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Jurassic peraluminous gneissic granites in axial zone of Peninsular Ranges, southern California

S.E. Shaw
GEMOC (National Key Centre for the Geochemical Evolution and Metallogeny of Continents), Department of Earth and Planetary Sciences, Macquarie University, NSW 2109, Australia

V.R. Todd
Palomar College, San Marcos, California 92069, USA

M. Grove
Department of Earth and Space Sciences, University of California, Los Angeles, California 90095, USA

ABSTRACT

The Peninsular Ranges batholith of southern California and Baja California, México, is well recognized as a prime example of an I-type Cretaceous batholith. Often overlooked, however, is a volumetrically significant amount of pre-Cretaceous gneissic granite in the axial zone of the batholith. New U-Pb zircon age data confirm that the metaluminous and peraluminous plutonic bodies were emplaced during the middle Jurassic. Also reported in this paper is a Jurassic U-Pb age for a metaluminous (I-type) tonalite-quartz diorite pluton that is spatially related to the peraluminous suites. This result suggests that other unrecognized Jurassic I-type plutons may also be present in the batholith.

Within San Diego County, Todd and Shaw (1985) recognized and mapped two suites of strongly deformed gneissic granites and migmatites. One is peraluminous (Harper Creek suite) while the other is transitional between metaluminous and peraluminous (Cuyamaca Reservoir suite). These rocks bear a striking resemblance to deformed and, in places, migmatitic, peraluminous (S-type) examples from the Lachlan fold belt and New England batholith of eastern Australia. The gneissic granite suites are known to extend north along the axial zone of the Peninsular Ranges batholith and cover an area at least 45 km wide by 150 km long. To the south, rocks of similar type are known to extend into Baja California, México, for at least 300 km. Chemical and isotopic studies of these Jurassic suites confirm that they meet the criteria necessary to define them as S-type and transitional I- to S-type, respectively. However, unlike the majority of Lachlan fold belt S-type granites that are high level and often associated with their volcanic equivalents, the Peninsular Ranges batholith suites were emplaced at much deeper levels, possibly as much as 11–16 km. The Harper Creek suite, of S-type gneissic granodiorite-tonalite plutons and associated Stephenson Peak migmatitic schist and gneiss facies, is strongly peraluminous and contains biotite, cordierite, sillimanite, abundant graphite, and ilmenite. It has elevated $\delta^{18}O$ up to +20 per mil, initial $^{87}Sr/^{86}Sr$ ratios (Sr_i) to 0.713, a high aluminum saturation index, and Na_2O/K_2O ratios that overlap those of the Lachlan fold belt S-type granites. The Cuyamaca Reservoir suite contains gneissic granodiorite-tonalite plutons, transitional between metaluminous and moderately peraluminous (I- to S-type), containing reduced biotite, subaluminous amphibole, orthopyroxene, titanite

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and ilmenite. It has values of $\delta^{18}O$ and Sr_i greater than the Cretaceous I-type granites but less than the Harper Creek suite. Leucosome melt phase accumulation from the Julian Schist diatexites to produce a restite-rich magma is considered the most likely origin for the Harper Creek suite. For the Cuyamaca Reservoir suite, possible source components include young mantle-derived magma, metaigneous and metasedimentary rocks formed in an arc environment, and evolved basinal fill metasedimentary rocks. The evolved metasedimentary component may be of Julian type.

The Harper Creek and Cuyamaca Reservoir suites comprise deformed, steep-walled, north-northwest–trending bodies up to 20 km long and with length-to-width ratios of 4:1. Textures range from strongly foliated to gneissic or mylonitic. Internal foliation that strikes parallel to the long dimension of the bodies and dips steeply to the east is defined by alignment of relict magmatic feldspar and quartz grains and recrystallized aggregates of quartz and biotite. The concordance of magmatic and subsolidus foliations in the Jurassic plutons and the continuity of these structures with regionally developed metamorphic fabrics in their wallrocks indicate that magmatic foliation was overprinted by high-temperature, post-magmatic solid-state foliation. Foliation in the Harper Creek and Cuyamaca Reservoir suites is concordant with the axial planes of outcrop-scale isoclinal folds, and map patterns suggest that the plutons underwent regional-scale isoclinal folding. Their fabric probably formed during multiple episodes of synintrusive deformation that began at least by the Late Jurassic and culminated by the middle Cretaceous.

Keywords: granite; Jurassic; migmatite; Peninsular Ranges; peraluminous; restite; S-type.

INTRODUCTION

The Peninsular Ranges batholith of southern California, U.S.A., and Baja California, México, is traditionally viewed as having evolved wholly within the Cretaceous (Silver and Chappell, 1988). Although the great volume of the Cretaceous plutonic rocks has obscured the earlier Mesozoic history of the region, the recognition of gneissic granites of Jurassic age (Thomson and Girty, 1994) in the southern California segment of the batholith extends the history of igneous activity significantly and increases the similarity of the Peninsular Ranges batholith to the Sierra Nevada batholith and other arc terranes along the western margin of North America.

Palinspastic restoration of Neogene dextral displacement on the San Andreas fault system places the Peninsular Ranges batholith about 300 km south of its present position, opposite the Sonoran Desert region (Gastil, 1993; Fackler-Adams et al., 1997). Whereas an impressive body of data describes Jurassic volcanic and plutonic rocks of the Mojave-Sonoran arc (Busby-Spera et al., 1990; Staude and Barton, 2001), relatively little is known about contemporaneous magmatic activity to the west, oceanward of the continental margin. However, Jurassic igneous activity and metamorphism has been reported in the Sierra San Pedro Martir area, Baja California (Schmidt and Paterson, 2002). Data presented in this paper show that a substantial part of the central Peninsular Ranges batholith, extending from at least 33° 45’ N in southern California and possibly as far south as 28° N in Baja California, is composed of a belt of Jurassic gneissic granites (Todd et al., l99la; Thomson and Girty, 1994) (Fig. 1).

Broad regions of gneissic granitoids (Stonewall Peak granodiorite) and “mixed rocks” that were believed to be of potential Jurassic age were originally mapped by Everhart (1951) and Merriam (1958) within the axial zone of the Peninsular Ranges batholith near 33°N. Todd and Shaw (1979, 1985) studied these heterogeneous rocks in more detail and were able to recognize two compositionally distinct suites. Following the original definition of I-type and S-type granites of Chappell and White (l974) and White and Chappell (1977) for rocks in southeastern Australia, and noting the similarity, Todd and Shaw (l985) described an S-type Harper Creek suite and a transitional I-type to S-type Cuyamaca Reservoir suite. Although the application of these terms to plutons in the Peninsular Ranges batholith was questioned by White et al. (1986), the gneissic granites more than meet the criteria of Chappell and White (1974, 1992) and White and Chappell (1988) for I- and S-type granites from southeastern Australia. The distinctive criteria used to define S-types in the Lachlan fold belt (and which also characterize the Jurassic peraluminous gneissic granites of the Peninsular Ranges batholith) include: molecular $Al_2O_3/(Na_2O+K_2O+CaO) >1.1$, equivalent to CIPW normative corundum >1%; relatively high SiO_2 values (>63%); less regular inter-element variations that I-types; reduced nature of the granites as reflected by ilmenite rather than magnetite and reduced biotite with red-brown pleochroism; cordierite present in less deformed granites; and elevated $Sr_i>0.708$ and $\delta^{18}O>+10$ per mil.

The word “granite” as used in this paper is a general term to describe coarse-grained rocks of igneous origin consisting mainly of quartz and feldspars. The term “gneissic granite” is used to

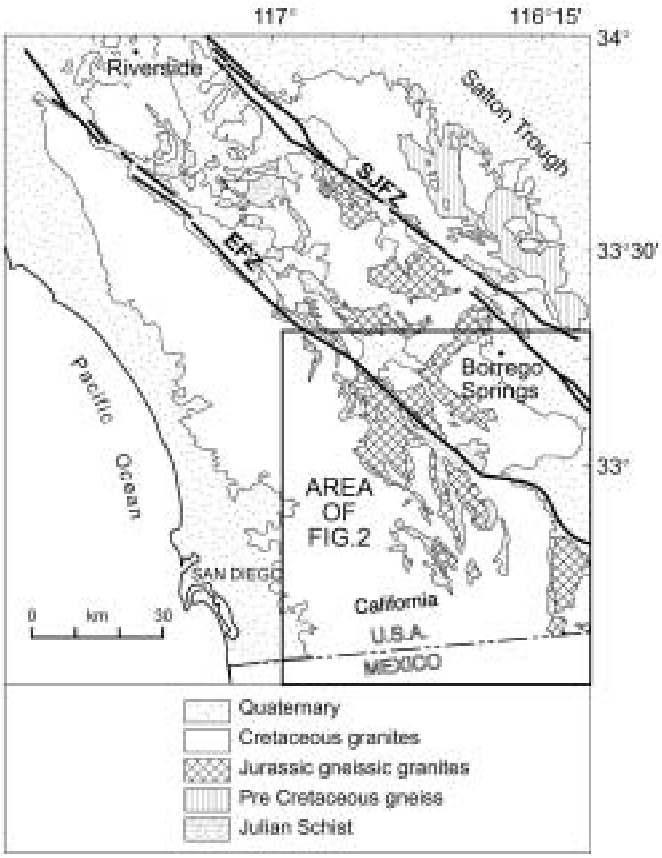

Figure 1. Map of the northern Peninsular Ranges batholith (after Jachens et al., 1986; Todd, 1978, 1994a, 1994b; Todd et al., 1988). Neogene San Andreas fault system: SJFZ—San Jacinto fault zone; EFZ—Elsinore fault zone. Jurassic gneissic granites are in part mylonitic and include migmatite and gneiss facies. Pre-Cretaceous gneiss includes orthogneiss (Harper Creek–type?) and anatexites, mylonites, and Paleozoic metasedimentary rocks.

describe metamorphosed "granite," although compositionally, the rocks referred to in this paper are mainly granodiorite and tonalite. The term "restite" is used to describe residuum or solid material in equilibrium with the melt that remains after a melting event.

The Jurassic Peninsular Ranges batholith gneissic granite belt is essentially parallel to, but lies ~150 km west of, the intracontinental Jurassic-Triassic Mojave-Sonoran arc. An estimate of the outcrop area of the Peninsular Ranges batholith Jurassic gneissic granites from Figure 1 is ~1155 km^2. While this represents a small proportion of the more than 1200-km-long, mainly Cretaceous batholith, the area is comparable with Australian examples of mixed I-S-type batholiths in the Lachlan fold belt: Berridale (1670 km^2), Murrumbidgee (1470 km^2), and Kosciusko (4000 km^2). The Cooma S-type gneiss complex in the Lachlan fold belt is, however, minor at 14 km^2 (areas from Chappell et al., 1991).

This paper discusses in more detail the field relationships, mineralogy, U-Pb zircon geochronology, and geochemistry of the Jurassic Harper Creek and Cuyamaca Reservoir suites. Descriptions of the metaluminous tonalite–quartz diorite plutons of the East Mesa suite that are also partly Jurassic in age (this paper) are given in Todd et al. (this volume).

GEOLOGICAL SETTING

The north-northwest–trending Peninsular Ranges batholith exhibits a strong transverse asymmetry (DePaolo, 1981; Gastil, 1983; Taylor, 1986; Gromet and Silver, 1987; Silver and Chappell, 1988; Todd et al., 1988; Walawender et al., 1990). This asymmetry is considered to reflect the change in granite source regions from oceanic lithosphere on the west to continental lithosphere on the east. The resultant west-to-east lithologic variations within plutons and prebatholithic wallrocks, which were noted in early studies, have been confirmed by subsequent geochemical, isotopic, and geophysical studies (Taylor, 1986; Shaw et al., 1986; Silver and Chappell, 1988; Walawender et al., 1990; Jachens et al., 1986, 1991; Wooden et al., 1997; Premo et al., 1998). Todd and Shaw (1985, their fig. 2) summarized significant boundaries and gradients that roughly divide the batholith into western and eastern zones. Prebatholithic rocks in the western zone of the Peninsular Ranges batholith are chiefly Mesozoic volcanic and sedimentary rocks, whereas those in the eastern zone consist of sedimentary and lesser volcanic rocks of Mesozoic, Paleozoic, and Late Proterozoic age (Gastil, 1993). The batholith in the western zone comprises relatively magnetic, dense plutonic and volcanic rocks and in the eastern zone, less dense, virtually nonmagnetic plutons and metasedimentary wallrocks. Jachens et al. (1986) modeled the steep gravity-magnetic gradient that separates the two zones as the expression of a tectonic boundary dipping 45–60° to the east and extending to depths of at least 10–12 km. In San Diego County, this geophysical boundary coincides with the western limit of Jurassic granites and is shown in Figure 2 as the boundary PSGR (pseudogravity gradient). Jachens (1992, personal commun.) defines the pseudogravity gradient as the "mathematical transformation of magnetic field data into pseudogravity anomalies, effectively converting the magnetic field to the gravity field that would be produced if all magnetic material were replaced by proportionally dense material." The $\delta^{18}O$ step of Taylor and Silver (1978) corresponds closely with the PSGR boundary (Fig. 2). In a depth subdivision of the Peninsular Ranges batholith (Gastil, 1975), the boundary of intermediate Zone B with deepest Zone C lies approximately 5 km to the west of the PSGR boundary. In the threefold subdivision of the Peninsular Ranges batholith based on rare earth element (REE) variation (Gromet and Silver, 1987), the boundary between western and central regions lies ~5 km to the east of the PSGR boundary (Fig. 2).

On the basis of these discontinuities, Todd and Shaw (1985) suggested they are indicative of a fundamental change in the Mesozoic crust, possibly a suspect terrane boundary formed by the accretion of a western volcanic arc and the continental margin to the east. Schmidt and Paterson (2002) describe the boundary as an example of crustal transition between juxtaposed oceanic and continental floored arcs, with Jurassic-Cretaceous deforma-

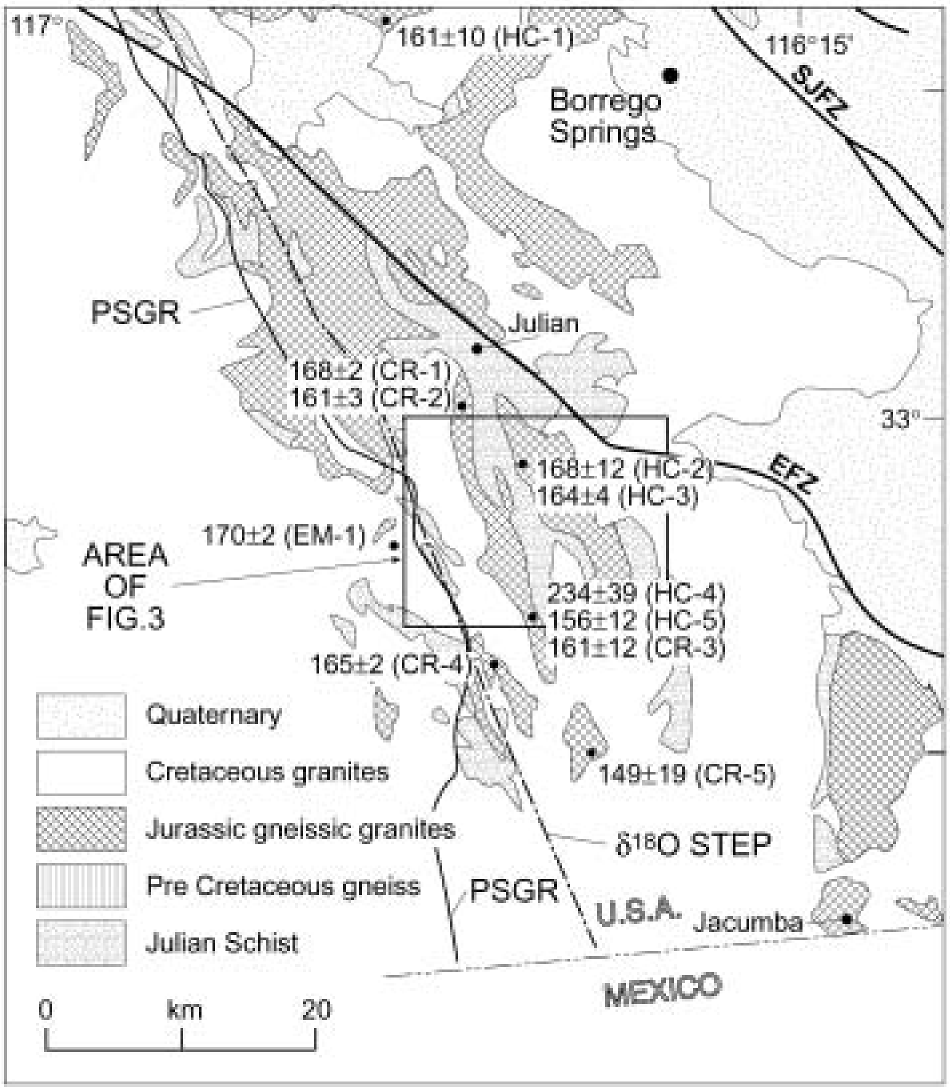

Figure 2. Enlargement of southern part of Figure 1 shows approximate sample locations for dated Jurassic plutons (see section on age of Jurassic granites). Locations have been combined in cases where samples were collected within a distance of approximately 1–2 km. HC—Harper Creek suite: HC-1, this paper; HC-2, Todd et al. (1991b, Rb/Sr whole-rock age); HC-3, this paper; HC-4 and HC-5, Thomson and Girty (1994). CR—Cuyamaca Reservoir suite: CR-1 and CR-2, this paper; CR-3, Thomson and Girty (1994); CR-4, this paper; CR-5, Thomson and Girty (1994). EM—East Mesa suite: EM-1, this paper. Unless specified, all are zircon $^{206}Pb/^{238}U$ ages. Also shown are SJFZ (San Jacinto fault zone) and EFZ (Elsinore fault zone) (as in Fig. 1), $\delta^{18}O$ step of Taylor and Silver (1978) and gravity-magnetic gradient of Jachens et al. (1986, 1991) shown here as pseudogravity gradient boundary (PSGR). Gravity and magnetic gradients coincide closely with western edge of terrain of Jurassic granites and Julian Schist and I-S line of Todd and Shaw (1985).

tion evident for at least 800 km along the axis of the Peninsular Ranges batholith.

The oldest plutonic rocks recognized in the Peninsular Ranges batholith are highly deformed Middle and Late Jurassic metaluminous to peraluminous granodiorites and tonalites that are exposed in the central zone of the Peninsular Ranges batholith (Todd et al., 1991a, 1991b; Thomson and Girty, 1994). Plutonic rocks of possible Triassic age (Girty et al., 1994) have also been reported (see section on age below). By the Late Jurassic, these granites and their Mesozoic and Paleozoic wallrocks formed a belt at least 45 km wide along the southwestern margin of North America (Figs. 1 and 2). About 40 km west of the belt of Middle and Late Jurassic plutons are scattered exposures of Late Jurassic to Early Cretaceous volcaniclastic-volcanic island arc-type rocks (Balch et al., 1984; Anderson, 1991; Wetmore et al., 2001). The Jurassic plutons were subsequently intruded on the west by voluminous Early Cretaceous I-type plutons (ca. 120–l05 Ma, Silver and Chappell, l988; Premo et al., 1998). Contacts between the Cretaceous and Jurassic plutons are broadly concordant and in detail suggest mylonitization and localized incipient melting of the Jurassic granites during Cretaceous intrusion. Between ca. l05 and 89 Ma, with a major pulse centered on 95 Ma (Walawender et al., l990; Kimbrough et al., 2001), large hornblende-bearing trondhjemite to garnetiferous two-mica monzogranite plutons (La Posta–type) intruded the eastern part of the Jurassic belt. Unlike the Early Cretaceous intrusions, the middle- to Late Cretaceous plutons are largely discordant to the belt, crosscutting it in the southern part of the study area (Fig. 3) and fragmenting or obliterating large parts of the belt on its eastern side.

The Peninsular Ranges block, deeply eroded and peneplaned by Eocene time (Krummenacher et al., 1975), underwent renewed uplift and apparent westward tilting as a result of the

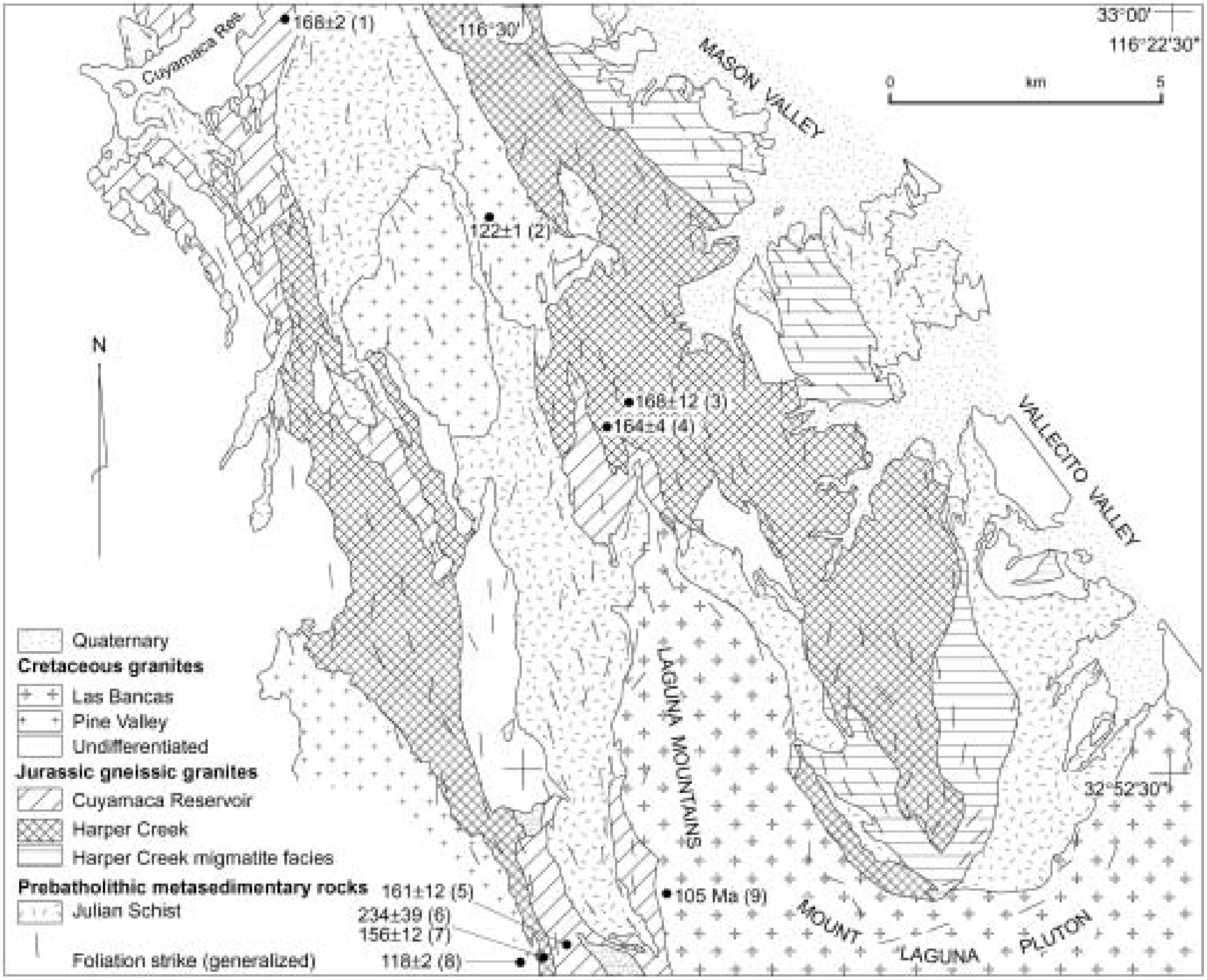

Figure 3. Generalized geologic map (after Todd, 1994b) of part of Peninsular Ranges batholith outlined in Figure 2. Trend lines indicate strike orientation of steep (60–90°), mainly eastward-dipping foliation in Jurassic gneissic granites, Cretaceous plutons, and metamorphic screens. Entire area is within Cretaceous Cuyamaca–Laguna Mountains shear zone of Todd et al. (1988). References for numbered age locations are: (1) this paper; (2) D.L. Kimbrough, 1994, personal commun.; (3) Todd et al. (1991b); 4) this paper; (5), (6), and (7) Thomson and Girty (1994); (8) D.L. Kimbrough, 1994, personal commun.; (9) L.T. Silver, 1979, personal commun.

Neogene development of the Gulf of California San Andreas rift-transform system. Progressively deeper levels of erosion across the batholith have produced a depth profile from volcanic levels on the west to mid-crustal depths in the central to eastern Peninsular Ranges batholith (see Grove et al., this volume).

PREBATHOLITHIC FRAMEWORK

The Jurassic granites were emplaced into predominantly metasedimentary country rocks, which are now represented as a large number of steeply dipping, northwest-trending screens (Figs. 1 and 2). The granites appear to have intruded as a north-trending belt across two or more northwest-trending prebatholithic wallrock terrains. Flyschlike metasedimentary wallrocks in the western part of the Jurassic intrusive belt are assigned to the early Mesozoic Julian Schist of Hudson (l922), whereas wallrocks in the central part of the intrusive belt, although lithologically similar to the Julian Schist, differ in that they contain ubiquitous thin marble interbeds as well as a greater proportion of quartz-rich metasandstone (Engel and Schultejann, 1984; Gastil and Miller, 1984; Todd et al., 1987, 1988). Todd et al. (1987) informally named these rocks the Jacumba sequence and tentatively assigned them a Mesozoic and/or Late Paleozoic age.

Julian Schist

The Julian Schist comprises a diverse assemblage of fine-grained, quartzo-feldspathic micaceous metasedimentary rocks, grading from semi-pelitic to feldspathic metaquartzite (metapelites to metapsammites). The predominant schists and quartzites are interlayered with smaller amounts of amphibolite and mafic schist, calc-silicate quartzite and gneiss, metaconglomerate, and rare marble and talc schist (Berggreen and Walawender, 1977; Todd and Shaw, 1979; Germinario, 1993; Grove, 1987). The overall fine grain size, relict thin bedding, and well-preserved turbidite structures in the larger western screens of the Julian Schist indicate that the sediments were deposited on submarine fans or fan complexes (Detterman, 1984; Grove, 1987; Germinario, 1993; Reed, 1993; Gastil, 1993) in a forearc or interarc basin or basins (sandstone-shale belt of Gastil and Miller, 1984).

Amphibolite and mafic schist, which compose ~5% of the Julian Schist, are interlayered with and share the metamorphic foliation of the schist and quartzite facies of the unit. The sizes and geometries of these mafic bodies, as well as their interfingering contacts with the enclosing sediments, suggest penecontemporaneous extrusion of mafic volcanic rocks during marine deposition (Schwarcz, 1969; Grove, l987; Todd, l994a). In addition to mafic intrusives, some metaconglomerates contain clasts (silicic tuff and a variety of intermediate to mafic lithic clasts), indicating that parts of a Late Triassic–Early Jurassic arc were undergoing erosion during deposition.

Fragmentary fossil data for the Julian Schist and possible correlative units of the sandstone-shale belt in southern California and Baja California, México, suggest a Late Triassic–Jurassic depositional age (Gastil and Miller, l984; Todd et al., l988; Gastil, l993; Germinario, l993). Whole-rock Rb-Sr isochrons of the Julian Schist and related units also suggest a Jurassic and Triassic depositional age for the protoliths (Hill, 1984; Gastil et al., l988; Davis and Gastil, 1993). Therefore, because the Julian Schist was intruded by Middle and Late Jurassic plutons (this paper; Todd et al., 1991a, 1991b; Thomson and Girty, l994), its minimum age is constrained.

Metamorphism of the Julian Schist

Between the Middle Jurassic and the Late Cretaceous, the Julian Schist underwent multiphase synkinematic metamorphism. In general, peak-grade metamorphic conditions ranged from lower amphibolite facies in the western part of the Julian Schist belt to upper amphibolite facies in the eastern part (Todd et al., 1988). The metamorphic assemblage in pelitic and semi-pelitic schists in the lower amphibolite facies is quartz-biotite-muscovite-plagioclase-K-feldspar-andalusite ± staurolite ± sillimanite ± cordierite ± garnet. In the upper amphibolite facies, pelitic rocks contain quartz-biotite-muscovite-plagioclase- K-feldspar-sillimanite ± cordierite ± garnet. Based upon thermobarometry of metapelitic rocks of the Julian Schist across the central and eastern zones of the batholith, Grove (1994) recognized the existence of a steep, north-northwest–trending gradient, across which metamorphic pressures varied from ~3 kbar on the west to ~4.5 kbar on the east. The implied increase in structural depth, from ~7 to 10 km in the western zone to ~11 to 16 km in the eastern zone, was interpreted by Grove (1994) to mark the trace of a Late Cretaceous ductile, west-directed thrust or reverse fault.

Many of the larger screens have antiformal or synformal structures, the limbs of which host steeply plunging minor folds (Schwarcz, 1969; Berggreen and Walawender, 1977; Grove, 1987). Relict bedding and sedimentary structures are preserved within the larger screens in the western part of the Julian Schist belt, but in the eastern part, migmatization and polyphase folding have obliterated primary structures (Grove, 1987; Lampe, 1988; Todd, 1994a).

Jacumba Sequence

Sillimanite-grade, flysch-like metasedimentary wallrocks in the central part of the Jurassic intrusive belt (Jacumba Mountains, Fig. 2) consist of interbedded pelitic and semi-pelitic schists, metaquartzite, thin-bedded calcitic and dolomitic marble, calc-silicate rocks, amphibolite and mafic schist, and metachert. The metamorphic assemblage in pelitic rocks is the same as the upper amphibolite facies Julian Schist. Slight lithologic differences between these rocks and the Julian Schist suggest that the two units differ in lithofacies and/or depositional age (Engel and Schultejann, l984; Gastil and Miller, l984; Todd et al., l988; Walawender et al., 1991; Gastil, l993).

The Jacumba sequence and possible equivalent rocks in northeastern San Diego County are generally similar in lithology to middle Paleozoic to Triassic clastic rocks of slope-basin facies and interbedded basalt flows in Baja California (Leier-Engelhardt, 1993). For this reason, they are tentatively assigned a Mesozoic and/or late Paleozoic age.

Paleozoic Miogeoclinal Rocks

Several large wallrock screens occur in the northern and eastern Peninsular Ranges batholith, north of Borrego Springs and east of the San Jacinto Fault Zone (SJFZ; vertical ruling in Fig. 1). These screens are described as consisting of high-grade, pre-Cretaceous metamorphic and metaplutonic rocks, Cretaceous plutonic rocks, and Paleozoic metasedimentary rocks of miogeoclinal and slope-basinal affinities (Sharp, 1967; Theodore and Sharp, 1975; Miller and Dockum, 1983; Hill, 1984; Erskine, 1986; Wagner, 1996). Common features of the pre-Cretaceous screens include upper amphibolite facies, quartz- and carbonate-rich metasedimentary rocks and anatexites, orthogneiss, and minor amphibolite. As the orthogneiss component of some of these screens is lithologically similar to the Stephenson Peak migmatite and gneiss facies of the Harper Creek suite, we speculate that Jurassic metamorphism and plutonism may have extended eastward into Paleozoic miogeoclinal crust.

JURASSIC GRANITES

Gneissic granites form a belt at least 45 km wide and 150 km long within the axial zone of the Peninsular Ranges batholith, and similar rocks are reported as far south as 28° N in Baja California (R.G. Gastil, 1991, personal commun.). In San Diego County, they occupy an area of ~1155 km^2, or roughly 13% of the total outcrop area of prebatholithic-screen rocks and granite. The western margin of the Jurassic intrusive belt approximates to the $\delta^{18}O$ step and the pseudogravity gradient (PSGR) boundary (Fig. 2), with most plutons in the western part of the intrusive belt located within an Early Cretaceous north-northwest–trending ductile shear zone in the Cuyamaca and Laguna Mountains, named the Cuyamaca–Laguna Mountains shear zone (Todd et al., 1988; Thomson and Girty, 1994). The area of Figure 3 is entirely within the Cuyamaca–Laguna Mountains shear zone, the shear zone being approximately 15 km wide at this location. The eastern part of the Jurassic intrusive belt extends at least to Borrego Springs (Fig. 1) and the Borrego Valley.

The Jurassic granites can generally be distinguished in outcrop from adjacent Cretaceous granites by their reddish-, orange-, and yellow-brown weathered surfaces (resulting from the oxidation of biotite and trace iron sulfides), abundant quartz and mica, aluminous minerals, and strong deformational fabric. The granites have been mapped as two intrusive suites, each of which comprises a number of separate bodies. The suites are: 1) the strongly peraluminous Harper Creek gneissic granite, which includes the Stephenson Peak facies of associated migmatitic schist and gneiss; and 2) the metaluminous to moderately peraluminous Cuyamaca Reservoir gneissic granite. The Cuyamaca Reservoir suite and more obvious magmatic facies of the Harper Creek suite consist chiefly of fine- to coarse-grained, medium- to dark-gray gneissic granodiorite and tonalite, but in addition, the Harper Creek suite includes mixtures of orthogneiss with paragneiss, feldspathic metapsammitic and metapelitic migmatites (diatexites), and discrete enclaves of refractory metasedimentary rocks and amphibolite. The Jurassic suites are ilmenite-series granites according to the classification of Ishihara (1977).

The intrusive relationships between the two suites change gradually from west to east across the Jurassic belt in a manner reminiscent of Gastil's (1975) depth zonation. To the west, where contacts between the two suites are sharp, the marginal Cuyamaca Reservoir rocks display higher level features and tend to be relatively coarse-grained and leucocratic (near-pegmatitic) or may be notably fine-grained against the Harper Creek gneissic granite, suggesting that, at least in these localities, Cuyamaca Reservoir magma intruded near-solidus Harper Creek plutons. Further east, contacts between the Harper Creek and Cuyamaca Reservoir plutons are more diffuse and marked by zones of gradation and interlayering that typically measure from a few meters to several tens of meters across strike. In the central part of the Jurassic belt, intrusive relationships appear to be at a deeper level and gradational zones may be as broad as 1 km, with Harper Creek plutons containing scarce lenticular bodies of Cuyamaca Reservoir gneissic granite up to 2 km long and oriented parallel to gneissic foliation. From west to east across the intrusive belt, gradational changes in the Cuyamaca Reservoir suite include mafic to felsic compositions, amphibole rather than biotite as the dominant mafic mineral, and an increase in mica-rich clots and enclaves of Julian-type metasediments.

Harper Creek Suite

The Harper Creek gneissic granites consist of biotite granodiorite and tonalite with lesser monzogranite. Where magmatic textures are least altered by subsolidus deformation and recrystallization, the mineral assemblage is plagioclase, quartz, biotite, and (in granodiorite and monzogranite) K-feldspar. Accessories include variable amounts of graphite, muscovite, tourmaline, sillimanite (mostly fibrous habit), cordierite, garnet, and andalusite.

The suite is characterized by abundant, uniformly distributed metasedimentary enclaves in a wide range of sizes. Rocks of the suite have a gneissic texture that is marked by thin (0.5 mm to several millimeters thick), evenly spaced lenticular biotite-rich clots oriented parallel to the foliation. In outcrop, reaction rims between larger, relatively intact pelitic schist enclaves and surrounding orthogneiss suggest that the biotitic clots are remnants of partly melted and/or assimilated metapelitic rocks. Mylonitic gneiss and schist of the Harper Creek suite contain thin (less than 1 cm thick) quartz stringers, boudins, and blebs that are oriented parallel to foliation. Modal analyses show that some mylonitic Harper Creek rocks plot in the field of quartz-rich granitoids on the International Union of Geological Sciences modal diagram of Streckeisen (1973). In addition to the abundant metasedimentary enclaves, the Harper Creek gneissic granite contains large tabular enclaves, or rafts, of calc-silicate–bearing metaquartzite and amphibolite, ranging in size from 10 cm to several meters long, aligned parallel to the foliation (Fig. 4A). Both enclaves and rafts are similar in lithology and metamorphic grade to the most refractory rocks in the adjacent metasedimentary screens. The fact that their distribution appears to be unrelated to distance from pluton walls suggests that the majority of these enclaves represent restite (possible "resistate" in the sense of Vernon et al., 2001) from the partial melting of supracrustal rocks in the source region rather than material stoped from walls or roof during intrusion. Amphibolite inclusions that grade laterally over short distances to bodies of fine-grained metagabbro are similar to layers of totally reconstituted amphibolite/gabbro intercalated within the Julian Schist. The majority of these inclusions are inferred to have been inherited from similar refractory mafic rocks in the Julian Schist.

Eastern plutons of the Harper Creek suite are partly mantled by migmatitic and gneissic rocks that Todd (1978) referred to as "the migmatitic schist and gneiss of Stephenson Peak." Now included as a facies of the Harper Creek suite, these rocks comprise orthogneiss, mylonitic gneiss, paragneiss, and pelitic and psam-

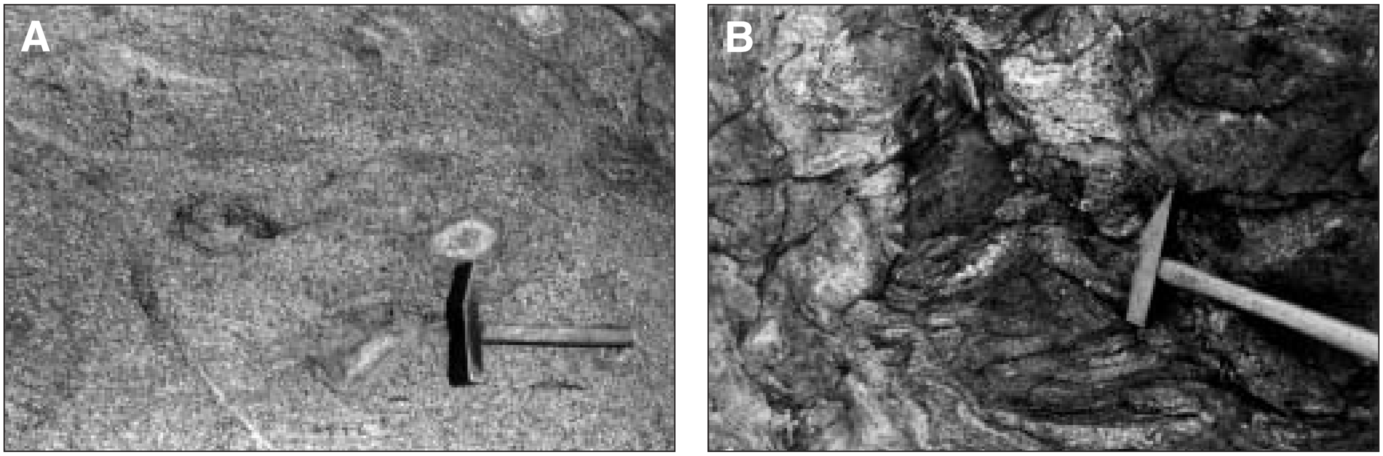

Figure 4. A: Horizontal outcrop surface of Harper Creek suite showing steeply dipping northwest-southeast foliation. Gneissic foliation is defined by flattened to ovoid enclaves of calc-silicate quartzite, amphibolite, and biotite schist that are similar in lithology to adjacent Julian Schist. B: Outcrop photograph of folded migmatite gneiss showing development of patchy granite leucosome and large single grains of white magmatic plagioclase. These rocks, The Stephenson Peak migmatitic schist and gneiss, are included as facies of Harper Creek suite and commonly occur as mantle around its plutons. Granite-migmatite relationships are similar to high-grade migmatitic zone surrounding the western part of Cooma granodiorite gneiss, Lachlan fold belt, eastern Australia (Vernon et al., 2001).

mitic schist with abundant granitic leucosome. From west to east, the bodies of Stephenson Peak facies vary from thin (<100 m) discontinuous rinds, to more sizable (1.5–2 km) envelopes surrounding Harper Creek plutons, to broad regions of diatexites and orthogneiss. Contacts between metamorphic enclaves and layers and Harper Creek orthogneiss are diffuse, wispy, or gradational. The migmatitic schist and gneiss of the Stephenson Peak facies is characterized by irregular patches of granite leucosome (Fig. 4B), discrete tabular layers of calc-silicate–bearing metaquartzite, amphibolite, mafic schist, and marble that range in thickness from a few centimeters to tens of meters. The facies forms the dominant part of the eastern Jurassic belt and appears to be composed of ultrametamorphosed metasedimentary and metavolcanic prebatholithic rocks similar to those of the Julian Schist and Jacumba sequence. The association of Harper Creek gneissic granite and migmatite in the Peninsular Ranges batholith closely parallels that of the Cooma granodiorite gneiss and migmatite in the Lachlan fold belt of eastern Australia (Vernon et al., 2001). They proposed that partial melting of prebatholithic feldspathic metapsammitic rocks was the major process in the formation of the Cooma granodiorite and that enclaves of metapsammitic residuum were restite. The formation of metapelitic (as distinct from metapsammitic) migmatite was considered by Vernon et al. (2001) to have formed earlier in the pressure-temperature (P-T) melting path, with partial melting to form leucosome limited by the amount of water released through breakdown reactions involving muscovite in the metapelite. Vernon et al. (2001) would view the earlier-formed metapelitic migmatite enclaves as "resistate" rather than restite. Similarly, outcrop-sized enclaves of metapelitic migmatite in eastern Harper Creek plutons suggest a resistate rather than restite association if the Harper Creek magma formed dominantly by partial melting of feldspar-rich metapsammitic rocks.

To the northeast of the study area in the Peninsular Ranges batholith (Fig. 3), Engel and Schultejann (1984, p. 662–663) suggested that pre-Cretaceous metamorphic wallrocks were "…feldspathized and granitized and in many places converted to gneisses almost indistinguishable from sheared and refoliated margins of [Cretaceous] plutons". These authors attributed the origin of the gneisses, which are described as containing "…relict marble beds, patches of laminated feldspathic quartzite, and other relict sedimentary features…", to amphibolite-grade metamporphism and metasomatism of wallrocks during Cretaceous intrusion. We suggest that such gneisses are S-type orthogneisses derived in large part from the anatexis of metasedimentary rocks, the character and age of which varied from west to east across the Jurassic continental margin.

Cuyamaca Reservoir Suite

Gneissic granite plutons of the Cuyamaca Reservoir suite are composed mainly of biotite- and hypersthene-biotite granodiorite and tonalite. The least deformed and recrystallized rocks contain plagioclase, subequal quartz and biotite ± orthopyroxene ± hornblende ± K-feldspar as a relict magmatic assemblage. Less common magmatic mafic assemblages in the granodiorite and tonalite are biotite–subaluminous amphibole (actinolite) and hypersthene-actinolite-biotite. Many samples fall between granodiorite and tonalite. The suite contains fine-grained mafic to intermediate enclaves (microgranitoid enclaves of Vernon, 1983) and, less commonly, enclaves of the Julian Schist. Many enclaves are dioritic in composition, have a mineral assemblage similar to that of the host rock, and have relict igneous textures. Lenticular biotite-rich aggregates ~1–2 cm long are aligned parallel to a strong mineral foliation in the Cuyamaca Reservoir plutons. Next to contacts with the larger metasedimentary screens, narrow marginal zones consist of mylonitic gneiss, in which metasedimentary enclaves (mainly calc-silicate quartzite) are concentrated near the contacts.

The Cuyamaca Reservoir suite is similar in many ways to the Hillgrove suite of the New England batholith, eastern Australia (Shaw and Flood, 1981). Plutons of the Hillgrove suite have intruded into trench-complex metasedimentary rocks, are biotite-rich, and contain actinolitic amphibole and ilmenite. Several plutons have well-developed regional metamorphic aureoles with migmatites and gneisses developed at the highest grades. The Hillgrove suite has elevated Sr_i and $\delta^{18}O$ relative to the main I-type suites of the New England batholith but lower than those in the Cuyamaca Reservoir suite.

U-Pb Geochronology

The suspicion that some deformed granites in the Cuyamaca and Laguna Mountains were Jurassic was initially based on their regional metamorphic fabric (e.g., Everhart, 1951). This was confirmed by conventional U-Pb ion exchange mass spectrometry on zircon separates (Girty et al., 1994; Thomson and Girty, 1994; G.H. Girty, 1998, personal commun.). Two samples of the Harper Creek suite yielded minimum crystallization ages of 156 ± 12 Ma and 234 ± 39 Ma, and a sample of the Cuyamaca Reservoir suite yielded an age of 161 ± 12 Ma. Girty et al. (1994) reported zircon from two additional Harper Creek samples as having Late Jurassic minimum ages. Thomson and Girty (1994), on the basis of the 234 Ma age, suggested that Triassic and Jurassic rocks are interleaved in the Scove Canyon segment of the Cuyamaca–Laguna Mountains shear zone. However, large uncertainties in age determinations due to an inherited Proterozoic (~1.5 Ga) zircon component limit the significance of age correlations. Todd et al. (1991a) reported a Rb-Sr isochron age for five whole-rock samples of 168 ± 12 Ma from a separate body of Harper Creek. A summary of available age data is plotted in Figures 2 and 3.

In Situ Single Zircon Ages—LAM-ICPMS

U-Pb age data are presented (Table 1, Fig. 5, A and B) for 26 single zircon grains from a sample of Harper Creek gneissic granite (WS-91-1). The analyses were conducted at Macquarie University on mounted and polished zircon grains using a laser ablation microprobe attached to a quadrapole induced-coupled-plasma mass spectrometer (LAM-ICPMS) following the procedure of Jackson et al. (1992). A Tera and Wasserburg (1972) isochron diagram (Fig. 5B) shows that, as pointed out by Thomson and Girty (1994), the Harper Creek gneissic granite contains a large inherited component of Paleozoic and Proterozoic zircons, many exhibiting contamination with common Pb. As the source of the common Pb is probably from the surface of the epoxy mount rather than the zircon itself (see discussion below), $^{207}Pb/^{206}Pb$ ages, although included in Table 1, are not considered meaningful. However, the six zircons that are grouped about a Jurassic age give, when pooled, a $^{207}Pb/^{206}Pb$ versus $^{238}U/^{206}Pb$ isochron age of 160.7 ± 9.5 Ma at 2σ standard errors (Fig. 5B).

In Situ Single Zircon Ages—Ion Microprobe

Zircons were extracted from seven samples of potential Jurassic age (Cuyamaca Reservoir samples CP-153, CP-178, WD-81A; Harper Creek sample 5-84-SS-2; East Mesa I-type tonalite samples TS-23, CP-155, 6-139-17) using conventional crushing, density, and magnetic methods. Epoxy mounts were prepared for hand-selected grains that were chosen on the basis of euhedral form and absence of inclusions and/or optically visible cores of older zircon. Isotopic U-Pb analysis was performed using the University of California at Los Angeles Cameca ims 1270 ion microprobe. AS-3 standard zircon (Paces and Miller, 1993) was employed to determine Pb/U relative sensitivity (see Appendix 3 of Grove et al., this volume, for additional analysis details). Unless otherwise stated, all U-Pb ages quoted are $^{206}Pb/^{238}U$ ages with ± 1σ standard error. Calculation of radiogenic Pb/U ratios was performed using ^{204}Pb as a proxy for common Pb. The vast majority of common Pb affecting our analyses is contributed from the periphery of the ion pit rather than the sample itself (Dalyrmple et al., 1999). Direct measurement of this component indicates that it is very similar in composition to anthropogenic Pb from the Los Angeles basin (see Sañudo-Wilhelmy and Flegal, 1994).

Zircon U-Pb age results (Table 2) are illustrated in $^{206}Pb/^{238}U$ versus $^{207}Pb/^{235}U$ concordia plots in Figure 6. As indicated, Middle Jurassic U-Pb zircon ages were obtained from three samples of the Cuyamaca Reservoir suite (CP-178, CP-153, and WD-81A), one sample of the Harper Creek suite (5-84-SS-2), and one sample of the East Mesa suite. The two additional samples from the East Mesa suite (CP-155 and 6-139-17) contained zircon that yielded Early Cretaceous U-Pb ages between 115 and 120 Ma and have not been included in Figure 6. The East Mesa suite consists of pyroxene-biotite-hornblende quartz diotite and tonalite plutons restricted to the Cuyamaca–Laguna Mountains shear zone. Because some East Mesa plutons interfinger with and/or grade to adjacent plutons of the Cuyamaca Reservoir suite, and others have petrographic and textural characteristics that indicate a Cretaceous age, Todd et al. (this volume) designate the suite as Jurassic and Cretaceous.

The typically low (50–200 ppm) U-contents of zircons from all of the samples that we have studied result in relatively poorly defined $^{207}Pb/^{235}U$ ages after correction for common Pb have been applied. Hence, while all results statistically overlap concordia, the $^{207}Pb/^{235}U$ age data we obtained are too imprecise to allow us to conclude with a high degree of confidence that the U-Pb analyses are unaffected by inheritance and/or Pb loss. In spite of this, only one sample (WD-81A) yielded clear evidence for such phenomena (Fig. 6C). For example, the rim of one zircon from WD-81A yielded an early Cretaceous $^{206}Pb/^{238}U$ age (118 ± 3 Ma), while two additional highly discordant (and much older) results from other grains indicated incorporation of inherited Pb. Given the high spatial resolving power of the ion microprobe, we feel that it is likely that we were generally able to avoid the problematic behavior that was prominent in previous isotope dilution studies of these rocks (Thomson and Girty, 1994; Murray and Girty, 1996) and

TABLE 1. ZIRCON Pb/U SINGLE GRAIN ANALYSES, HARPER CREEK, SAMPLE WS-91-1

	Isotope ratios				Age estimates (Ma)			
Grain	$^{206}Pb/^{238}U$	± 2σ	$^{207}Pb/^{206}Pb$	± 2σ	$^{206}Pb/^{238}U$	± 2σ	$^{207}Pb/^{206}Pb$	± 2σ
24	0.02391	0.00106	0.06769	0.00850	152.3	6.7	858.9	250.4
2	0.02496	0.00072	0.08143	0.00710	158.9	4.5	1231.9	166.3
34	0.02533	0.00080	0.04898	0.00622	161.3	5.0	147	285.2
15	0.02538	0.00084	0.07407	0.00750	161.6	5.0	1043.5	197.7
8	0.02562	0.00140	0.07471	0.01914	163.1	8.9	1060.7	476.3
18	0.02585	0.00094	0.07421	0.00682	164.5	6.0	1047.1	180.2
32	0.03017	0.00170	0.05125	0.01598	191.6	10.7	252.1	647.6
11	0.04359	0.00092	0.05778	0.00274	275.1	5.7	521.2	102.4
26	0.04393	0.00134	0.05581	0.00528	277.2	8.2	444.4	204.2
12	0.04514	0.00130	0.06860	0.00444	284.6	8.0	886.8	131.1
6	0.04793	0.00178	0.09596	0.00818	301.8	11.0	1547	156.1
20	0.05016	0.00096	0.08176	0.00314	315.5	5.9	1239.7	73.9
16	0.05055	0.00150	0.07677	0.00538	317.9	9.3	1115.1	136.8
23	0.05316	0.00128	0.08693	0.00432	333.9	7.9	1359	94.0
4	0.05478	0.00134	0.09440	0.00448	343.8	8.2	1516.1	88.2
21	0.05648	0.00154	0.10040	0.00658	354.2	9.4	1631.5	119.4
22	0.05696	0.00122	0.05955	0.00292	357.1	7.5	587.1	104.5
14	0.06080	0.00132	0.07379	0.00338	380.5	8.1	1035.7	91.0
29	0.07263	0.00250	0.09176	0.00710	452.0	15.0	1462.3	143.8
3	0.07479	0.00192	0.08872	0.00428	465.0	11.5	1398	90.8
13	0.09337	0.00208	0.09805	0.00394	575.4	12.2	1587.4	74.0
30	0.11549	0.00354	0.06685	0.00436	704.6	20.5	833	133.2
7	0.11947	0.00358	0.11013	0.00626	727.5	20.7	1801.6	101.8
28	0.15504	0.00492	0.07923	0.00488	929.1	27.5	1178	119.3
25	0.18826	0.00580	0.08266	0.00540	1111.9	31.5	1261.2	125.0
31	0.25992	0.00714	0.09923	0.00418	1489.4	36.5	1609.7	77.4

Note: LAM-ICPMS analyses, Macquarie University. Ablation diameter, 30 μ. Wavelength, 266 nm. Method after Jackson et al. (1992).

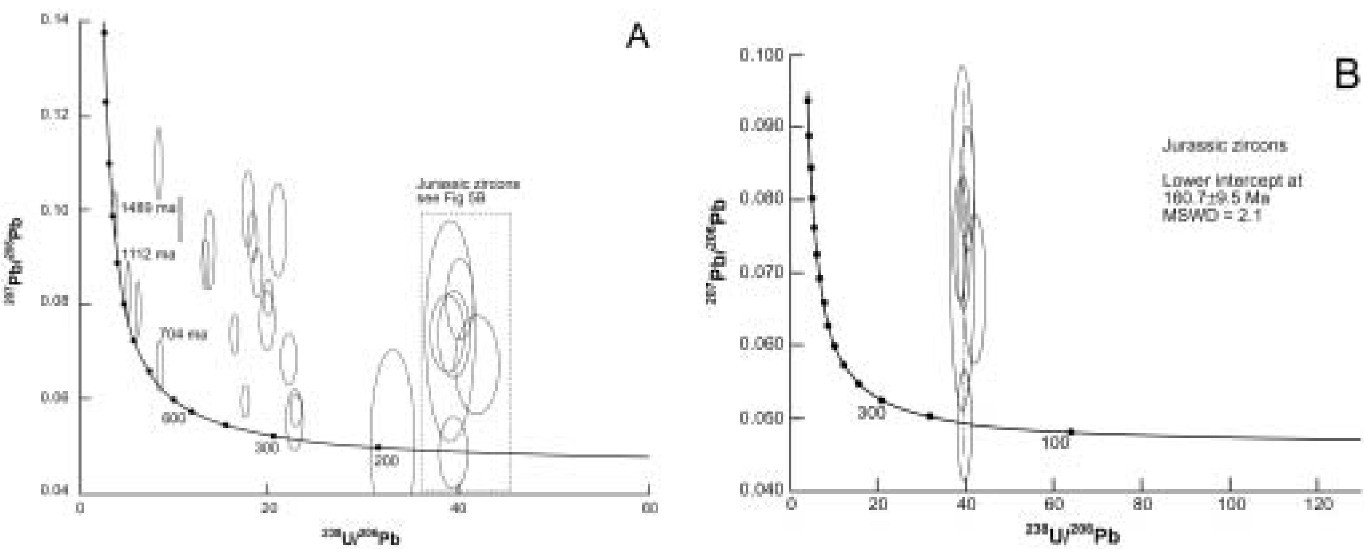

Figure 5. A: Tera and Wasserburg (1972) isochron diagram of 26 single zircon analyses for Harper Creek sample WS-91-1, dated using procedure of Jackson et al. (1992). Ages based on $^{238}U/^{206}Pb$ show spread from Jurassic to Proterozoic. Discordant $^{207}Pb/^{206}Pb$ ages in some zircons indicate significant common lead in direction of elongation of error ellipses and are probably due to surface contamination of sample. B: Isochron diagram of six Jurassic zircons outlined in A, giving calculated least-squares age of 160.7 ± 9.5 Ma 2σ standard errors. MSWD 2.1 (mean square of weighted deviates).

TABLE 2. IONPROBE* U-Pb AGE RESULTS FROM ZIRCON

Sample	Suite	$^{206}Pb/^{238}U^{\#}$ ± 1σ (Ma)	$^{207}Pb/^{235}U^{\#}$ ± 1σ (Ma)	MSWD	Number**
CP-153	Cuyamaca Reservoir	168 ± 2	169 ± 18	3.5	11/11
CP-178	Cuyamaca Reservoir	161 ± 3	154 ± 25	3.4	10/10
WD-81A	Cuyamaca Reservoir	165 ± 2	164 ± 11	4.9	11/14[##]
5-84-SS-2	Harper Creek	164 ± 4	171 ± 44	3.8	10/10
TS-23	Tonalite of East Mesa	170 ± 2	157 ± 29	3.5	10/10

*UCLA's Cameca ims1270. See Appendix 3 in Grove et al. (this volume, Chapter 14) for additional details on analytical methods.
[#]Weighted mean age. Standard errors (± 1σ) shown have been scaled by the square root of the MSWD.
**Number of spot analyses used to calculate mean age/total number of spot analyses measured.
[##]Excluded measurements for WD-81A include rim analysis of 118 Ma, and two highly discordant analyses (206 Ma and 564 Ma).

provisionally conclude that the majority of the analyses we have obtained are concordant and represent the age of intrusion.

Weighted mean U-Pb ages are presented in Table 2 and represented diagrammatically by filled ellipses in Figure 6. In nearly all instances, the mean square of weighted deviates (MSWD) values are significantly in excess of unity. We believe that the most likely explanation for this is instrumental instability that is not adequately taken into account in the error propagation. For example, our ability to reproduce $^{206}Pb/^{238}U$ ratios from AS-3 standard zircon during the analysis sessions in question was about ±2 % (1σ). Hence, to ensure that the stated uncertainties adequately reflect instrumental instability, we have scaled the calculated standard errors by the square root of the MSWD in all cases where the MSWD exceeded unity.

Complete data tables for each sample investigated have been included in the GSA Data Repository[1] and on the CD-ROM that accompanies this volume.

Structural Characteristics

The Jurassic granites form steep-walled, north-northwest–elongate bodies as much as 5 km wide and 20 km long. The rock textures in these bodies range from strongly foliated to mylonitic gneiss. Foliation strikes parallel to the sides of the bodies; dips are steep (60–90°) to the east and defined by alignment of relict magmatic feldspar and quartz grains and by recrystallized aggregates of quartz and biotite. A penetrative lineation plunges steeply to the east in the plane of the foliation. Modification of igneous grains in the Jurassic granites is variable: foliated granites contain aligned relict subhedral feldspar and biotite phenocrysts and interstitial quartz, whereas in mylonitic rocks, the larger feldspar grains are porphyroclasts. Most quartz and biotite form fine-grained recrystallized aggregates, and both K-feldspar and biotite are extensively altered to muscovite. The apparent concordance of magmatic and subsolidus foliations in the Jurassic plutons and the continuity of plutonic foliation and lineation with regionally developed metamorphic fabrics in their wallrocks indicate that magmatic foliation was overprinted by a high-temperature, post-magmatic solid-state foliation (Paterson et al., 1989).

The foliation in the Harper Creek and Cuyamaca Reservoir suites is axial planar to scattered outcrop-scale isoclinal folds, and map patterns of several bodies suggest that the plutons were folded isoclinally about near-vertical axes on a regional scale prior to Cretaceous intrusion (Todd, 1994b). The fabric of these suites probably formed during multiple episodes of synintrusive deformation that began at least as early as the Late Jurassic (Todd et al., 1994) and culminated by the middle Cretaceous (Thomson and Girty, 1994). Apparently, deformation was under way in the Triassic and Jurassic prior to Jurassic intrusion, because both the Harper Creek and Cuyamaca Reservoir plutons contain, as xenoliths, the detached hinges of isoclinal folds in refractory Julian Schist rock-types.

The fabric of the Jurassic plutons was not fully developed until the Cretaceous period. Foliation and lineation in Early Cretaceous plutons within the Cuyamaca–Laguna Mountains shear zone have orientations similar to those in the adjacent Jurassic plutons, and in some outcrops, foliation is continuous between Jurassic and Cretaceous plutons, crossing contacts at a high angle (Fig. 3). Studies of two bodies of the Harper Creek gneissic granite located within the Cuyamaca–Laguna Mountains shear zone indicate that mylonitic fabrics developed during two periods of Cretaceous deformation (Girty et al., 1993; Thomson and Girty, 1994).

Petrographic Characteristics

Microstructures of the Harper Creek (including the Stephenson Peak facies) and Cuyamaca Reservoir suites indicate high-temperature strain and recrystallization of magmatic grains during late-(?) and post-crystallization subsolidus deformation. In thin section, quartz and feldspar are generally similar between

[1]GSA Data Repository item 2003XXX, chemical analyses of 62 Jurassic plutonic rocks and 12 eastern-zone prebatholithic rocks (Julian Schist, orthoamphibolites), Peninsular Ranges batholith, San Diego County, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA, at editing@geosociety.org, www.geosociety.org/pubs/ft2003.htm, or on the CD-ROM accompanying this volume

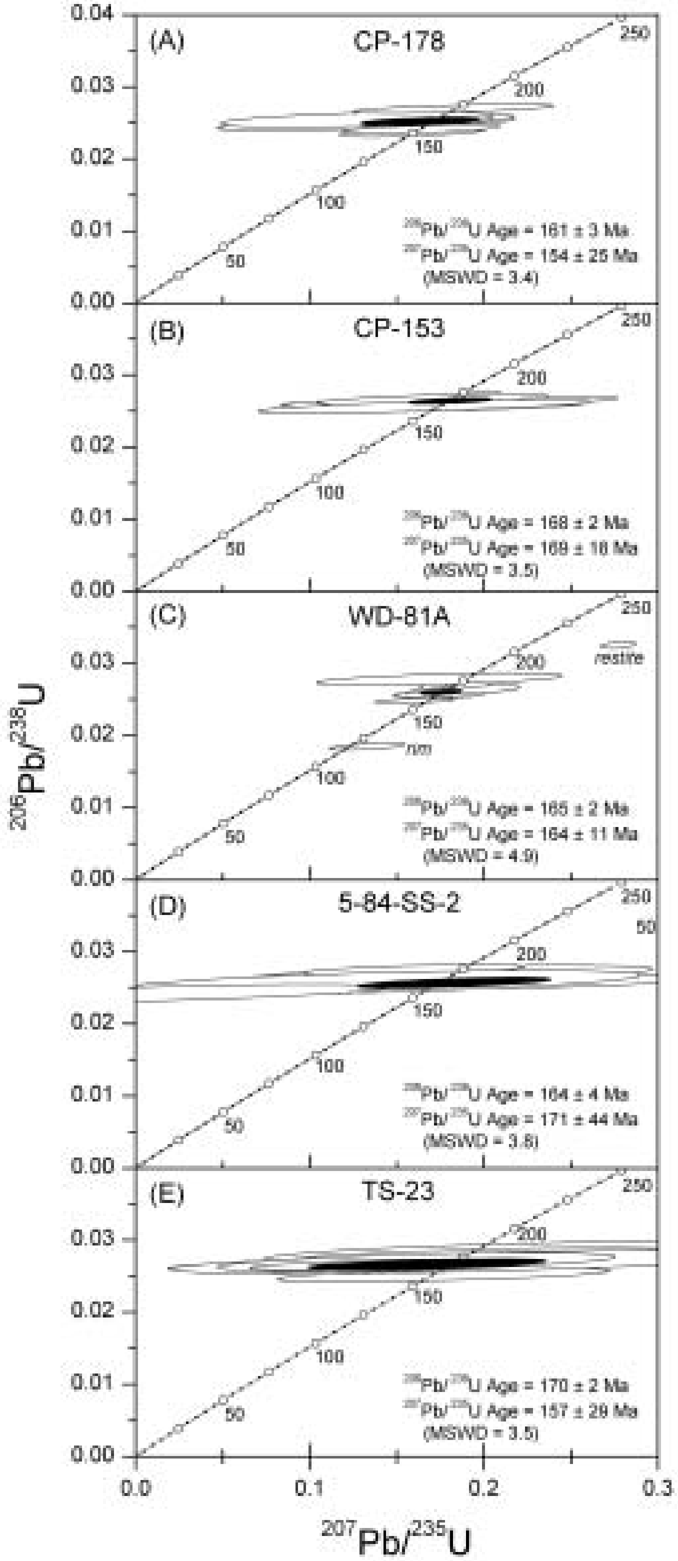

Figure 6. Concordia diagrams $^{206}Pb/^{238}U$ versus $^{207}Pb/^{235}U$ for zircons from five Jurassic plutons. Cuyamaca Reservoir suite CP-178 (A), CP-153 (B), and WD-81A (C). Harper Creek suite 5-84-SS-2 (D). East Mesa suite TS-23 (E). Age errors in Ma ± 1σ standard error.

the two suites, although mafic and accessory mineral contents are different (discussed below).

The plagioclase composition of the Harper Creek suite varies from oligoclase to andesine, whereas that of the Cuyamaca Reservoir suite is typically sodic to calcic andesine. Cores of normally zoned grains in both suites are as calcic as sodic labradorite, and rims are as sodic as oligoclase. Scarce relict grains showing magmatic zoning are more common in the Cuyamaca Reservoir suite (Fig. 7A) than in the Harper Creek suite (Fig. 7B), but typically zoning has been obliterated by wedge-shaped deformation twins, kinking, fracturing, and incipient subgrain formation.

Quartz in the Harper Creek and Cuyamaca Reservoir suites varies in size and degree of recrystallization from strain-shadowed relict magmatic grains ~10 mm long to granoblastic grains less than 1 mm across (Fig. 7B). In rocks that contain the best-preserved magmatic quartz, the mineral commonly is interstitial to plagioclase. In both suites, 3–5 mm lenticular multi-grain quartz aggregates and, in the most deformed rocks, ribbons from 0.5 to 1 cm long are common.

Potassium feldspar content in the granodiorite and tonalite of both suites ranges from almost zero to 12%. The mineral is present as 5- to 15-mm-long lenticular grains, as smaller 0.5–3 mm irregular interstitial grains, and as fine-grained granoblastic aggregates with quartz, plagioclase, and biotite. Although most K-feldspar is recrystallized, the larger grains within the granodiorite of both suites show relict magmatic textures such as subhedral shape, small euhedral inclusions of early-crystallized plagioclase, and Carlsbad twins. Potassium feldspar-plagioclase grain contacts are marked by extensive replacement of plagioclase by K-feldspar as well as by nucleation of myrmekite on plagioclase grains (replacement of adjacent K-feldspar, Vernon, 1991).

Biotite in both suites has a "foxy" red-brown color characteristic of reduced biotite in S-type granites (Hine et al., 1978; Shaw and Flood, 1981; Clemens and Wall, 1988). The mineral occurs as 1- to 5-mm-long, relatively equant, subhedral and anhedral grains (relict magmatic grains) and as 2- to 5-mm-long decussate aggregates interleaved with graphite, ilmenite, and minor chlorite (recrystallized biotite).

Muscovite is restricted to the Harper Creek suite, varies in abundance from zero to 13%, and is present mainly as decussate aggregates with biotite and also as sheaf-like aggregates or ragged poikiloblasts that overprint feldspar grains and quartzofeldspathic matrix. The mineral is most abundant in mylonitic Harper Creek rocks. Nearly equant, 1–4 mm anhedral or subhedral muscovite grains with ragged terminations (possibly relict phenocrysts) are rare. Textural relations suggest that most muscovite of the Harper Creek suite is secondary.

Prominent accessory minerals of the Harper Creek suite are graphite, apatite, zircon, tourmaline, allanite, and ilmenite. Graphite, in some cases, makes up more than 2% of the rock. Zircon in Harper Creek rocks occurs as two populations: 1) small colorless euhedral magmatic grains; and 2) pale yellowish-tan, subhedral, broken or rounded inherited grains. Irregular to subhedral partial overgrowths of colorless zircon were observed on some colored

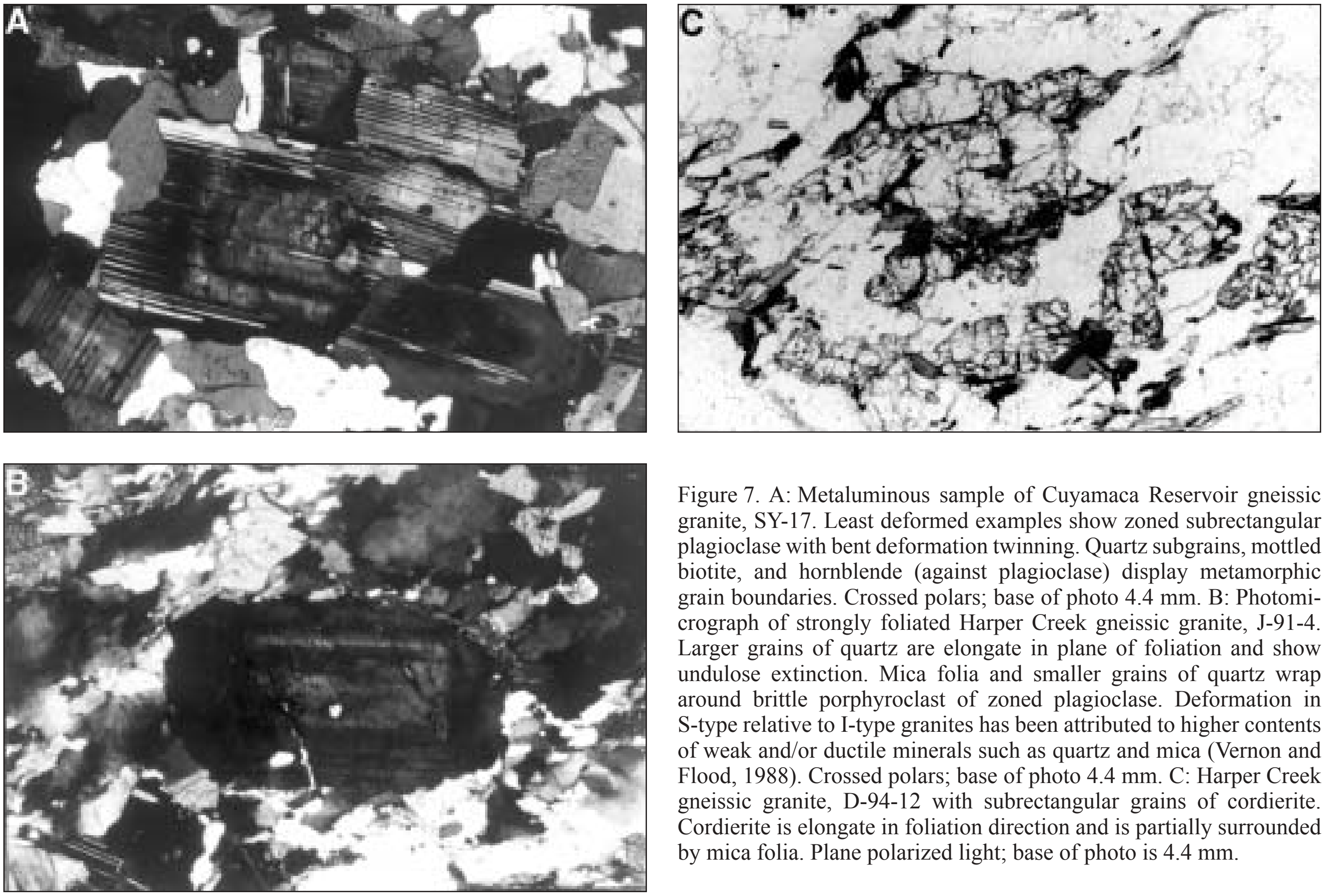

Figure 7. A: Metaluminous sample of Cuyamaca Reservoir gneissic granite, SY-17. Least deformed examples show zoned subrectangular plagioclase with bent deformation twinning. Quartz subgrains, mottled biotite, and hornblende (against plagioclase) display metamorphic grain boundaries. Crossed polars; base of photo 4.4 mm. B: Photomicrograph of strongly foliated Harper Creek gneissic granite, J-91-4. Larger grains of quartz are elongate in plane of foliation and show undulose extinction. Mica folia and smaller grains of quartz wrap around brittle porphyroclast of zoned plagioclase. Deformation in S-type relative to I-type granites has been attributed to higher contents of weak and/or ductile minerals such as quartz and mica (Vernon and Flood, 1988). Crossed polars; base of photo 4.4 mm. C: Harper Creek gneissic granite, D-94-12 with subrectangular grains of cordierite. Cordierite is elongate in foliation direction and is partially surrounded by mica folia. Plane polarized light; base of photo is 4.4 mm.

zircon cores. Tourmaline that is pleochroic in yellow and yellow-green hues is a minor but ubiquitous mineral in Harper Creek rocks. It occurs mainly as 1–2 mm anhedral grains within micaceous folia and, more rarely, as euhedral or subhedral grains.

One or more of the aluminosilicate minerals sillimanite, cordierite, garnet, and andalusite are found in variable amounts in Harper Creek rocks. Fibrous sillimanite and, less commonly, small prismatic grains of sillimanite, are associated with muscovite along biotite cleavage planes and in relict plagioclase and K-feldspar phenocrysts. Harper Creek rocks in the Cuyamaca and Laguna Mountains contain pseudomorphous aggregates of muscovite, chlorite, and fibrous sillimanite that partly or wholly replace cordierite. Other rocks contain relatively abundant, essentially unaltered cordierite, mostly as small anhedral grains associated with biotite, but, in a few rocks, phenocryst-sized rectangular grains are interpreted as magmatic (Fig. 7C). Scarce garnet porphyroblasts occur within aggregates of biotite, muscovite, and quartz in some thin sections.

The least altered Cuyamaca Reservoir rocks contain the relict magmatic assemblage plagioclase-orthopyroxene-quartz-biotite ± K-feldspar ± hornblende. Less common magmatic mafic assemblages in granodiorite and tonalite are biotite-subaluminous amphibole (actinolite) and hypersthene-actinolite-biotite. Hypersthene, in places partly bordered by clinopyroxene, is relatively abundant as subhedral or skeletal magmatic grains and is commonly replaced by biotite and/or actinolite. Actinolite also occurs as well-formed prismatic grains in both granodiorite and tonalite, with or without pyroxene. Accessory minerals include titanite, apatite, ilmenite, zircon, and allanite. Titanite, which may form as much as 1% of the rock, occurs as: 1) subhedral grains up to 3 mm across associated with, or included within, biotite (primary titanite); or 2) narrow rims on ilmenite, tiny seed-like grains along biotite cleavage planes, and in decussate aggregates of mafic minerals (secondary titanite). Clear, colorless zircons are euhedral to subhedral, with a few zoned grains containing rounded (inherited?) tan cores.

Geochemical Variations

Whole-Rock Chemistry

Analyses of eight typical granites from the Harper Creek and Cuyamaca Reservoir suites for major oxides and trace elements are listed in Table 3. Sixty-two analyses from these two suites and 20 from the Julian Schist and interbedded rocks are included in

TABLE 3. COMPOSITION OF SELECTED JURASSIC GNEISSIC GRANITES

	Harper Creek suite				Cuyamaca Reservoir suite				Cooma*
	MP-35	10-79-E	WS-91-1	WS-94-4	D-29	CP-178	ML-5	SY-12	C1
SiO_2	66.07	68.27	71.10	74.47	61.75	65.49	68.49	73.13	72.00
TiO_2	0.54	0.59	0.58	0.48	1.06	1.12	0.88	0.65	0.54
Al_2O_3	15.70	14.65	13.96	12.61	15.29	14.25	14.03	12.93	13.72
Fe_2O_3	0.71	1.04	0.87	0.35	0.78	0.45	0.43	0.57	0.59
FeO	2.74	2.97	2.42	1.77	4.48	4.06	4.25	1.94	3.48
MnO	0.10	0.05	0.06	0.03	0.12	0.08	0.06	0.04	0.06
MgO	1.85	1.72	1.66	1.25	3.95	3.29	1.47	0.78	1.76
CaO	3.22	3.26	1.71	2.63	5.89	4.47	3.07	2.14	0.95
Na_2O	2.75	2.34	1.54	2.70	2.89	2.63	2.54	2.44	1.49
K_2O	4.03	2.96	3.37	1.60	2.71	2.79	3.39	3.96	3.73
P_2O_5	0.30	0.12	0.06	0.14	0.16	0.19	0.16	0.08	0.13
H_2O+	1.26	1.10	1.47	0.76	1.12	1.14	0.47	0.55	n.d.
H_2O-	0.10	0.09	0.21	0.10	0.07	0.11	0.09	0.05	n.d.
CO_2	0.20	0.74	2.71	1.74	0.13	0.31	0.03	0.18	n.d.
Carbon	n.d.	n.d.	0.74	0.47					
Trace elements (parts per million)									
Ba	1360	1261	1353	715	959	867	942	1697	765
Cr	39	63	77	47	171	150	31	25	56
Cu	18	41	20	16	19	23	36	<1	n.d.
Ga	19	17	14	15	24	17	19	12	n.d.
Nb	n.d.	n.d.	14	12	12	13	n.d.	9	n.d.
Ni	18	25	7	6	16	16	5	8	24
Pb	22	25	18	11	14	12	18	16	35
Rb	143	98	126	71	101	112	139	135	153
Sr	224	223	182	221	174	156	133	106	127
Th	<2	12	8	8	16	10	11	14	22
U	4	4	3	<3	<3	<3	4	5	4
V	52	108	122	68	126	71	59	39	39
Y	35	22	23	21	50	38	48	40	23
Zn	100	108	82	78	89	75	106	41	n.d.
Zr	81	179	155	176	136	192	228	214	201
C (CIPW)#	0.96	1.67	4.67	1.66			0.60	0.74	5.51
ASI (Mol)**	1.065	1.129	1.503	1.151	0.831	0.921	1.045	1.061	1.670

*Cooma gneiss, Lachlan fold belt, eastern Australia (taken from White and Chappell, 1988).

#C (CIPW) is percent normative corundum.

**ASI (Mol) is the aluminum saturation index as $Al_2O_3/(Na_2O+K_2O+CaO)$.

n.d. = not determined.

Sample locations as latitude and longitude:

Sample	Lat. W	Long. N	Sample	Lat. W	Long. N
MP-35	116°29′03″	32°56′19″	D-29	116°33′13″	32°48′03″
10-79-E	116°31′26″	32°48′52″	CP-178	116°33′58″	33°00′02″
WS-91-1	116°38′49″	33°18′26″	ML-5	116°28′32″	32°51′15″
WS-94-4	116°35′58″	32°22′44″	SY-12	116°40′18″	33°01′31″

the GSA Data Repository and on the CD-ROM accompanying this volume (see footnote 1). Unlike the Early Cretaceous I-type granites of the Peninsular Ranges batholith, the Jurassic granites have the characteristics of Australian S-type granites (Hine et al., 1978; White and Chappell, 1988; Shaw and Flood, 1981). In particular, the Jurassic granites have higher values of K_2O, TiO_2, MgO, Ba, Rb, Cr, Ni, Zn, and Nb, and lower values of CaO and Na_2O (Todd and Shaw, 1995). In common with S-type granites of eastern Australia, the Harper Creek rocks have a limited SiO_2 range from 65% to 75% and compositionally overlap for most major oxides, but there are some differences. The Harper Creek suite and the Lachlan fold belt S-type granites, including a Cooma gneiss sample, are plotted on a Na_2O versus K_2O diagram in Figure 8. The Lachlan fold belt analyses, apart from a few scattered points, have Na_2O values equal to or less than K_2O, a consequence of Na_2O being differentially removed from sediments by seawater (Hine et al., 1978). Overall, the trend of the Lachlan fold belt S-types is toward increasing Na_2O with increasing K_2O, and in general the Harper Creek analyses fall within the field of the Lachlan fold belt S-types (Fig. 8). However, at lower K_2O values, the data field overlaps that of the Julian Schist, indicating a significant restite (migmatite) component of the Harper Creek magma at those values.

The Harper Creek gneissic granites are strongly peraluminous while those of the Cuyamaca Reservoir vary between metaluminous and peraluminous compositions (Table 3). The range in composition of these rocks is shown on an aluminum saturation index (ASI) versus FeO (total) diagram in Figure 9. For comparison, a sample of Cooma gneiss analyzed by White and Chappell (1988) is included in Table 3 and Figure 9. The horizontal line at an ASI of 1 (Fig. 9) separates metaluminous and peraluminous compositions. Two arrowhead lines define a field that constrains the Julian Schist and the Harper Creek granites. The field converges toward the minimum-melt composition of the simple granite system, based on an ASI value of 1 and an FeO value close to 1.5%. The extreme scatter of Harper Creek points, ranging from Julian metasedimentary compositions toward the granite minimum, is strongly suggestive of a considerable proportion of restite being retained during ascent and emplacement.

Also plotted on Figure 9 are samples of the metaluminous to peraluminous Cuyamaca Reservoir suite. In general, the trend is toward the granite minimum, but some are significantly peraluminous and not in the direction that would be expected from crystal fractionation of an initial metaluminous Cuyamaca Reservoir magma alone, suggesting at least two source components were involved. It has already been noted that the composition of the Cuyamaca Reservoir suite varies geographically. To the west, the plutons are more mafic and contain amphibole as a major phase, while to the east, biotite becomes significant,

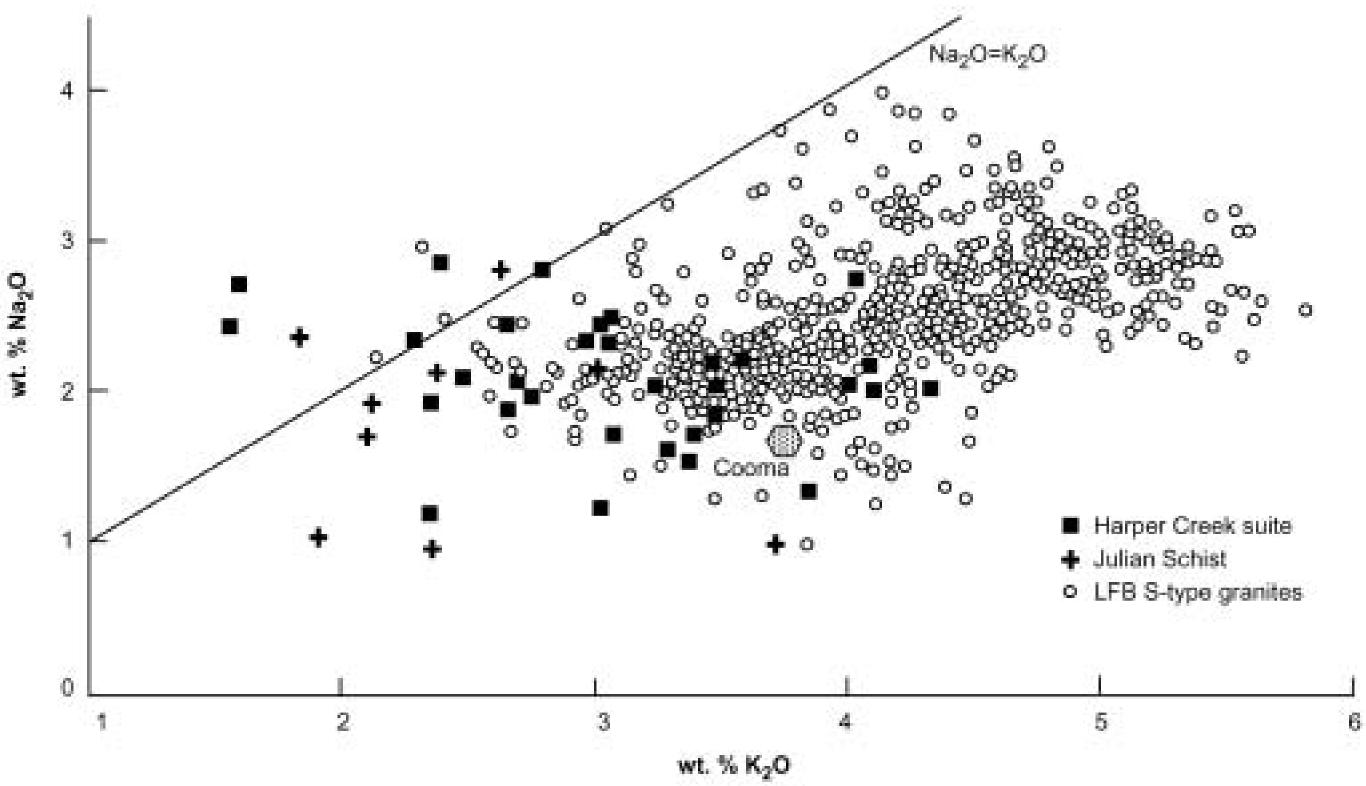

Figure 8. Na_2O versus K_2O diagram of Harper Creek suite and Julian Schist. Lachlan fold belt (LFB) S-type granites of eastern Australia, including Cooma granodiorite gneiss, are added for comparison. Samples of Harper Creek with $Na_2O > K_2O$ probably reflect high restite content of Julian Schist.

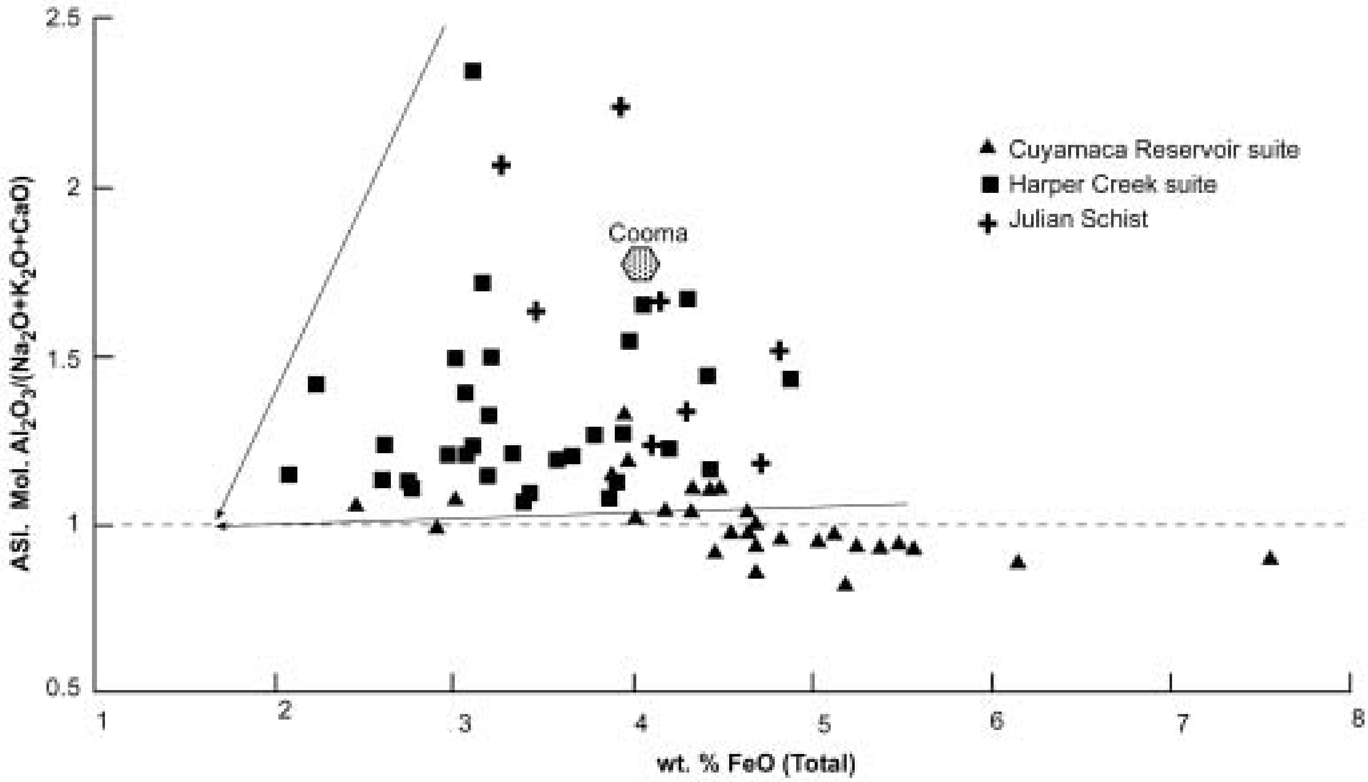

Figure 9. ASI versus FeO (total) diagram of Harper Creek S-type suite, Cuyamaca Reservoir transitional I-type to S-type suite, and Julian Schist. Lachlan fold belt Cooma granodiorite gneiss is added for comparison. Harper Creek samples are bounded by two arrowed lines and converge toward minimum melt composition of simple granite system at an ASI value of 1.0. Harper Creek samples lie between Julian Schist compositions and granite minimum. Cuyamaca Reservoir samples are metaluminous at mafic compositions but at lower FeO values are weakly to moderately peraluminous, reflecting a compositional change from west to east across the Jurassic intrusive belt.

the plutons are more felsic, and enclaves of Julian Schist are more common.

Geochemical variations between the Harper Creek and Cuyamaca Reservoir granites (Fig. 10) show considerable overlap on Harker diagrams for TiO_2, P_2O_5, CaO, Ba, Rb, Y, Zr, and Sr. Neither suite is characterized by regular chemical trends, apart from CaO and Ba for the Cuyamaca Reservoir. The scatter of data points for the Harper Creek samples, particularly at high SiO_2 contents, would suggest that crystal fractionation and restite removal at minimum melt compositions were ineffective.

$^{87}Sr/^{86}Sr$ *and* $\delta^{18}O$

Published Sr_i data for the Cretaceous granites demonstrate a general uniformity parallel to the axis of the batholith and a progressive increase from about 0.704 in the western margin of the batholith to 0.709 in the eastern zone (Silver and Chappell, 1988). Isopleths of $\delta^{18}O$ for the Cretaceous granites show a similar parallelism to the axis of the batholith, rising from +6 to +8.5 per mil in the western zone to +9 to +12 per mil in the eastern zone (Taylor and Silver, 1978). The higher eastern values indicate that these magmas were derived from, or had interacted with, rocks formed at near-surface environments (O'Neil and Chappell, 1977; O'Neil et al., 1977; Taylor and Silver, 1978; Shaw and Flood, 1981).

The Jurassic gneissic granites are significantly different from the Cretaceous I-type granites with respect to Sr_i and $\delta^{18}O$ values. Table 4 lists whole-rock $\delta^{18}O$ and Sr_i values for the Harper Creek and Cuyamaca Reservoir suites. The age used for the Sr_i calculation was taken as 161 Ma. Initial $^{87}Sr/^{86}Sr$ ratios and $\delta^{18}O$ values of the Harper Creek suite vary from 0.7105 to 0.7129 and +14.8 to +19.8 per mil respectively, whereas those of the more mafic Cuyamaca Reservoir suite vary from 0.7064 to 0.7082 and +11.8 to +13.8 per mil. Differences between the Jurassic and Cretaceous plutonic rocks are clearly shown on a plot of Sr_i versus $\delta^{18}O$ (Fig. 11) where all but four Cretaceous samples have Sr_i values <0.705 and $\delta^{18}O$ values $\leq$+9.0 per mil. The Jurassic gneissic granites form a roughly linear array with a positive slope in the high Sr_i-high $\delta^{18}O$ region of the diagram, with a small but distinct gap between the Harper Creek and Cuyamaca Reservoir samples.

The $\delta^{18}O$ and Sr_i values of the Jurassic gneissic granites, particularly those of the Harper Creek suite, are higher than those reported for the Lachlan fold belt and New England batholith

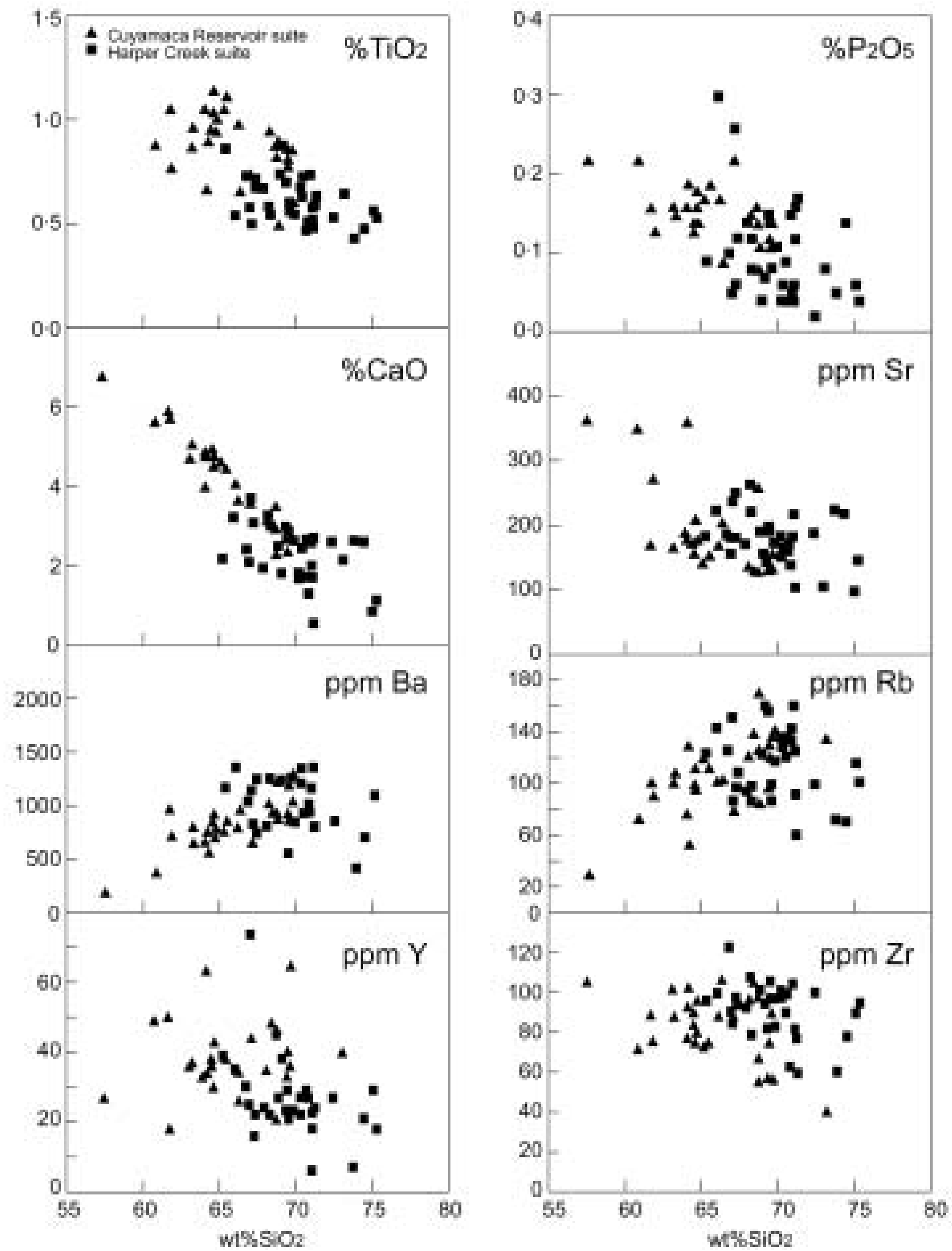

Figure 10. Selected Harker diagrams of Harper Creek and Cuyamaca Reservoir suites. Neither suite displays regular inter-element variation patterns and there is considerable compositional overlap between the suites.

S-type granites (O'Neil and Chappell, 1977; O'Neil et al., 1977), and therefore it is unlikely that these rocks were derived solely from a mantle or young oceanic crustal source (O'Neil et al., 1977; Shaw and Flood, 1981). While the overall chemical and isotopic behavior of the Harper Creek can well be explained by partial melting of a single component, the Julian Schist and related metasediments, the Cuyamaca Reservoir suite from Figure 11 requires an origin from possibly three source components: 1) young mantle-derived magma; 2) relatively young metaigneous and metavolcanic mafic rocks formed in an arc environment; and 3) a metasedimentary component equivalent to or less isotopically evolved than the Julian Schist. The isotopic gap between the mantle-derived Cretaceous (and presumably the Jurassic) I-type magmas and the Cuyamaca Reservoir suite (Fig. 11) would not argue strongly for a significant contribution from a young mantle-derived component.

Rare Earth Compositions

REE data for 17 rock samples of Julian Schist, Harper Creek suite, and Cuyamaca Reservoir suite are presented in Table 5. The values are plotted as standard chondrite-normalized (C1, Sun and McDonough, 1989) patterns in Figure12.

TABLE 4. STRONTIUM AND OXYGEN ISOTOPIC ANALYSES

Sample number	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$ measured	$^{87}Sr/^{86}Sr$ present day	$^{87}Sr/^{86}Sr$ initial	$\delta^{18}O$* rock
Cuyamaca Reservoir suite#						
CP-153	114	150	2.200	0.71143	0.70640	11.8
CP-178	113	160	2.054	0.71202	0.70732	12.1
D-29	111	167	1.934	0.71153	0.70710	11.8
J-38	139	154	2.605	0.71401	0.70805	13.8
TS-55	81.2	180	1.303	0.71083	0.70785	12.4
TS56	103	168	1.772	0.71224	0.70818	12.4
10-79-H	79	176	1.300	0.70970	0.70672	12.2
Harper Creek suite#						
J-40	133.7	189.9	2.041	0.71613	0.71146	15.9
Jhc-1	65.8	50.2	3.796	0.72159	0.71290	19.8
Jhc-2	159.3	164.2	2.813	0.71740	0.71096	17.9
10-79-E	99	216	1.319	0.71372	0.71070	14.8
10-79-G	90	211	1.238	0.71332	0.71049	16.2
10-79-M	119	185	1.862	0.71509	0.71083	16.7
MP-35	143.1	227.4	1.824	0.71415	0.70998	17.7
Julian Schist**						
10-79-I	76	320	0.686	0.71308	0.71103	16.7
10-79-J	121	258	1.358	0.71342	0.70936	17.2
10-79-L1	145	144	2.922	0.71874	0.71001	17.7

Note: Oxygen analyses, USGS Menlo Park. Analyst J. R. O'Neil. Rb and Sr analyses by XRF, Macquarie University. Variance $^{87}Rb/^{86}Sr$=1.0%. $^{87}Sr/^{86}Sr$ analyses, Centre for Isotope Studies, Sydney. Variance = 0.5%.

*Relative to SMOW as per mil.

#Initial $^{87}Sr/^{86}Sr$ calculated at 161 Ma.

**Initial $^{87}Sr/^{86}Sr$ calculated at 210 Ma.

Sample locations as latitude and longitude:

Sample	Lat. N	Long. W	Sample	Lat. N	Long. W
CP-153	33º00′22″	116º3′20″	10-79-L1	33º00′13″	116º37′17″
CP-178	33º00′02″	116º33′58″	J-40	33º05′33″	116º31′26″
D-29	32º48′03″	116º33′13″	Jhc-1	32º56′02″	116º26′49″
J-38	33º05′26″	116º34′48″	Jhc-2	32º56′02″	116º26′49″
TS-55	33º00′10″	116º42′29″	10-79-E	32º48′52″	116º31′26″
TS-56	32º57′22″	116º38′50″	10-79-G	32º46′46″	116º38′29″
10-79-H	32º46′41″	116º38′31″	10-79-M	32º04′14″	116º32′36″
10-79-I	32º53′35″	116º26′49″	MP-35	32º56′19″	116º29′03″
10-79-J	33º03′47″	116º34′06″			

Gromet and Silver (1987), in a regional study of the Cretaceous gabbros and granites across the batholith, showed a systematic variation of REE patterns from west to east. Their western region, which corresponds to the western zone as used in this paper, has as its boundary to the east regional discontinuities such as the PSGR gradient and $\delta^{18}O$ gap, age step, gabbro line, and other major isopleths (Todd and Shaw, 1985, their Fig. 2). Despite the widespread distribution of Jurassic foliated granites in Gromet and Silver's (1987) central and eastern regions, there was no acknowledgment of the existence of pre-Cretaceous granites within the Peninsular Ranges batholith.

The Julian Schist samples consisting of quartz-feldspar-mica semi-pelitic schist are characterized by REE abundances (Fig. 12A) between 10 and 100 times chondrite. The light rare earth elements (LREE) are enriched relative to heavy rare earth elements (HREE), slopes from middle rare earth elements (MREE) to HREE are fairly flat, and there are minor negative Eu anomalies. The patterns show a remarkable similarity to the

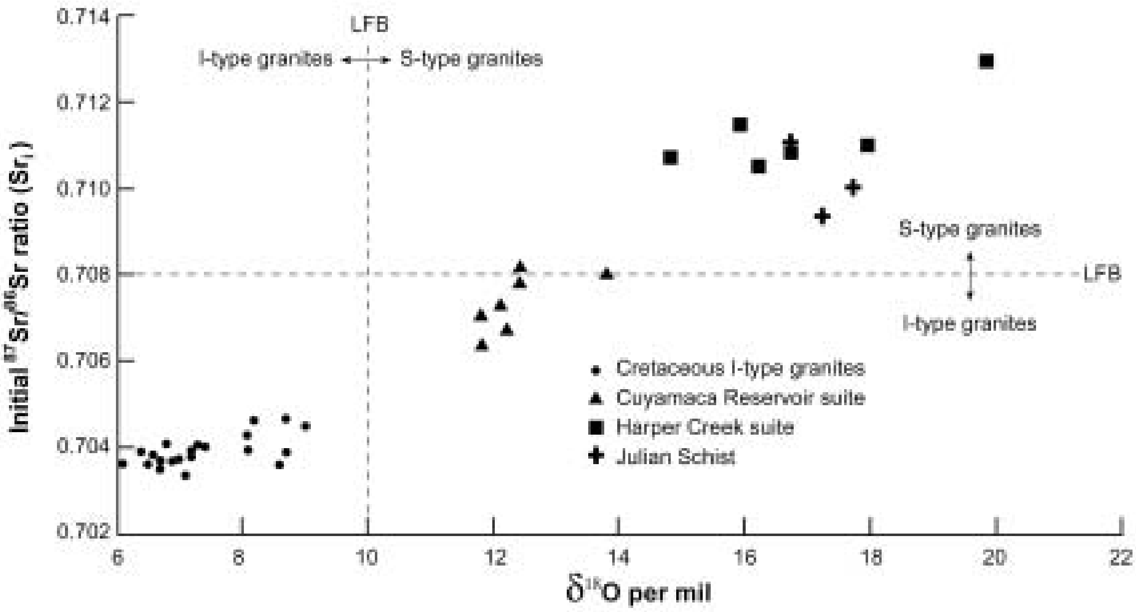

Figure 11. Initial $^{87}Sr/^{86}Sr$ (Sr_i) versus $\delta^{18}O$ diagram. Samples of Harper Creek suite plot in S-type field of Lachlan fold belt granites, eastern Australia (Chappell and White, 1992). Samples of Cuyamaca Reservoir suite are transitional between I-type and S-type fields and plot between isotopically depleted Cretaceous granites from eastern zone of Peninsular Ranges batholith and isotopically evolved Julian Schist metasediments.

TABLE 5. RARE EARTH ELEMENT ANALYSES OF JURASSIC GNEISSIC GRANITES

Sample number	La	Ce	Nd	Sm	Eu	Tb	Yb	Lu
Cuyamaca Reservoir suite								
10-79-H	11.70	24.30	12.70	4.72	0.90	1.04	2.82	0.36
CP-153	25.50	51.50	24.80	6.31	0.88	0.96	2.92	0.43
CP-178	26.49	52.99	24.61	6.27	0.96	0.94	3.06	0.45
D-29	26.52	54.78	27.12	7.39	1.05	1.16	4.23	0.59
J-38	14.96	35.63	23.28	7.20	0.95	1.35	6.38	0.90
ML-30	26.60	54.90	27.00	7.69	1.00	1.34	5.40	0.75
TS-55	22.84	44.63	21.27	5.18	1.10	0.79	2.98	0.43
Harper Creek suite								
10-79-E	34.90	64.30	28.80	6.13	1.16	0.70	2.07	0.28
10-79-G	24.91	48.42	21.22	5.09	1.38	0.66	1.98	0.28
10-79-M	36.67	69.52	31.73	6.58	1.09	0.75	2.37	0.35
J-40	40.80	78.30	32.40	7.16	1.05	0.80	2.79	0.45
MP-35	10.20	19.40	10.00	3.38	1.19	0.79	3.18	0.45
Julian Schist								
5-84-10	39.00	71.60	30.90	6.83	1.34	0.93	3.59	0.53
5-84-SS-1	29.00	56.60	23.50	5.36	1.22	0.74	3.03	0.41
10-79-I	22.14	37.32	17.22	4.11	0.89	0.58	2.48	0.37
10-79-J	38.46	74.06	33.43	7.51	1.50	1.08	3.74	0.56
10-79-L1	32.25	59.90	28.31	5.63	0.98	0.72	2.80	0.43

Note: REE, as parts per million, by neutron activation, USGS, Reston.
Sample locations not mentioned in Tables 3 and 4 as latitude and longitude:

Sample	Lat. N	Long. W
ML-30	32°44′36″	116°27′08″
5-84 10	33°03′47″	116°34′06″
5-84-SS-1	32°58′47″	116°31′28″

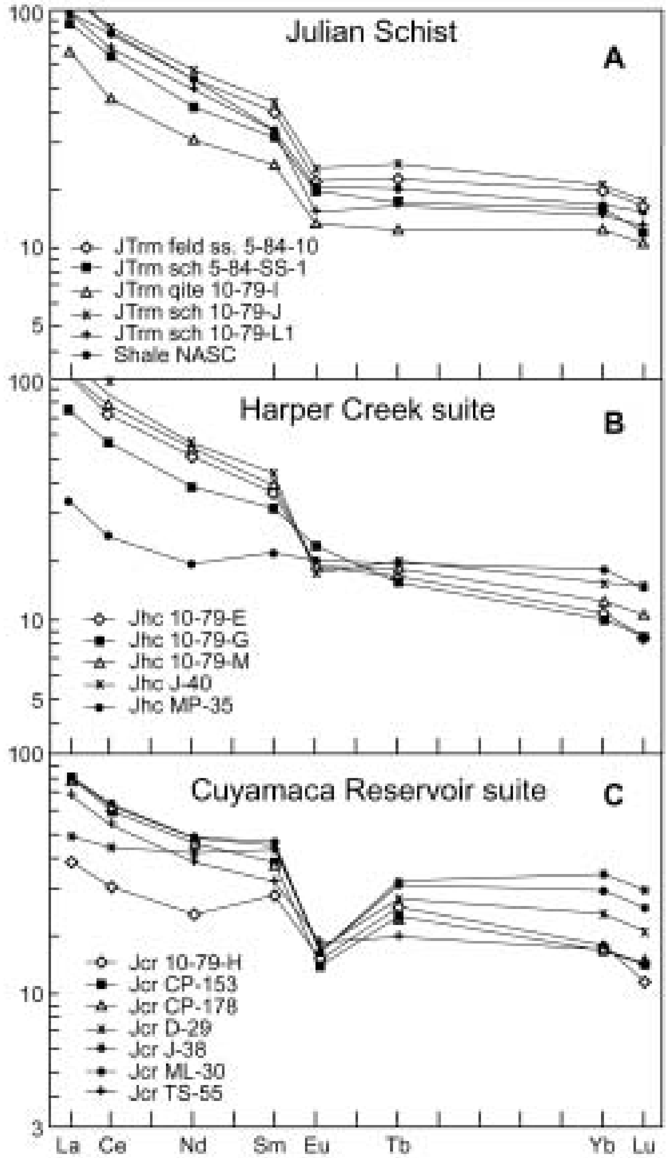

Figure 12. Various REE patterns. A: Julian Schist (quartzo-feldspathic sandstones and semi-pelites). North American shale composite (of Haskin et al., 1966) is added for comparison. B: Harper Creek suite, similar to (A) but with more pronounced negative Eu anomaly. C: Cuyamaca Reservoir suite, transitional between metaluminous and peraluminous compositions. Patterns similar to Cretaceous quartz gabbro of Gromet and Silver (1987) but with distinctly higher $\delta^{18}O$ and Sr_i values.

North American shale composite (of Haskin et al., 1966) and to a group of Australian shales and lithic sandstones from Proterozoic to Triassic in age (Nance and Taylor, 1976).

The Harper Creek suite, apart from one sample, has REE patterns (Fig. 12B) that significantly overlap the field of the Julian Schist and differ only in having slightly more pronounced negative Eu anomalies. This close relationship is to be expected, as the Harper Creek suite shows a complete gradation from igneous plutons containing elongate migmatitic and other lithologically recognizable enclaves of Julian Schist in varying stages of disaggregation to migmatized schist and gneiss. The Harper Creek suite REE patterns are similar to the pattern of the Bundarra S-type suite of the New England batholith, eastern Australia, for which an origin by partial melting of trench-complex metasedimentary rocks has been proposed (Shaw and Flood, 1981).

The Cuyamaca Reservoir suite (Fig. 12C) has flatter REE patterns and moderately negative Eu anomalies relative to the Harper Creek suite. In terms of REE patterns, mineralogy (reduced biotite and subaluminous amphibole), elevated Sr_i, and transitional I- to S-type character, the Cuyamaca Reservoir relates closely to the Hillgrove suite of the New England batholith (Shaw and Flood, 1981). The Cuyamaca Reservoir REE patterns also show similarities to the 50-Ma granodiorites from the eastern Gulf of Alaska (fig. 14 *in* Barker et al., 1992), which is considered to have formed from the partial melting of young trench-fill sediments in a forearc environment. The Alaskan granodiorites also have ASI values greater than 1, Sr_i values between 0.705 and 0.707, and FeO/Fe_20_3 ratios greater than 3:1. Values greater than 3:1 indicate reducing conditions during crystallization, probably with ilmenite as an oxide phase (Todd et al., this volume).

PETROGENESIS

Plate-Tectonic Setting of Jurassic Magmatism

The new U-Pb zircon crystallization ages for the Harper Creek and Cuyamaca Reservoir suites presented in this paper cluster in the Middle Jurassic and include a Jurassic age for an I-type pluton in the Peninsular Ranges batholith (East Mesa pluton). Plate-tectonic models for the early Mesozoic (Triassic–Early Jurassic) Cordilleran convergent margin characterize the forearc region as a zone of extension or transtension within which a system of ophiolite-floored basins and fringing volcanic arcs developed (Busby-Spera, 1988; Saleeby and Busby-Spera, 1992; Busby and Saleeby, 1993; Busby et al., 1998). Available data for the Peninsular Ranges indicate that Late Triassic to Early and Middle(?) Jurassic deposition took place in fault-bounded forearc and/or interarc basins formed during crustal extension and rifting marginal to North America (Germinario, 1993; Gastil, 1993; Busby and Saleeby, 1993, Busby et al., 1998). The depositional basin of the Julian Schist is inferred to have been floored by oceanic or ophiolitic crust (Todd et al., 1988; Saleeby and Busby-Spera, 1992; Grove, 1993; Gastil, 1993; Thomson and Girty, 1994). Thomson and Girty (1994, p. 1114) speculated that the "….plutons of the Cuyamaca–Laguna Mountains shear zone were emplaced into the sedimentary fill of a composite forearc basin which was probably floored by highly extended continental crust on the east and oceanic crust on the west." Titanium enrichment in metabasalt layers within the Julian Schist suggests that these were distal volcanic products of fringing intraoceanic arc. Miogeoclinal metasedimentary rocks of predominantly Paleozoic age in the eastern part of the Peninsular Ranges are considered to have overlain fragments of the North American craton (Gastil, 1993). Schmidt and Paterson (2002) suggest the Cuyamaca–Laguna Mountains shear zone is a segment of a Jurassic-Cretaceous deformation zone that extends along the length of the Peninsular Ranges batholith and which marks a lithospheric discontinuity between an oceanic floored arc to the west and a continental floored arc to the east.

The major source of the Julian Schist detritus apparently lay to the east, where late Proterozoic sialic basement and overlying Precambrian-derived epiclastic rocks were undergoing rapid uplift and unroofing as the result of early Mesozoic sinistral transform faulting along the truncated Paleozoic miogeocline (Saleeby and Busby-Spera, 1992; Busby and Saleeby, 1993; Germinario, 1993). Mafic flows and tuffs were extruded onto submarine fans and mafic feeder dikes and/or sills intruded these sediments. Gastil et al. (1981) and Gastil (1993) reported arc-volcanic strata of Triassic and Jurassic age in Baja California. Multigrain U-Pb analysis of detrital zircons in the Julian Schist and related units in southern California, however, are interpreted as mixtures of late Paleozoic and early Mesozoic grains, together with a blend of Precambrian zircons averaging ca. 1540 Ma (Gastil et al., 1988; Gastil and Girty, 1993).

Along the California Cordilleran margin, continental arc magmatism is interpreted to have persisted at least into the Middle Jurassic, migrating westward during Early and Middle Jurassic time across earlier-accreted Triassic–Early Jurassic forearc terranes and subduction complexes (Saleeby and Busby-Spera, 1992). A profound tectonic change occurred in the late Middle to early Late Jurassic with the development of spreading ocean basins floored by ophiolite west of North America. Fringing island arcs grew along the outer edge of the ophiolite basins in a tectonic regime dominated by extension and dextral transform faulting parallel to the continental margin. By the Late Jurassic, ophiolite development had reached its final stages, the fringing arcs were being accreted to North America, and transform faulting along the continental margin had shifted to sinistral motion (Saleeby and Busby-Spera, 1992; Busby and Saleeby, 1993).

Models for the origin of the Harper Creek and Cuyamaca Reservoir suites depend upon the timing of pre-Cretaceous magmatic activity in the Peninsular Ranges. As discussed above, plutons of the Harper Creek and Cuyamaca Reservoir suites yield Middle and Late Jurassic isotopic ages. The seemingly anomalous location of the Peninsular Ranges Jurassic magmatic "arc," 100–200 km west of the Jurassic Mojave-Sonoran continental-margin arc, apparently reflects its plate-tectonic environment in the upper plate of the Mojave–Sonoran subduction zone. In the late Middle Jurassic to Late Jurassic, the Mojave-Sonoran arc was the site of regional extension or transtension coeval with batholithic emplacement (Tosdal et al., 1989; Busby-Spera et al., 1990; Saleeby and Busby-Spera, 1992; Schermer and Busby, 1994). During this period, subduction was sinistral-oblique and sinistral strike-slip faults cut across the arc. In southeastern California and southwestern Arizona, regional metamorphism was contemporaneous with crustal extension and rifting and with Middle and Late Jurassic granitic intrusion. Tosdal et al. (1989) speculated that an oceanic rift- and transform-fault system existed west of the continental-margin arc in Middle to Late Jurassic time. In this region, fault-bounded basins floored by ophiolite were the site of volcanogenic, oceanic, and terrigenous deposition and of mafic igneous activity while fragments of continental crust and/or rifted island arc segments were transported southward along sinistral transform faults. In southeastern California and southwestern Arizona, regional metamorphism was contemporaneous with crustal extension and rifting and with coeval Middle and Late Jurassic granitic intrusion (Tosdal et al., 1989).

Mixing of Source Rock Components

Variations within the Cretaceous I-type plutonic rocks of the Peninsular Ranges batholith have been explained as the result of mixing of two or more source components (DePaolo, 1981; Hill et al., 1986; Shaw et al., 1986; Gromet and Silver, 1987). Notably, DePaolo (1981), through Sm-Nd and Rb-Sr isotopic studies, proposed a model for the batholith that involved the interaction of depleted mantle magmas with a crustally derived component. The crustal component was considered to have been derived mostly from the mantle in the Mesozoic, but the remaining material was of Precambrian continental derivation. The models of DePaolo (1981) and Gromet and Silver (1987) do not address the origin of the volumetrically significant Jurassic metaluminous and peraluminous granites within the batholith.

In southeastern Australia, Chappell and White (1974, 1992) proposed that magmas of the Lachlan fold belt were derived by partial melting of either an igneous or a sedimentary source component to produce an I-type (metaluminous) or an S-type (peraluminous) partial melt and a residuum (restite). They suggested that most of the variation within an intrusive suite was a result of variable degrees of separation of melt and restite prior to crystallization, with crystal fractionation important only after all restite was removed. In the case of some granites transitional between metaluminous and peraluminous compositions, a single physically mixed-source of infracrustal and supracrustal rocks has been suggested (Shaw and Flood, 1981; Barker et al., 1992). Shaw and Flood (1981) argue that the Uralla plutonic suite of the New England batholith, although metaluminous, has some S-type affinities, such as higher Sr_i than the main I-type suites of the batholith, $\delta^{18}O$ only marginally less than +10 per mil, ilmenite present in some plutons, and high normative hypersthene to diopside ratios. They envisage the source rocks as a physical mixture possibly along interfaces of trench-complex metasedimentary and metaigneous material. Barker et al. (1992) describe three granodiorite plutons intrusive into an accretionary prism in the eastern Gulf of Alaska. The composition and isotopic overlap of the granodiorite and flyschoid sediments, together with modelling and reference to melting experiments of greywacke indicate the granodiorite was derived from the metasediments by high degrees of melting. Basaltic magmas are seen by Barker et al. (1992) as a heat source rather than a melt component mixing with other crustally derived magmas.

As an alternative to the single component model of Chappell and White (1974), two- and three- source components for granites of the Lachlan fold belt have been proposed (Gray, 1984, 1995; Collins, 1996, 1998). Collins (1998) argued that the observed range of both I- and S-type magmas could be generated by a sequence of melting and mixing events involving three source components: (1) mantle generated basalt; (2) melts of older, lower crustal metaigneous rocks; and (3) melts of younger, middle crustal sedimentary rocks. According to Collins (1998), the S-type magmas of the Lachlan fold belt were generated in the middle crust by mixing and hybridization of partly melted Ordovician sedimentary rocks (diatexites) with I-type tonalite magmas derived by melting and mixing of mantle basalts and Cambrian greenstones; in effect, S-type magmas are interpreted as highly contaminated I-type magmas.

Origin of the Harper Creek Suite

Similarities in mineralogy, whole-rock geochemistry and isotopic ratios between the Harper Creek metagranite and metasedimentary rocks of the Julian Schist indicate that the Harper Creek magmas originated by partial melting of the abundant quartzo-feldspathic and semi-pelitic rocks of the Julian Schist and perhaps other, similar flysch units such as the Jacumba sequence. Supporting evidence for such an origin includes field relations, in which Harper Creek plutons are: 1) mantled by anatectic migmatites and gneisses of the Stephenson Peak facies; 2) interlayered and interfolded with migmatites and high-grade metasedimentary rocks; and 3) contain uniformly distributed Julian Schist–type refractory enclaves (restite).

The intermediate chemical compositions and moderately high Sr_i and $\delta^{18}O$ values of the Harper Creek suite reflect the heterogeneous character of the metasedimentary source (Julian Schist and related units). Protoliths of these metasedimentary rocks include Late Proterozoic and Paleozoic crustal sources as well as immature Mesozoic arc materials (Gastil et al., 1988; Gastil and Girty, 1993); the arc materials are, in part, only slightly older than Harper Creek magmatic activity.

According to the paleotectonic reconstruction discussed above, Middle to Late Jurassic Harper Creek magmatism occurred in a setting dominated by extension and transform faulting parallel to the continental margin, which was located well to the east. Peraluminous Harper Creek magmas may have formed at mid-crustal depths in the predominantly metasedimentary early Mesozoic forearc wedge during subduction and rifting, continentward (east) of the site of Cuyamaca Reservoir melting and mixing. Unlike Cuyamaca Reservoir plutons, Harper Creek plutons do not contain metaigneous enclaves or metaluminous phases, which indicates that I-type basaltic/tonalitic magmas did not rise and mingle with this segment of the wedge. Because there are no Harper Creek Sr_i and $\delta^{18}O$ values that differ significantly from the Julian Schist, the contribution of an I-type lower crustal component must have been minor. The relative lack of pegmatites in Harper Creek plutons and the fact that muscovite is largely of subsolidus origin suggest that the magmas were relatively anhydrous and thus able to rise to considerable distances above the site of melting. Accordingly, we consider that the discrete Harper Creek plutons on the western side of the Jurassic belt resulted from crystallization of S-type magma that rose into the upper crust.

Stephenson Peak Facies

We suggest that the Stephenson Peak facies of the Harper Creek suite in the eastern Peninsular Ranges batholith represents the metasedimentary partial melts and restite (diatexites) that remained at or near the mid-crustal source region after more homogeneous bodies of Harper Creek magma separated from restite and rose into the upper crust. In the source region, up to 50% melt plus restite phases may have co-existed in near-equilibrium (White and Chappell, 1977). Low degrees of separation of melt and restite in the Stephenson Peak facies suggest that this material did not rise in the crust more than a few kilometers from the site of melting. The Cooma granodiorite gneiss of the Lachlan fold belt may be an analogous case: Collins (1998) suggested that Cooma-type melts were hydrous and thus incapable of moving far from their middle crustal source region. He further suggested that large migmatitic complexes such as those of Cooma-type in the Lachlan fold belt (and of Stephenson Peak–type in the Peninsular Ranges batholith?) underlie S-type granitoid batholiths to be exposed only after deformation and uplift.

Origin of the Cuyamaca Reservoir Suite

The westernmost Cuyamaca Reservoir plutons have relatively mafic compositions that in thin section suggest disequilibrium among the minerals hornblende, prismatic actinolite, hypersthene, clinopyroxene, and biotite. Preliminary pressure estimates based on hornblende geobarometry from a western pluton, sample SY-91-1, are 3.5 kb (J.M. Hammarstrom, 2002, personal commun.), implying a minimum depth of crystallization of 9–11 km. In the field, Cuyamaca Reservoir plutons contain metaigneous enclaves and interfinger/grade to bodies of I-type hornblende tonalite-quartz diorite (of East Mesa type?), suggestive of magma mingling and/or mixing. Concurrent with a west-to-east change from mafic to felsic compositions, there is a tendency for: 1) inclusions of Julian Schist and mica-rich clots to be more abundant in eastern Cuyamaca Reservoir plutons; 2) a change from amphibole to biotite as the main mafic phase; and 3) contacts between Cuyamaca Reservoir and Harper Creek plutons to become broader and more diffuse. Compositionally, Cuyamaca Reservoir major- and trace-element variation diagrams (Fig. 10) do not follow the systematic trends of fractionation or simple two-component mixing models. The aluminum saturation index varies from metaluminous to moderately peraluminous compositions (Fig. 9) in a manner that cannot be explained by crystal fractionation of an originally homogeneous magma. On a Sr_i versus $\delta^{18}O$ diagram (Fig. 11), the data fall between the more primitive Cretaceous I-type granite suites and the more evolved Harper Creek suite.

The above characteristics suggest an origin for the Cuyamaca Reservoir magmas from a possible three-component source rather than from a single component of arc-type metaigneous and metasedimentary rocks. Three possible source components are: 1) young mantle-derived magma; 2) a mixture of arc-type metaigneous and metasedimentary rocks; and 3) a mid-crustal, predominantly metasedimentary source consisting of Julian Schist and related units.

During Middle to Late Jurassic time, basaltic melts probably underplated and invaded Early Mesozoic-accreted arc-type terranes that comprised the oceanic crust west of the Mojave-Sonoran continental-margin arc and above its subduction zone (Tosdal et al., 1989, and references cited therein). In this offshore zone of rifting and extension, young mantle-derived magma may have been associated with a fringing island arc or may have risen through fracture zones and/or thinned oceanic crust, acting both as a source of heat necessary for partial melting and possibly as a component, mixing with the generated crustal melts. Probable Middle and Late Jurassic fringing island arc-type rocks are reported from several localities in the Continental Borderland of southern California (Gastil et al., 1981; Sorenson, 1988) and metamorphosed Late Jurassic arc rocks underlie Early Cretaceous volcanic rocks unconformably in coastal San Diego County (Balch et al., 1984; Anderson, 1991; C.T. Herzig, 1994, personal commun.). On the Vizcaíno Peninsula, two Late Jurassic source terranes, one exposing I-type hornblende tonalite and a second exposing high-level peraluminous granitic rocks with abundant red-brown biotite (Harper Creek–type?) were contributing detritus to the Eugenia and Perforada Formations by latest Jurassic-earliest Cretaceous time (Kimbrough et al., 1987). These formations overlap an early Mesozoic oceanic arc assemblage.

The older arc-type metaigneous and metasedimentary rocks that formed the oceanic crust outboard of the Mojave-Sonoran arc probably included accreted Triassic and Early Jurassic arc-ophiolites, subduction complexes, and forearc/interarc supracrustal sediments. Partial melting in lower crustal I-type granulitic-amphibolitic rocks, with or without magma mixing of the young mantle-derived magma, could have produced rocks such as the East Mesa suite, the bodies of hornblende tonalite that occur within western Cuyamaca Reservoir plutons, and perhaps other, as yet unrecognized, Jurassic I-type plutons. The Agua Caliente tourmaline-bearing biotite tonalite of Baja California (Schmidt and Paterson, 2002) with an age of 164.3 ± 2.3 Ma could be a further example.

A possible third source-component for the Cuyamaca Reservoir magmas may have been mid-crustal, predominantly metasedimentary assemblages such as the Julian Schist and related units, which, as noted above, are composed of mixtures of Precambrian-derived detritus with early Mesozoic arc material. Mixing of mantle-derived melts and/or lower crustal I-type melts with melts of such isotopically evolved material could produce the Cuyamaca Reservoir magmas with their transitional geochemical and isotopic characteristics.

CONCLUSIONS

Systematic variations within the Jurassic plutonic belt reflect an increasing depth profile from west to east across the Peninsular Ranges batholith (due largely to late Cenozoic tectonism), as well as apparent west-to-east changes in the source regions of the Jurassic magmas. Plutons of the relatively mafic Cuyamaca Reservoir suite are located in the western part of the belt, closer to an inferred western, oceanic lithospheric source. Sharp contacts and little or no evidence of in-situ mixing between those plutons and plutons of the Harper Creek suite in this area suggest that plutons of both suites are relatively high level. Deeper crustal levels and higher proportions of continentally derived source material are suggested in the central part of the belt by more peraluminous compositions of Cuyamaca Reservoir plutons and by broad gradational contacts between them and Harper Creek plutons, which in turn display a greater degree of mixing. In the eastern part of the belt, broad regions appear to be underlain by pre-Cretaceous peraluminous orthogneisses and metasedimentary anatectic migmatites (diatexites). In San Diego County, these rocks comprise the Stephenson Peak facies of the Harper Creek suite. In the Sierra San Pedro Martir area, Baja California, Schmidt and Paterson (2002) have described similar deformation structures as those that occur in the Cuyamaca–Laguna Mountains shear zone. Migmatitic orthogneisses are present in both areas, an extended history of Jurassic-Cretaceous magmatism and deformation is similar in both areas, and significant changes in recorded metamorphic pressures occur across the Cuyamaca–Laguna Mountains shear zone and Sierra San Pedro Martir deformation zones. To account for the strong lateral changes in rheology in the Sierra San Pedro Martir deformation zone, Schmidt and Paterson (2002) reason that extensive heating by sheeted tonalites along the lithospheric discontinuity was sufficient to focus deformation. Although Jurassic tonalite magmas from the East Mesa suite (Todd et al., this volume) could provide some heat in the Cuyamaca–Laguna Mountains shear zone and to the east, it is more probable that the Cuyamaca Reservoir suite and possible volcanic precursors may have been a significant source of heat for the high-grade metamorphism, partial melting of the Julian Schist, and magma mixing within and to the east of the Cuyamaca–Laguna Mountains shear zone.

The geochemical and isotopic characteristics of the Jurassic granites of the Peninsular Ranges batholith suggest that aspects of both mixing and restite models may be required to explain the observed variations within the Harper Creek and Cuyamaca Reservoir suites. In the broadest sense, geochemical variations within both suites can be explained by mixing of recycled Precambrian crustal materials with Mesozoic arc rocks. However, at the present level of erosion, separation of refractory metasedimentary and metavolcanic restite from peraluminous melts during partial fusion of early Mesozoic forearc strata in the middle crust was apparently a dominant process for the Harper Creek magmas. For the Cuyamaca Reservoir suite, partial melting of forearc strata with input of younger basalt/tonalite magma originating in the mantle/lower crust may have been a primary process in its formation. The Middle Jurassic age of an I-type pluton (East Mesa suite) that is spatially associated with a Cuyamaca Reservoir pluton suggests partial melting of early Mesozoic lower crustal metaigneous rocks with input of young mantle basalt as indicated by the presence of synplutonic basaltic dikes and mafic enclaves. Widespread mixing of magmas among these three suites at the present level of erosion is considered minimal, although complex contacts suggest that some interaction occurred.

SPECULATIONS

The above petrogenetic model for the origin of the Peninsular Ranges batholith Jurassic granites relies on offshore Jurassic plate-margin reconstructions for the southwestern Cordillera that are not well documented; therefore, the model is speculative. However, the data available for the Jurassic suites support a west-to-east transition from oceanic-ophiolitic source to a transitional source or one with a significant contribution of North American cratonic detritus. More speculative is the suggested vertical transition from lower crustal to mid-crustal sites of melting. As stated, our model for the Middle to Late Jurassic images the complex earlier Mesozoic history of subduction and arc magmatism, repeated seafloor spreading, and margin-parallel tectonic transport and accretion, although the degree of transpression relative to orthogonal convergence has been questioned (Schmidt and Paterson, 2002). We envision the Peninsular Ranges batholith Jurassic intrusive belt as forming in a setting offshore of the Mojave-Sonoran arc during a period of seafloor spreading, ophiolite development, and growth of transient fringing oceanic arcs. The production of Jurassic magmas ranging from I-type through S-type reflects the complexity of the oceanic-ophiolitic or transitional crust inherited from similar early Mesozoic plate-tectonic events. The presence of craton-derived Mesozoic metasedimentary strata as a source component is not unique to the Peninsular Ranges batholith, although the recognition of S-type granitic plutons apparently is. Based upon different chemical and isotopic compositions of plutons in the Sierra Nevada batholith, Kistler (1990) recognized and named two different lithospheric source regions separated by a tectonic boundary: a western Panthalassan type and an eastern North American type. Isotopic ratios of plutons that intruded Panthalassan lithosphere indicate a greater sedimentary component in the source than that for plutons that intruded North American lithosphere. Plutons that intruded Panthalassan lithosphere were described as "strongly contaminated" by Ague and Brimhall (1988). Indeed, such plutons might be expected to occur wherever subduction/magmatic arc activity results in the recycling of material from ancient cratons.

Jachens et al. (1991) speculated that the western edge of the Jurassic intrusive belt and the coincident gravity-magnetic boundary represent either: 1) a suture between continental crust of North America and an inferred Jurassic magmatic arc built on oceanic crust; or 2) a younger fault that reactivated the suture. We suggest that this structure originated as a Late Jurassic-Early Cretaceous transpressive or orthogonal convergent (Schmidt and Paterson, 2002) deformation zone located outboard of the craton at what was to become the new continental margin. In San Diego County, this deformation zone separated an eastern terrane of mid- to upper-crustal S-type, transitional I- to S-type, and I-type plutons and their high-grade metamorphic wallrocks from the shallow, volcanic part of a Late Jurassic island arc on the west (east-over-west sense of motion). By latest Jurassic-earliest Cretaceous time, both I-type and mixed I-S-type "arcs" were accreted to North America, were emergent, and actively undergoing erosion (Kimbrough et al., 1987). Cretaceous plutons stitched across the boundary, welding the Jurassic arc to North America. The prominent geophysical, geochemical, and isotopic asymmetries of the Cretaceous batholith are a direct reflection of the eastward migration of the Cretaceous magmatic arc across this fundamental boundary (Todd et al., this volume). Moreover, since deformation occurs more readily in quartz- and mica-rich granites (Vernon and Flood, 1988), we believe that the Jurassic belt localized Early Cretaceous contractional strain, leading to the development of the Cuyamaca–Laguna Mountains shear zone.

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