma. These analyses yield a mean Th/U value of 0.30 ± 0.20 and were from oscillatory zoned regions in Cl-imagery that likely reflect igneous crystallization.
Accordingly, we regard 122 ± 3 Ma as a geologically meaningful upper bound upon the depositional age of the sedimentary protolith.

**Epidote-Amphibolite Facies**

Five muscovite ± biotite ± garnet epidote-amphibolite facies metagraywackes yielded zircon that was generally more subhedral to euhedral grains than the well-rounded grains common in the amphibolite unit metagraywackes. Cathodoluminence imaging and Th/U indicate that metamorphic recrystallization of detrital grains was less common than in the amphibolite unit. Only 13 of 135 or 10% of analyses of the zircons from epidote-amphibolite facies metagraywackes had Th/U < 0.1. The U-Pb age distribution indicates a diminished cratonal provenance relative to amphibolite facies metagraywackes; only about 20% of the zircons yield Middle Proterozoic ages (Fig. 4G). Although fewer analyses are available, the overall age distribution of Middle Proterozoic zircons in epidote-amphibolite metagraywackes is broadly similar to that of the amphibolite facies rocks (Fig. 4B). The relative lack of Proterozoic grains is accompanied by a proportionate increase of < 200 Ma grains. Distinct peaks are present at 115 and 126 Ma with a strong peak at 150 Ma (Fig. 4B). The five youngest detrital zircons from the epidote-amphibolite metagraywackes yield an age of 113 ± 3 Ma, with a mean Th/U value of 0.42 ± 0.05.

**Epidote-Blueschist Facies**

Thirty analyzable zircons were recovered from a single metagraywacke sample intercalated with Na-amphibole + clinozoisite + albite-bearing mafic rocks (glaucophanic greenschists of Platt, 1975; Sorensen, 1986). Six similar-appearing samples from equivalent field settings failed to yield zircon. This poor zircon yield probably reflects abundant volcanic detritus in the protolith. Nearly all of the zircons recovered consisted of clear subhedral to
euhedral grains. There was no evidence for metamorphic zircon growth either petrographically or in terms of measured Th/U. Two distinct maxima occur at 103 and 142 Ma (Fig. 4C). Although results from the epidote-blueschist facies unit are transitional between those obtained from the amphibolite and epidote-amphibolite facies units and lower grade portions of the Catalina Schist (Fig. 4G), the epidote-blueschist facies unit appears more closely allied with the lower grade units based upon the age of the youngest zircons detected in each population. The five youngest zircons measured yielded an average U-Pb age of 101±3 Ma with a mean Th/U of 0.63 ± 0.23.

**Lawsonite-Blueschist Facies and Lower Grade Rocks**

Data bearing upon the sedimentary protoliths of the lawsonite-blueschist and lawsonite-albite, and albite-actinolite facies units are discussed together because of the highly similar results produced (Fig. 4D-G). Overall, we obtained 164 analyses from 11 lawsonite-blueschist facies rocks; 59 analyses from 3 albite-actinolite facies rocks, and 82 analyses from 7 lawsonite-albite facies metagraywackes. No metamorphic zircon growth was detected; Th/U values were all > 0.1. The Ca-rich metagraywackes prevalent in low-grade portions of the Catalina Schist overwhelmingly contain clear subhedral to euhedral grains. Most of these grains yield U-Pb ages of 95 to 130 Ma (Fig. 4D-F). Late Jurassic-Early Cretaceous zircon was greatly subordinate and only trace quantities of Proterozoic zircon were present. Because the youngest zircon U-Pb ages measured from the three lowest-grade units all agreed within error (96 ± 3 Ma, 97 ± 3 Ma, and 98 ± 3 Ma for lawsonite-blueschist, actinolite-albite, and lawsonite-albite respectively), they define a 97 ± 3 Ma upper bound for the depositional age of the sedimentary protolith for all of the low-grade Catalina Schist units.

**Garnet-bearing Blueschist Block in Lawsonite-Blueschist Melange**
Garnet-bearing blueschist blocks occur within melange zones in the lawsonite-blueschist unit. Phengite $^{40}\text{Ar}^{39}\text{Ar}$ results for two such blocks indicate that they formed prior to peak-grade recrystallization in the amphibolite unit (Grove and Bebout, 1995). To confirm this relationship we undertook Rb-Sr measurements of phengite, Na-amphibole, and whole rock for one of the samples (330-4B; Fig. 5A). Results are shown in Figure 5B. Phengite, Na-amphibole, and whole rocks results do not define a statistically meaningful isochron. However, the model Rb-Sr age defined by phengite + Na-Amphibole, 135 Ma, is identical to the total gas age calculated from $^{40}\text{Ar}^{39}\text{Ar}$ step-heating results obtained from 330-4B phengite.

**DISCUSSION**

**Evaluation of the Nascent Subduction Model for the Catalina Schist**

The origin of the amphibolite facies rocks of the Catalina Schist is central to assessing the tectonic significance of this terrane. Amphibolite facies metamorphism and anatexis are atypical features of subduction complexes and are thought to reflect unusual processes along a convergent margin. High-temperature conditions accompany the earliest developmental stages of subduction, before the hanging wall is refrigerated, (Platt, 1975; Cloos, 1985; Peacock, 1987). Although ridge subduction, and/or slow underflow of very young oceanic crust (e.g., Peacock, 1987, 1992; Hacker, 1990, 1991; Peacock et al., 1994) can also produce high-temperature metamorphism in subduction zones, there is no evidence that either process affected the mid-Cretaceous southwestern margin of North America.

The detrital zircon results (Fig. 6) require substantial modification of Platt’s (1975) hypothesis that amphibolite through blueschist facies rocks formed in an inverted metamorphic aureole in response to transient heating during nascent subduction at ca. 115 Ma. Sequential accretion of the major tectonic units of the Catalina Schist over an ~20 m.y. interval is required.
As indicated in Figure 6A, the protolith of the amphibolite unit metagraywackes was deposited between 122 Ma (youngest detrital zircon ages) and 115 Ma (the time of peak-grade recrystallization). In contrast, the sedimentary protolith of the lawsonite-blueschist and lower grade units was deposited after 97 Ma or at least 18 m.y. after peak-grade recrystallization of the amphibolite unit (Fig. 6C-F).

Available data for the Catalina Schist amphibolite unit indicate that it cooled slowly (ca. 25°C/m.y.) from peak grade conditions (650 to 700°C) at 115 Ma to K-Ar muscovite closure temperatures (350-400°C) at 105 to 100 Ma (Fig. 6A; Grove and Bebout, 1995). The Catalina epidote amphibolite unit also cooled slowly (Fig. 6B): a 7 to 12 m.y. interval separated peak-grade recrystallization of epidote amphibolite mafic gneisses and cooling through phengite K-Ar closure. Slow cooling exhibited by the amphibolite unit is compatible with very slow subduction or subduction of very young oceanic lithosphere. However, environments such as these are far too warm to permit formation of lawsonite-blueschist and lower temperature metamorphic rocks.

Within the Catalina Schist, only the lawsonite-blueschist and lower grade rocks record metamorphic conditions that require formation within a low-T, high-P/T subduction zone. The amphibolite and epidote-amphibolite units were retrograded under greenschist facies conditions, as was the epidote blueschist unit (Platt, 1976; Sorensen, 1986; Bebout and Barton, 1989; Grove and Bebout, 1995). Given their present association with lawsonite blueschist and lower-grade rocks, we find it remarkable that the amphibolite and epidote-amphibolite units exhibit such little evidence for equilibration at blueschist facies conditions (Platt, 1976; Sorensen, 1986; Bebout and Barton, 1989; Grove and Bebout, 1995). Blueschist facies overprinting of higher grade rocks is the rule for subduction complexes of the North American Cordillera (Ernst, 1988). For example, tectonic blocks of garnet amphibolite in the Franciscan and Shuksan subduction
complexes are heavily overprinted by blueschist facies assemblages (e.g., Brown et al., 1982; Cloos, 1985; Wakabayashi, 1990). In contrast, evidence for blueschist facies overprinting within the Catalina Schist amphibolite unit is obscure and limited to sparse veining by pumpellyite and trace amounts of lawsonite in white mica + zoisite + albite assemblages that replace oligoclase in amphibolite facies rocks (Grove and Bebout, 1995). We interpret the available thermochronology and greenschist facies overprinting of the high-grade units of the Catalina Schist to indicate that they occupied a subcrustal position characterized by 400-500°C temperatures prior to 100 to 95 Ma. The lack of blueschist facies overprinting within the amphibolite and epidote-amphibolite units indicates to us that these rocks were sufficiently distant from an active subduction zone to be largely unaffected by the low geothermal gradient (<5-10°C/km) recorded by the Franciscan and other Cordilleran subduction complexes during the mid-Cretaceous.

Provenance Ties Linking the Catalina Schist to the Peninsular Ranges Batholith

King et al. (2007) measured whole rock Pb isotopes from mélange matrix sampled from the amphibolite, lawsonite-blueschist, and lawsonite-albite units of the Catalina Schist. Although details of the U-Th-Pb systematics indicate the possible elemental redistributions during the devolatization history of the schist (see King et al., 2007), it is remarkable how well the measured Pb isotopic compositions of the sediment-dominated lawsonite-blueschist and lawsonite-albite mélange matrix agree with those measured from granitoids of the adjacent PRB (Fig. 7A and 7B; Kistler et al., 2003). The whole rock Pb isotopic compositions from the mafic and ultramafic-dominated mélange of the Catalina Schist amphibolite unit also overlap strongly with the PRB (Fig. 7C) but tend to more radiogenic values than the lower grade units (Fig. 7A and 7B). The amphibolite unit mélange results are more difficult to interpret than those from the
lower grade units because the sources of Pb are far less uncertain due to the scarcity of sedimentary material within the mélange (see Bebout and Barton, 1993, 2002). Nevertheless, they are consistent with a cratonal provenance similar to that exhibited by southeastern Arizona (Wooden and Miller, 1990) and can be differentiated from the Transverse Ranges/Mojave province (Fig. 7C; Barth et al., 1995, J. Wooden, D. Coleman, and A. Barth, unpublished data). Because the early Mesozoic wallrocks of the PRB are dominated by craton-derived Proterozoic zircon that indicate a similar source (see below), we speculate that the Proterozoic zircon contributed by from the cratonally-derived sediment eroded from PRB wallrocks recrystallized during mélange formation, liberating the highly radiogenic Pb that was dispersed throughout the amphibolite unit mélange as it was recrystallized and metasomatically altered (Bebout and Barton, 1993, 2002; King et al., 2007).

In the same manner as the Pb isotope data discussed above, the detrital zircon results from the Catalina Schist are also indicate a close genetic relationship between the Catalina Schist and the PRB: (1) Early Mesozoic, Paleozoic, and Proterozoic detrital zircon from the two terranes are similar; and (2) mid-Cretaceous detrital zircon age distributions from the lawsonite-blueschist and lower-grade Catalina Schist are virtually identical to the distribution of crystallization ages from the northern PRB and the detrital age distribution from the adjacent Upper Cretaceous forearc rocks.

Lower Mesozoic flysch wallrocks of the PRB crop out within the deeply denuded axial zone of the northern batholith (Fig. 2) and represent a major potential source of Early Mesozoic, Paleozoic, and Proterozoic zircon. A composite detrital zircon distribution based upon results obtained by Morgan et al. (2005) from three different localities in southern California (Bedford Canyon Fm., French Valley area, and Julian Schist; localities 1-3 in Fig. 2) and northern Baja
California (Vallecitos area; locality 4 in Fig. 2) is shown in Figure 8. As indicated, the age distribution defined by > 200 Ma detrital zircon from the Catalina Schist (Fig. 8A) is consistent with a source region similar to the PRB wallrocks (Fig. 8B). The expected age distribution for detritus originating from the southwest North American craton is represented by late Miocene to Recent sands from the Colorado River system (Fig. 8C). Well-defined peaks at 1.0 to 1.2, 1.45, and 1.65 to 1.75 Ga are distinctive of the basement assemblage (Gehrels and Stewart, 1998). Early Paleozoic and latest Neoproterozoic (400-650 Ma) zircon is contributed by supracrustal cover rocks that were sourced from the Appalachian and Ouachita orogenic belts (Dickinson and Gehrels, 2003). Finally, Permo-Triassic zircon is derived from the earliest Phanerozoic magmatic arcs established along the southwestern North American margin (Barth and Wooden, 2006; González-León et al., 2006).

The lawsonite-blueschist and lower-grade rocks of the Catalina Schist contain Upper Cretaceous forearc strata of the Peninsular Ranges are very similar (Fig. 8E), as is the distribution of U-Pb zircon crystallization ages from the adjacent batholith (Fig. 8 F; Silver and Chappell, 1988; Walawender et al., 1990; Kistler et al., 2003; D.L. Kimbrough, unpublished data). The fact that the Catalina Schist results are skewed to slightly older ages is readily understood in terms of the Cenomanian age of the protolith. Cenomanian age forearc strata from the PRB are also skewed to older ages since they were deposited prior to massive exhumation of the ca. 95 Ma La Posta plutonic suite within the eastern batholith (Mahoney et al., 2005). The majority of the results from the Upper Cretaceous forearc rocks are Campanian or Maastrichtian and hence enriched in detritus from the eastern plutonic zone.

In order to further evaluate the strength of the provenance tie between the Catalina Schist and the Peninsular Ranges, we have carried out ternary mixing calculations. We have identified
three distinctive components: (1) “lower Mesozoic wallrocks” represented by the Bedford Canyon, French Valley, Julian Schist, and Vallecitos flysch localities (Fig. 9A; Morgan et al., 2005); (2) “Early Cretaceous volcanics” that are represented by results from Lower Cretaceous sandstones and volcanics within the volcanic arc (Fig. 9B; Alsleben et al., 2005; D.L. Kimbrough, unpublished data); and (3) “Upper Cretaceous forearc” represented by Cenomanian-Maastrichtian strata distributed along the western margin of the Peninsular Ranges batholith (Fig. 9C; Mahoney et al., 2005). Cumulative probability plots for these three components are shown in Figure 9D along with equivalent curves representing the major units of the Catalina Schist.

We have linearly mixed these three components to obtain the best-fit to the detrital zircon age distributions from the Catalina Schist. Results for the amphibolite unit, epidote-amphibolite unit, epidote-blueschist unit, and lawsonite-blueschist and lower-grade units are shown in Figures 9E to 9H respectively. The relative proportions of the three end members required to produce these fits are shown in Figure 9I. As indicated in Figure 9I, the detrital zircon provenance signature of the earliest accreted material within the amphibolite unit of the Catalina Schist is best approximated by a 68:32 mixture of sediment derived predominantly from lower Mesozoic wallrocks and the Early Cretaceous volcanic arc, respectively. Such a high proportion of cratonally derived material within the amphibolite unit metasediments makes sense given the abundance of middle Proterozoic zircon, the aluminous and quartz-rich nature of the protolith, and the fact that pegmatites, metasediments, and mélangé sampled from the amphibolite unit tend to yield a relatively radiogenic Sr, Nd, and Pb isotopic signature (i.e., similar to evolved arc crust; Bebout and Barton, 1993, 2002; King et al., 2006, 2007).
The provenance signature of sediment shed from the Peninsular Ranges batholith shifted dramatically between early Aptian to early Cenomanian time (Mahoney et al., 2005). A parallel shift is recorded by the Catalina Schist (Fig. 4). In the case of the epidote-amphibolite unit, the relative proportions of lower Mesozoic wallrock to Early Cretaceous volcanic arc sediment diminishes to 44:56 (Fig. 9I). Compositionally, there is still a resolvable cratonal input in that epidote-amphibolite metagraywackes. A further shift is exhibited by the epidote-blueschist metagraywackes. Their age distribution is well modeled by a 20:58:22 mixture of lower Mesozoic wallrock, Early Cretaceous volcanic arc, and Upper Cretaceous forearc sediment respectively (Fig. 9I). The provenance shift is completed by the time the protolith of the lawsonite-blueschist and lower grade metagraywackes was deposited. The detrital zircon age distribution of the latter are well described by a 49:51 mixture of Early Cretaceous volcanic arc and Upper Cretaceous forearc sediment respectively (Fig. 9I).

**Implications of Pre-115 Ma High-P/T Tectonic Blocks within Low-Grade Catalina Schist**

Grove and Bebout (1995) reported $^{40}\text{Ar}/^{39}\text{Ar}$ results for two garnet-bearing, blueschist blocks within the lawsonite-blueschist unit of the Catalina Schist that indicated that the blocks predated 115 Ma peak-grade metamorphism within the amphibolite unit. Because the pronounced age gradients yielded by both samples reached 150 to 160 Ma at the highest temperatures of gas release, Grove and Bebout (1995) speculated that the garnet-bearing blueschist blocks formed during the Middle Jurassic and had the same tectonic significance as similar high-grade blocks present within the Franciscan Complex along the margin to the north (Wakabayashi, J., and Dumitru, 2007 and references cited therein) and within the western Baja terrane along the margin to the south (Baldwin and Harrison, 1989, 1992). The concordance of the 135 Ma Rb-Sr and K-Ar model ages from sample 330-4B (Fig. 5) indicate that $^{40}\text{Ar}/^{39}\text{Ar}$
results from it cannot be explained by excess $^{40}$Ar contamination. Petrographic observations from sample 330-4B indicate that the K-Ar and Rb-Sr model ages are most sensibly interpreted as mixed ages that resulted from partial replacement of 0.1-1 mm diameter Middle Jurassic phengite by much finer grained (< 50 µm) mid-Cretaceous phengite. This evidence for prior (Middle Jurassic?) subduction metamorphism along the margin is in good agreement with the geologic history of the PRB. Conversely, pre-115 Ma subduction along the segment of the margin underplated by the Catalina Schist makes it difficult to explain how mantle lithosphere comprising the over-riding hanging wall could have retained sufficient heat to produce amphibolite facies metamorphism within a newly formed subduction zone at 115 Ma.

**Subduction Erosion Model for the Formation of the Catalina Schist**

The results and geologic relationships described above lead us to conclude that the amphibolite, epidote-amphibolite, and possibly the epidote-blueschist units of the Catalina Schist of the Catalina Schist are unlikely to have formed within the same subduction zone in which the lawsonite-blueschist and lower-grade rocks of the Catalina Schist were also accreted and metamorphosed. Whereas studies of plate kinematics involving the Farallon, Kula, and Pacific plates are not definitive. They indicate that continuous subduction of comparatively old oceanic crust likely prevailed along the southwestern North American margin throughout the mid-Cretaceous (Engebretsen et al., 1986; Stock and Molnar, 1988). Accordingly, we favor a model in which the Catalina amphibolite and epidote-amphibolite units represent Early Cretaceous forearc strata and basement (Fig. 10A) that were underthrust and metamorphosed beneath a forearc thrust that was well-separated from the deeper subduction megathrust (Fig. 10B). A modern analogue for paired megathrusts depicted in Figure 10B is imaged by deep seismic reflection data from the northern Cascadia subduction zone (Calvert, 2004). Forearc thrusting on
this scale is capable of displacing forearc rocks to subcrustal positions and is consequently an important manifestation of subduction erosion.

In the Catalina Schist, high-P/T amphibolite facies metamorphism and partial melting may have occurred when a portion of the Early Cretaceous forearc (Fig. 10A) was underthrust to a position beneath the western margin of the northern Peninsular Ranges batholith between 122 to 115 Ma (Fig. 10B). This event coincided broadly with previously documented intra-arc thrusting within the adjacent PRB (Gastil et al., 1981; Todd et al., 1988; Silver and Chappell, 1988; Thompson and Girty, 1994; Busby et al., 1998; Johnson et al., 1999; Schmidt et al., 2002; Wetmore et al., 2002; Schmidt and Patterson, 2002; Wetmore, 2003; Busby, 2004). Shortening of the forearc could have been triggered by lateral expansion related to batholith emplacement coupled with collapse of thin, hot backarc crust that had separated the PRB from the southwestern craton margin in the Early Cretaceous.

Underthrusting of the Early Cretaceous forearc rocks to a stalled position beneath the western margin of the magmatic arc (Fig. 10B) explains the heat source required to produce amphibolite facies metamorphism and anatexis within the Catalina amphibolite unit. It also explains why the amphibolite unit experienced protracted greenschist facies metamorphic conditions after peak metamorphism occurred at 115 Ma. Greenschist facies, rather than blueschist facies overprinting is expected if the amphibolite unit was metamorphosed close to the magmatic arc and far-removed from the subduction zone.

Based upon detrital zircon results, the epidote-amphibolite unit was accreted after 113 ± 3 Ma or several m.y. after peak grade recrystallization in the overlying amphibolite unit (Fig. 6B). Accretion of the epidote-amphibolite unit created an imbricate thrust stack beneath the western margin of the PRB (Fig. 10C). Peak grade recrystallization of the epidote-amphibolite unit
occurred at 110 to 107 Ma based upon $^{40}$Ar/$^{39}$Ar hornblende results from the mafic gneiss immediately underlying the amphibolite unit (Fig. 6B). Just as in the case of the amphibolite unit, greenschist facies metamorphic conditions likely persisted within the epidote-amphibolite unit for up to 10 m.y. based upon $^{40}$Ar/$^{39}$Ar phengite results from epidote amphibolite unit metagraywackes that indicate cooling below 400 to 350°C was delayed until ca. 97 Ma (Fig. 6B).

Only fragmentary constraints are available for the metamorphic evolution of the epidote-blueschist unit. Previous work had indicated the rocks of the epidote-amphibolite unit and epidote blueschist unit were contiguous (the Catalina greenschist unit of Platt, 1976) and shared a common greenschist facies overprint (Platt, 1976; Sorensen, 1986). Our detrital zircon results from a single sample of epidote blueschist metasediment intercalated with clinozoisite-albite-bearing blueschists cast doubt upon the likelihood that the epidote-amphibolite and epidote-blueschist units were closely related prior to ca. 101 Ma. Results from this single sample indicate that the epidote blueschist rocks were accreted after $101 \pm 3$ Ma or roughly 6 m.y. after peak grade recrystallization of the epidote-amphibolite unit had occurred. The epidote-blueschist assemblages could reflect subcrustal accretion beneath the collapsing forearc at an intermediate position between the subduction zone and the arc. Greenschist facies overprinting of the epidote blueschist assemblages resulted as the rocks were underthrust beneath the epidote amphibolite and higher grade rocks of the Catalina Schist between 101 to 97 Ma.

The lawsonite-blueschist and lower-grade rocks of the Catalina Schist were accreted after $97 \pm 3$ Ma during a major Cenomanian pulse of syn-batholith erosion (Fig. 10D; Kimbrough et al., 2001). Evidence that accretion took place within the subduction zone is provided by the lawsonite-blueschist and lower-grade mineralogy and the fact that phengite $^{40}$Ar/$^{39}$Ar ages and
the youngest detrital zircon ages coincide, indicating that accretion, metamorphism, and cooling took place very rapidly (Fig. 6D). Merging of the higher grade units of the Catalina Schist with the subduction complex likely occurred by ca. 95 Ma and triggered tectonic denudation via low-angle extensional faulting which attenuated the crust and established the presently observed contact relationships within the Catalina Schist (e.g., Platt, 1986).

**Devolatilization and Fluid Flow within the Catalina Schist**

Recent studies (Bebout et al., 1999, 2007; Bebout, in press) have discussed varying extents of devolatilization in the tectonometamorphic units of the Catalina Schist within the context of varying prograde P-T paths experienced in a subduction zone setting (also see discussion in Grove and Bebout, 1995). Whereas the subduction erosion model (Fig. 10) does not affect the general conclusions of this previous work as they pertain to devolatilization histories, the sources and pathways for infiltrating fluids that metasomatized the higher grade rocks of Catalina Schist do require further consideration within the context of the subduction erosional model presented here. Bebout and Barton (1989) and Bebout (1991a,b) proposed that, in the amphibolite unit, the O isotope compositions in melange and other more permeable zones reflect infiltration by externally-derived aqueous fluids with δ18O inherited from prior equilibration with similar metasedimentary rocks but at lower temperatures (~300-600°C). The source of the fluids that ascended into the underplated (but still hot) amphibolite unit was primarily related to 400-550°C chlorite breakdown reactions in rocks equivalent to those within the lower grade units of the Catalina Schist (Bebout, 1991a,b; also see discussion in Grove and Bebout, 1995).

In the subduction erosion model (Fig. 10), fluids with appropriate lower-T metasedimentary signatures derived from the subducting slab are required to traverse a
significant thickness (10’s of kilometers) of mantle lithosphere in order to infiltrate the amphibolite unit at near peak-grade metamorphic conditions. We suggest that this could have occurred without substantial modification of the $\delta^{18}O$ of the fluids. Devolatilization of subducting oceanic crust and trench fill is considered capable of hydrating mantle lithosphere over broad regions (Peacock, 1993, Humphreys et al., 2003). Infiltration of aqueous fluids derived from the subduction zone likely began during, or prior to, the Middle Jurassic. This previous history was likely sufficient to thoroughly hydrate and isotopically re-equilibrate the intervening lithospheric mantle with lower-temperature, slab-derived fluids prior to the onset of Early Cretaceous subduction erosion of the forearc. Pervasive hydration of mantle lithosphere underlying the forearc region is consistent with observations, in modern forearcs, of extensive hydration in the hanging-wall leading to distinctive seismic velocity signatures (Bostock et al., 2002; Broucher et al., 2003) and forearc serpentinite diapers (Fryer et al., 1995). As continued subduction erosion brought the loci of forearc thrusting nearer to the active subduction thrust system (see Figs. 10C, 10D, and 10E), progressively lower-T rocks such as those in the epidote-amphibolite and epidote-blueschist rocks would similarly have been infiltrated by these fluids emanating from the still lower-T (lawsonite-blueschist) domains produced in the active subduction zone thrust.

**Relationship to Emplacement of the La Posta Tonalite-Trondjhemite-Granodiorite Suite**

At around 100 Ma, the locus of Peninsular Ranges batholith (PRB) magmatism shifted abruptly eastwards in conjunction with intrusion of the voluminous La Posta tonalite-trondjhemite-granodiorite plutonic suite (Fig. 2; Gastil et al., 1975; Silver & Chappell, 1988; Walawender et al., 1990; Tulloch and Kimbrough, 2003; Kimbrough and Grove, 2006). La Posta suite emplacement involved a sustained (ca. 98 to 92 Ma) magmatic flux of > 100
km$^3$/m.y./km strike length over more than 1200 km (Kimbrough and Grove, 2006). The high Sr, Ba, Sr/Y, Na$_2$O, Al$_2$O$_3$, and highly fractionated REE patterns exhibited by these plutons indicate deep, garnet-involved melting of a fundamentally mafic source region (Gromet and Silver, 1987). Oxygen and Rb-Sr isotopic measurements reported by Taylor (1986), Gromet and Silver (1987), Silver and Chappell (1988), Hill and Silver (1988), and Kistler et al. (2003) have revealed that the northern La Posta belt features elevated $\delta^{18}$O (9 to 12 ‰ whole rock equivalent) at intermediate initial $^{87}$Sr/$^{86}$Sr values (typically 0.704 to 0.708). The supracrustal input implied by these combined isotopic attributes cannot be accounted for by high-level assimilation of highly radiogenic, Early Mesozoic, cratonally-derived flysch host rocks (Shaw et al., 2003). These characteristics are more readily explained by partial melting and/or devolatilization of isotopically primitive low-grade metasedimentary and metavolcanic rocks and altered oceanic crust within the deep source region (Taylor, 1986; Gromet and Silver, 1987).

Emplacement of the La Posta belt plutons involved an eastern relocation of the locus of magmatism within the Peninsular Ranges over a strike length of 1200 km (Fig. 10D; Kimbrough and Grove, 2006). This event was the initial stage of an eastern sweep of magmatism into northern Mexico that occurred during the Late Cretaceous-early Tertiary (e.g., Silver and Chappell, 1988; McDowell et al., 2001; Staude and Barton, 2001; Henry et al., 2003). Subduction erosion (Fig. 10D) could have delivered Catalina Schist into close proximity to the La Posta source region between 100 to 95 Ma. The coincidence of crustal thickening of the former back arc extensional basin (compare Fig. 10A with 10D), focused asthenospheric corner flow (Fig. 10D), and massive devolatilization of Catalina Schist (Fig. 10D) may set up the the optimal conditions required to trigger the La Posta TTG flare-up (Kimbrough and Grove, 2006). Recent $\delta^{18}$O measurements performed by Kimbrough and Grove (2006) indicate that La Posta
plutons have high $\delta^{18}O$ only in the northern segment of the batholith, where it might have been underplated by the Catalina Schist (Fig. 2). This is a good indication that Catalina Schist subduction erosion was sufficiently important to modify the underlying causes and source region characteristics of the La Posta belt magmatism. An important implication is that substantial fractions of subducted forearc supracrustal material can be recycled via subduction erosion processes and incorporated into arc batholith magmas over a short (<10 Ma) time interval.

**Relationship to Laramide Underthrusting**

The accretion of the lowest-grade units of the Catalina Schist beneath the western Peninsular Ranges and the eastern relocation of the La Posta plutonic belt to the eastern Peninsular Ranges at 95 Ma can be viewed as important precursors to the Laramide craton-ward shift of arc magmatism and contractional deformation (Coney and Reynolds, 1977; Dickinson and Snyder, 1978). During the Laramide episode of Late Cretaceous-early Tertiary deformation, large tracts of southern California and southwestern Arizona were underplated by high-pressure/temperature Pelona and related schists, with no intervening lithospheric mantle preserved (Fig. 10E; Ehlig, 1968, 1981; Crowell, 1968, 1981; Yeats, 1968; Haxel and Dillon, 1978; Haxel et al., 2002; Burchfiel and Davis, 1981; Jacobson, 1983, 1990; Jacobson et al., 1988, 2002, 2007; Dillion et al., 1990; Malin et al., 1995; Wood and Saleeby, 1997; Saleeby, 2003; Grove, 2003a). Late Cretaceous eclogitic xenoliths have been recovered from kimberlite pipes as far east as northeastern Arizona (Usui et al., 2003).

The earliest recognized accretion of the distinctive Pelona and related schists occurred at 91 ± 1 Ma along the southwesternmost tip of the Sierra Nevada batholith (Rand Schist within the San Emigdio Mountains; Saleeby, 2003; Grove et al., 2003a). Lawsonite-blueschist and lower grade Catalina Schist represent nearly equivalent material underplated at a slightly older time
Thermochronology and detrital zircon results obtained from the northwest-southeast trending belt of schist exposures have revealed that schist underplating associated with the Laramide event was widespread by 80 to 70 Ma beneath the mid-Cretaceous arc (Jacobson, 1990; Jacobson et al., 2000; Barth et al., 2003; Grove et al., 2003a). By 70 to 60 Ma, cratonal rocks were being underplated by schist at positions well east of the medial Cretaceous arc (Grove et al., 2003a; Usui et al., 2003; Jacobson et al., 2007).

Laramide shallow subduction processes tectonically removed the deep lithospheric mantle roots of the La Posta belt within the northern Peninsular Ranges batholith between 80 and 65 Ma (Fig. 10E). Receiver function seismic studies indicate that the deep crust and lithospheric mantle roots no longer exist beneath within the northeastern PRB (Ichinose et al., 1996; Lewis et al., 2001). Whereas the PRB exhibited predominately syn-to late-batholithic cooling in its southern extent (>85 Ma; e.g., Ortega Rivera, 2003), the northeastern segment of the batholith was further characterized by a delayed and very significant pulse of rapid cooling between 80 and 65 Ma (Krummenacher et al., 1975; George and Dokka, 1994; Lovera et al., 1999; Grove et al., 2003b). Lovera et al. (1999) and Grove et al. (2003b) have attributed this cooling to denudation related to the removal of lithospheric mantle and the underplating of schist during Laramide shallow subduction. Holk et al. (2006) reported that Laramide-age deformation within the northern Peninsular Ranges was associated with infiltration of high $\delta^D$ and $\delta^{18}O$ fluids that are most readily explained by devolatilization of subducted oceanic crust or underplated volcanogenic sediments. The impact of the Laramide Orogeny on the northern Peninsular Ranges appears to have contrasted significantly with its effect upon the southern Sierra Nevada batholit, where an inflection in the Laramide subduction zone (Pickett and Saleeby, 1993; Malin
et al., 1995; Saleeby, 2003) apparently allowed the deep crust and upper mantle to be preserved until much more recently (Ducea and Saleeby, 1996, 1998; Zandt et al., 2004).

**CONCLUSIONS**

(1) Major tectonometamorphic units of the Catalina Schist where successively accreted over a 15 to 20 m.y. interval beginning with the amphibolite unit at ca. 120 to 115 Ma and concluding with the lawsonite-blueschist and lower-grade lithologies by 97 to 95 Ma.

(2) The amphibolite unit resided at high temperatures for a protracted period of time (10 to 15 m.y.) whereas the lawsonite-blueschist and lower grade units most likely formed near or within the subduction zone were deposited, accreted, metamorphosed, and cooled over a time interval too brief to be resolved by the methods employed in this study (< 3 m.y.).

(3) The provenance of the Catalina Schist metasedimentary rocks shifted as a function of time of accretion (and now, metamorphic grade). Metagraywackes from the earliest-accreted amphibolite unit were derived from an early Aptian sediment that was apparently originated from erosion of Late Triassic-Jurassic flysch wallrocks and Early Cretaceous volcanics of the Peninsular Ranges batholith. Successively younger accreted materials became enriched with Early Cretaceous plutonic zircon from the Peninsular Ranges. The last-accreted lawsonite-blueschist and lower-grade rocks were derived from Turonian sediment with a detrital zircon provenance virtually identical to similarly-aged sediment within the Peninsular Ranges forearc.

(4) The Catalina Schist likely does not represent a synchronously formed, inverted metamorphic aureole formed during nascent subduction. The highest grade portions of the complex appear to have formed by a subduction erosion process in which portions of the forearc were underthrust beneath the western margin of the Peninsular Ranges batholith at ca. 122 to 115
Ma. Progressive subduction erosion of the forearc by continued underthrusting in the forearc region ultimately juxtaposed the higher grade units of the Catalina Schist with the subduction complex by 97 to 95 Ma. Only the lawsonite-blueschist and lower grade rocks are considered to have originated along the dominant subduction zone thrust within a high-P/T thermal regime characteristic of such settings.

(5) The accretion of the Catalina Schist marked the initial stage of shallowing subduction and inboard migration of magmatism and sediment underplating that culminated in the Late Cretaceous-early Tertiary Laramide Orogeny. The sedimentary protolith of the youngest Catalina schist units are nearly identical in age and provenance to the oldest representative of the eugeoclinal schists emplaced beneath the southwesternmost Sierra Nevada batholith at ca. 92 Ma.

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FIGURES

**Figure 1** Geologic setting of the Catalina Schist. Location map provided in upper left shows the distribution of high-pressure/temperature subduction complexes and Cretaceous-tertiary
batholiths in southwestern North America. The offshore distribution of subduction complexes is estimated primarily from Crouch and Suppe (1993), Bohannon and Geist (1998), Sedlock, 1988a,b, Bonini and Baldwin (1998), and Fletcher et al., (2006). The box outlines the location of southern California continental borderland. The simplified map of borderland shown in the lower right is based upon Bohannon and Geist (1998) while cross section Y-Y’ is after Crouch and Suppe (1993). Symbols P & P* denote formerly contiguous rocks that dextrally sheared and rotated clockwise during Middle Miocene rifting. Box shows location of Catalina Island. The geologic map and cross section X-X’ of Catalina Island in the upper left are after Platt (1976). The distribution of lawsonite-albite and actinolite-albite rocks is from Altheim et al. (1999). Note that epidote-blueschist and epidote-amphibolite rocks are distinguished from Platt’s (1976) greenschist unit. A coherent km-scale mass of epidote amphibolite facies metagabbro that directly underlies the amphibolite unit was included by Platt (1976) within his amphibolite unit.

Figure 2  Geologic map of the Peninsular Ranges batholith modified after Gastil et al., (1975) and Kimbrough et al., (2001). Generalized stratigraphic relationships of forearc region in the northern Vizcaino peninsula modified after Kimbrough et al. (2001), Kimbrough and Moore (2003) and references cited within these papers. Simplified stratigraphy of northern Santa Ana Mountains is based upon references provided within Lovera et al. (1999). Western limit of forearc is the restored pre-middle Miocene position based upon Bohanon and Parson (1998) reconstruction and our interpretation of areo magnetic data presented by Langenheim and Jachens (2003).

Figure 3  (A) Measured zircon U-Pb age vs. Th/U of Cretaceous forearc sedimentary rocks of the Peninsular Ranges batholith. Data from Mahoney et al. (2005) demonstrates that virtually all detrital igneous zircons derived from the Peninsular Ranges have Th/U > 0.1 (see also Williams
and Claesson, 1987). (B) Equivalent plot for zircon results from the Catalina Schist amphibolite unit. Based upon cathodoluminescence imaging and other criteria, we regard a Th/U = 0.1 as a meaningful cutoff to distinguish analyses in which the sputter pit overlapped metamorphic zircon growth (open symbols). The filled symbols represent detrital grains that we consider to be largely unaffected by metamorphic zircon growth (see also Kröner et al., 1994; and Rubatto, 2002). Zircon results with Th/U < 0.1 have been excluded from all subsequent plots and calculations.

**Figure 4** Relative probability plots of detrital zircon U-Pb age distributions from major units of the Catalina Schist on Catalina Island. (A) amphibolite (B) epidote-amphibolite (C) epidote-blueschist (D) lawsonite-blueschist (E) actinolite-albite (F) lawsonite-albite. Note that we use a split horizontal axis at 300 Ma and that relative probability plots between 300-3000 Ma have a 2x scaling factor to improve resolution of the overall age distribution, (G) True scale cumulative probability spectra for all units.

**Figure 5** (A) Garnet-bearing blueschist block in mélangé matrix of the lawsonite blueschist unit. (B) Rb-Sr results for phengite, Na-amphibole, and whole rock. A statistically meaningful isochron is not defined. Phengite and Na-amphibole define a model isochron (135 Ma) that is identical to the total gas $^{40}\text{Ar}/^{39}\text{Ar}$ age yielded by this sample.

**Figure 6** Estimated temperature-time histories for major units of the Catalina Schist. (A) amphibolite (B) epidote-amphibolite (C) epidote-blueschist (D) lawsonite-blueschist and lower grade units of the Catalina Schist. Yellow circles represent measured detrital zircon U-Pb ages. Maximum bounds upon the depositional age of the sedimentary protolith represent average and standard deviation of U-Pb ages of five youngest zircons with Th/U > 0.1 (15 youngest zircons for composite of the three lowest grade units). Data sources for other thermochronometry
include Suppe and Armstrong (1972), Mattinson (1986), Grove and Bebout (1995), and Anczkiewicz et al. (2004). Note that we have varied the bulk closure temperatures assigned to micas from 400°C for coarse grained muscovites within the high-grade units to 350°C for finer grained, less retentive white mica within lawsonite-albite rocks.

**Figure 7** Whole rock Pb isotopic data of King et al. (2007) for: (A) Lawsonite-albite; (B) Lawsonite-blueschist; and (C) Amphibolite unit mélange matrix of the Catalina Schist. Also shown are whole rock Pb isotopic data for northern PRB plutons (red circles; Kistler et al., 2003) and equivalent data for southeastern Arizona (orange circles; Wooden and Miller, 1990). The yellow field represents the generalized distribution of whole rock Pb yielded by the Transverse Ranges and Mojave province (based upon Barth et al., 1995 and unpublished data from J. Wooden, D. Coleman, and A. Barth).

**Figure 8** Relative probability plots (200-2200 Ma) of detrital zircon U-Pb ages from: (A) All results from the Catalina Schist. (B) Representative early Mesozoic flysch wallrocks of the northern PRB (Morgan et al. (2005; data originates from localities 1-4 in Figure 2. (C) Representation of the southwestern North American detrital zircon provenance signature as represented by Late Miocene-Recent sediments of the Colorado River system (D. Kimbrough and M. Grove, unpublished data). Yellow bands denote the primary Middle Proterozoic crystallization age maxima contributed by cratonic basement and supracrustal rocks of southwestern North America. (D) All detrital zircon U-Pb age results from the lawsonite-blueschist and lower grade units of the Catalina Schist; (E) Representative Late Cretaceous detrital zircon data from the forearc of the northern Peninsular Ranges batholith (Mahoney et al., 2006). (F) Pluton U-Pb zircon crystallization ages from the northern PRB (Silver and Chappell, 1988; Walawender et al., 1990; Kistler et al., 2003; D. Kimbrough, unpublished data).
Figure 9  Relative probability plots (0-2000 Ma) of detrital zircon U-Pb age results used as end members in ternary mixing calculations: (A) the early Mesozoic flysch wallrocks of the northern PRB from Morgan et al. (2005). (B) Early Cretaceous volcanic sandstones (Alslében et al., 2005) and volcanic rocks (D. Kimbrough, unpublished data) of the northern PRB. (C) Late Cretaceous forearc strata of the northern PRB (Mahoney et al., 2005). D. Cumulative probability plots of the three endmembers. Best-fit ternary-mixing results for: (E) amphibolite unit, (F) epidote-amphibolite unit, (G) epidote-blueschist unit, and (H) lawsonite-blueschist (and lower grade) units of the Catalina Schist. (I) Ternary diagram showing best-fit solutions. Note the general trend away from Peninsular Ranges wallrock and volcanic zircon signature exhibited by the oldest accreted units toward pluton-dominated zircon provenance that characterizes the youngest rocks within the Catalina Schist.

Figure 10  Subduction erosion model for the formation of the Catalina Schist. (A) Early Cretaceous geometry of convergent margin of northern PRB. The backarc extension and sedimentation shown are inferred from geologic relationships within east-central PRB and formerly adjacent mainland Mexico. Detrital zircon results from Early Cretaceous volcanic sediments (Alslében et al., 2005) require close proximity to southwestern North America. (B) Initial underthrusting of Early Cretaceous forearc beneath northwestern Peninsular Ranges and amphibolite facies metamorphism and anatexis at 115 Ma. Amphibolite unit stalls beneath western batholith at a position well-separated from active subduction zone. Evidence for intra-arc shortening at this time is detailed within the text. (C) Underthrusting of epidote-amphibolite unit during progressive subduction erosion of the Early Cretaceous forearc beneath northwestern PRB. Shortening in the backarc region preconditions the crust for deep melting (> 40 km) to generate La Posta magmatism at 95 Ma. (D) Accretion of the lawsonite-blueschist and lower
grade units of the Catalina Schist between at 95 Ma occurs concomitant with major pulse of
denudation and erosion within PRB (e.g., Kimbrough et al., 2001). Deep, garnet-involved,
melting to generate La Posta magmatism is facilitated by overthickened arc-ophiolite basement
in the former backarc, focused asthenospheric counter flow caused by initiation of flat subduction
and eastern relocation of magmatic arc, and massive devolatilization of deep underplated
Catalina Schist within the subduction channel. (E) Laramide flat subduction tectonically erodes
deep crustal root beneath eastern PRB between 80-65 Ma, causing renewed deep exhumation of
eastern Peninsular Ranges batholith (Grove et al., 2003b).
Figure 2
Figure 3

(A) Peninsular Ranges Forearc (N = 1527)

(B) Catalina Schist Amphibolite Unit (N = 169)

- **Th/U**
- **Zircon U-Pb Age (Ma)**

- **used in calculations**
- **omitted from consideration**
Figure 4

(A) Amphibolite (N = 129)

(B) Epidote Amphibolite (N = 122)

(C) Epidote Blueschist (N = 30)

(D) Lawsonite Blueschist (N = 164)

(E) Actinolite Albite (n = 59)

(F) Lawsonite Albite (n = 82)

(G) Cumulative Probability vs. Detrital Zircon U-Pb Age (Ma)
Figure 5

(A) A photograph of a rocky landscape.

(B) A diagram showing the relationship between Rb/Sr and Sr/Sr.

- Phengite
- Whole rock
- Na-amphibole

The dashed line represents a 135 Ma Reference Isochron.
Figure 6

(A) Amphibolite unit
- Maximum depositional age: 122 ± 3 Ma

(B) Epidote-amphibolite unit
- Maximum depositional age: 113 ± 3 Ma

(C) Epidote-Blueschist
- Maximum depositional age: 100 ± 3 Ma

(D) Lawsonite Blueschist & lower-grade units
- Maximum depositional age: 97 ± 3 Ma
Figure 7

(A) SE Arizona Proterozoic
- PRB granitoids
- lawsonite-albite unit melange

(B) SE Arizona Proterozoic
- lawsonite-blueschist unit melange

(C) SE Arizona Proterozoic
- amphibolite unit melange
Figure 8

(A) Catalina Schist (all units)

(B) northern PRB Early Mesozoic wallrocks

(C) Colorado River (5 to 0 Ma)

(D) Catalina Schist lawsonite- blueschist & lower grade

(E) northern PRB Late K forearc

(F) northern PRB pluton crystallization ages
Figure 9

(A) PRB Early Mz wallrocks

(B) PRB Early K volcanic arc

(C) PRB Late K forearc

(D) Peninsular Ranges batholith

(E) Amphibolite facies: measured vs. model fit

(F) Epidote amphibolite facies: measured vs. model fit

(G) Epidote blueschist facies: measured vs. model fit

(H) Lawsonite blueschist & lower grade facies: measured vs. model fit

(I) Early Mz Wallrocks

- Red: amphibolite
- Green: epidote amphibolite
- Cyan: epidote blueschist
- Blue: lawsonite blueschist & lower grade facies

Cumulative Probability vs. Zircon U-Pb Age (Ma)

Relative Probability

Zircon U-Pb Age (Ma)
Figure 10