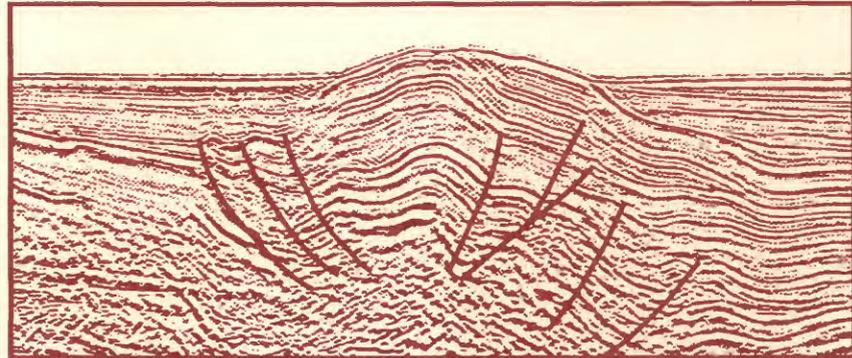# Conglomerates of the Upper Middle Eocene to Lower Miocene Sespe Formation along the Santa Ynez Fault—Implications for the Geologic History of the Eastern Santa Maria Basin Area, California

Reconnaissance Bulk-Rock and Clay Mineralogies of Argillaceous Great Valley and Franciscan Strata, Santa Maria Basin Province, California

Geophysical section offshore Santa Maria basin
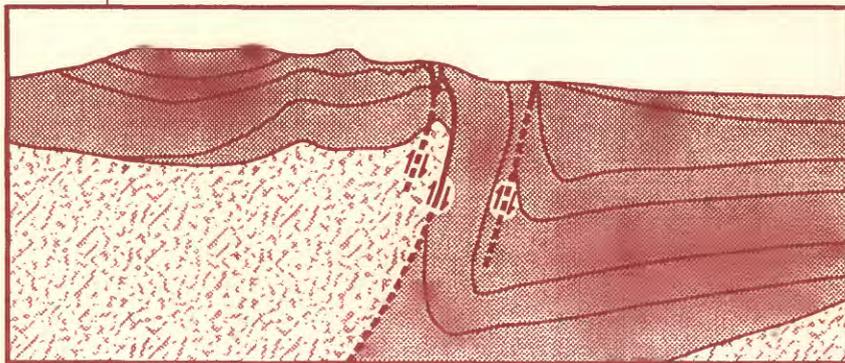
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Conglomerates of the Upper Middle Eocene to Lower Miocene Sespe Formation along the Santa Ynez Fault—Implications for the Geologic History of the Eastern Santa Maria Basin Area, California

By JEFFREY L. HOWARD

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By RICHARD M. POLLASTRO, HUGH MCLEAN, and LAURA L. ZINK

Chapters H and I are issued as a single volume
and are not available separately

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Chapter H

Conglomerates of the Upper Middle Eocene to Lower Miocene Sespe Formation along the Santa Ynez Fault—Implications for the Geologic History of the Eastern Santa Maria Basin Area, California

By JEFFREY L. HOWARD

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Conglomerates of the Upper Middle Eocene to Lower Miocene Sespe Formation along the Santa Ynez Fault—Implications for the Geologic History of the Eastern Santa Maria Basin Area, California

By Jeffrey L. Howard[1]

Abstract

The depositional history of the Sespe Formation was studied using sedimentary facies analysis, clast counts, paleocurrent and clast morphological measurements, and petrographic methods. Sedimentation is interpreted to have occurred mainly as part of two depositional sequences in a coastal-braid-plain forearc-basin setting. Both sequences are present in the Santa Ynez Mountains, whereas only the upper sequence is recognized in the eastern Santa Maria basin. The lower sequence is part of a late middle to late Eocene sedimentary offlap attributed to a high input of terrigenous sediment derived mostly from the Mojave Desert region. This sequence is capped by an intraformational erosional unconformity; much or all of the lower Oligocene section is inferred to be missing. The upper sequence is part of a late Oligocene to early Miocene sedimentary onlap and includes sediment derived from both Franciscan Complex and Mojave Desert sources. It probably represents basin backfilling in response to eustatic sea level rise. The intraformational erosional unconformity and associated upsection change to a Franciscan provenance were not necessarily created entirely by tectonism (Ynezan orogeny), as previously supposed. Regional relations suggest that the effects of a middle Oligocene eustatic sea level drop were also important. Eustatic effects may account for an upsection paleohydrological change from braided to meandering fluvial deposition in the upper depositional sequence. The observed sediment dispersal pattern is inconsistent with active faulting during sedimentation; the Ynezan orogeny can be explained by anticlinal warping of the San Rafael uplift. More recent horizontal offsets along the Santa Ynez and Little Pine Faults are suggested by the juxtaposition of contrasting lithofacies and sediment-source mismatches, but the sense, timing, and magnitude of such movements cannot be evaluated further without additional subsurface information.

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INTRODUCTION

The Santa Maria structural basin in southwestern California is a wedge-shaped region bounded on the northeast by the Little Pine Fault and on the south by the Santa Ynez and Lompoc Faults (fig. 1), therefore, it is much larger than the physiographic basin of the Santa Maria vicinity and includes areas which might be considered parts of either the San Rafael Mountains in the Coast Ranges Province or the Santa Ynez Mountains in the western Transverse Ranges Province. Although there is disagreement over the particular tectonic regime responsible, stratigraphic relations show that the Santa Maria basin formed, along with others like the Los Angeles basin, during middle Miocene time as a result of profound subsidence, faulting, and volcanism (Crowell, 1987). Several earlier diastrophic events also affected the Santa Maria basin; however, because this area was apparently a structural high during early Cenozoic time (Reed and Hollister, 1936; Dibblee, 1950, 1966), this "pre-basin" history is poorly preserved. A more extensive sedimentary record of these events is found in the nearby Santa Ynez Mountains (fig. 1). Nonmarine conglomerates of the upper middle Eocene to lower Miocene Sespe Formation crop out extensively in this area (fig. 2) and are of particular interest because they are well suited for paleogeographic studies. Unfortunately, a clearer understanding of the depositional history and tectonic implications of these deposits is clouded by numerous discrepancies in the literature regarding stratigraphic nomenclature, age and facies relations, and provenance. For example, it is uncertain how and when the Santa Ynez, Lompoc, and Little Pine Faults originated, and their displacement history is poorly defined.

The purpose of this investigation was to gather stratigraphic and sedimentologic data necessary to clarify the geologic history of the Sespe Formation. The area studied includes the eastern Santa Maria basin and the central and eastern Santa Ynez Mountains (figs. 1 and 2). The

conglomeratic parts of six stratigraphic sections of the Sespe Formation were studied intensively by using sedimentary facies analysis, paleocurrent measurements, clast counts, and petrographic methods. The intervening terrane between the measured sections was also studied in reconnaissance. A major aim of the study was to differentiate between first-cycle and multicycle clasts; hence, quartzite and metavolcanic clasts in conglomerates of the Sespe Formation were compared with those in older conglomerates on the basis of shape and lithology. This paper integrates data on the stratigraphy, paleontology, paleoenvironments, petrology, and provenance of the Sespe Formation conglomerates and discusses the implications of these deposits for the depositional and tectonic history of the eastern Santa Maria basin area.

PREVIOUS WORK

The Sespe Formation was named (Watts, 1897; Eldridge and Arnold, 1907) for a type section along lower Sespe Creek about 6 miles north of Fillmore, Calif. (Keroher and others,
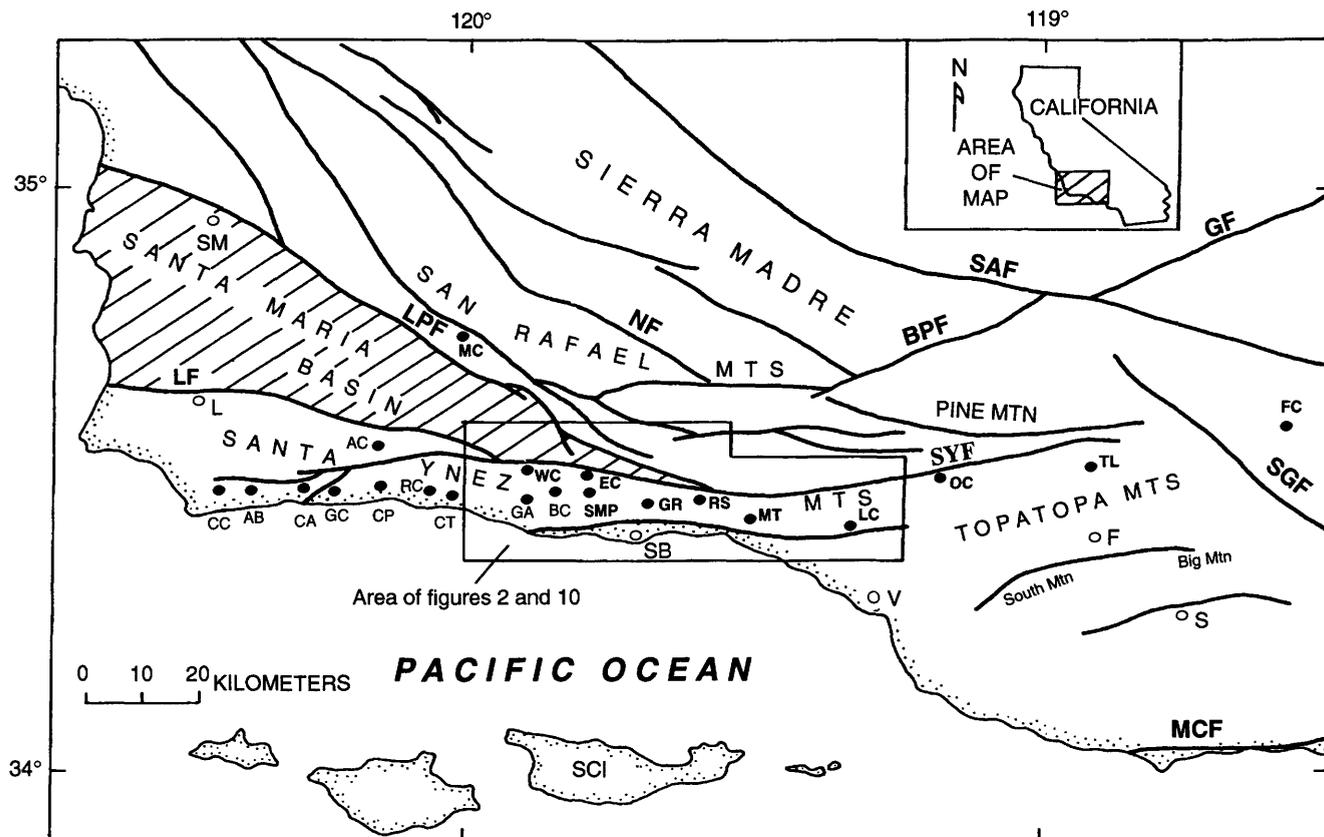
Figure 1. Map showing location of study area (area of figs. 2 and 10) in eastern Santa Maria basin and Santa Ynez Mountains. Location names (cities indicated by open circles): SM, Santa Maria; L, Lompoc; SB, Santa Barbara; V, Ventura; F, Fillmore; S, Simi; SCI, Santa Cruz Island. Fault names: SYF, Santa Ynez Fault; LF, Lompoc Fault; LPF, Little Pine Fault; NF, Nacimiento Fault; BPF, Big Pine Fault; SAF, San Andreas Fault; GF, Garlock Fault; SGF, San Gabriel Fault; MCF, Malibu Coast Fault. Sample localities: Espada Fm.—Manzana Creek (MC); Jalama Fm.—Romero Saddle (RS), Rose Canyon (OC); San Francisquito Fm.—Fish Creek (FC); Cozy Dell Shale—west Camino Cielo (WC); Sespe Fm.—Glen Annie Canyon (GA), east Camino Cielo (EC), San Marcos Pass State Highway 154 (SMP), Gibraltar Road (GR), Montecito (MT), Lake Casitas (LC), Sespe Creek (TL). Other localities: Canada del Cojo (CC); Arroyo Bulito (AB); Canada del Aqua Caliente (CA); Gaviota Canyon (GC); Alisal Creek (AC); Canada de la Posa (CP); Canada del Refugio (RC); Canada del Capitan (CT); Bartlett Canyon (BC).

1966). The type location is in the Topatopa Mountains about 40 km east of the study area (fig. 1, locality TL). This rock unit was subsequently mapped throughout the Ventura basin region (Kew, 1919, 1924), in the Santa Ana (English, 1926) and Santa Monica (Hoots, 1931) Mountains, and in the Santa Maria basin (Dibblee, 1950, 1966). The Sespe Formation was defined as the sequence of sparsely fossiliferous, nonmarine clastic sedimentary rocks lying between fossiliferous marine Eocene and marine Miocene strata. It includes a conspicuous sequence of red beds in some areas but is generally composed of gray or brown strata. Relevant stratigraphic nomenclature[2] is given in figure 3.

The position of the lower contact of the Sespe Formation in the Santa Ynez-Topatopa Mountains has been defined inconsistently because of disagreement over the stratigraphic affinity of a sequence of interbedded gray or brown sandstone and red mudrock physically transitional between the Coldwater Sandstone and the Sespe Formation. Watts' (1897) term Sespe brownstone formation

obviously referred to the overlying sequence of red beds, but Eldridge and Arnold (1907) included these "transitional beds"[3] as part of their Sespe Formation. Kew (1924) criticized this terminology on the basis that Watts' original definition had priority and consequently placed the lower contact at the base of the red bed sequence. Eschner (1969) drew the lower contact in the type area at the top of the uppermost strata containing marine fossils, thereby excluding most of the "transitional beds" from the Sespe Formation. Similar discrepancies are found in the literature pertinent to the parts of the Santa Ynez Mountains covered by this investigation (fig. 4). Thus, the "transitional beds" are equivalent to Lian's (1952) Hay Hill unit, which he later referred to informally as an unnamed lower member of the Sespe Formation (Lian, 1954). Dibblee (1966) also mapped these strata as the basal part of the Sespe Formation but did not discuss them as a separate unit. McCracken (1972) and Lander (1983), however, regarded the "transitional beds" as the uppermost part of the Coldwater Sandstone.

In the type area, the upper contact of the Sespe Formation is difficult to define because the uppermost part of the formation interdigitates with or grades into the overlying Vaqueros Formation. Eschner (1969) placed the lower contact of the Vaqueros Formation at the base of the lowest stratum containing marine mollusks. In the Santa Ynez Mountains and in the eastern Santa Maria basin, where the upper contact of the Sespe Formation contact is usually

[2]Note that important changes in the boundaries between certain chronostratigraphic and biochronologic units have resulted in significant modifications in the age assignments of some rock units from those found in the older literature. The Refugian Stage, once considered to be partly Oligocene, is now regarded as exclusively late Eocene; the Zemorrian Stage, once considered to include much of the Miocene, is now regarded as mostly Oligocene (Bartow, 1991). The Whitneyan, Orellan, and Chadronian stage boundaries have been changed following Swisher and Prothero (1990) and Prothero and Swisher (1992). The Eocene-Oligocene boundary age has been changed from 36.6 Ma to 33.6 Ma following Berggren and others (1992), Prothero and Swisher (1992), and Cande and Kent (1992).

[3]Not to be confused with the "transition stage" shown in figure 3.
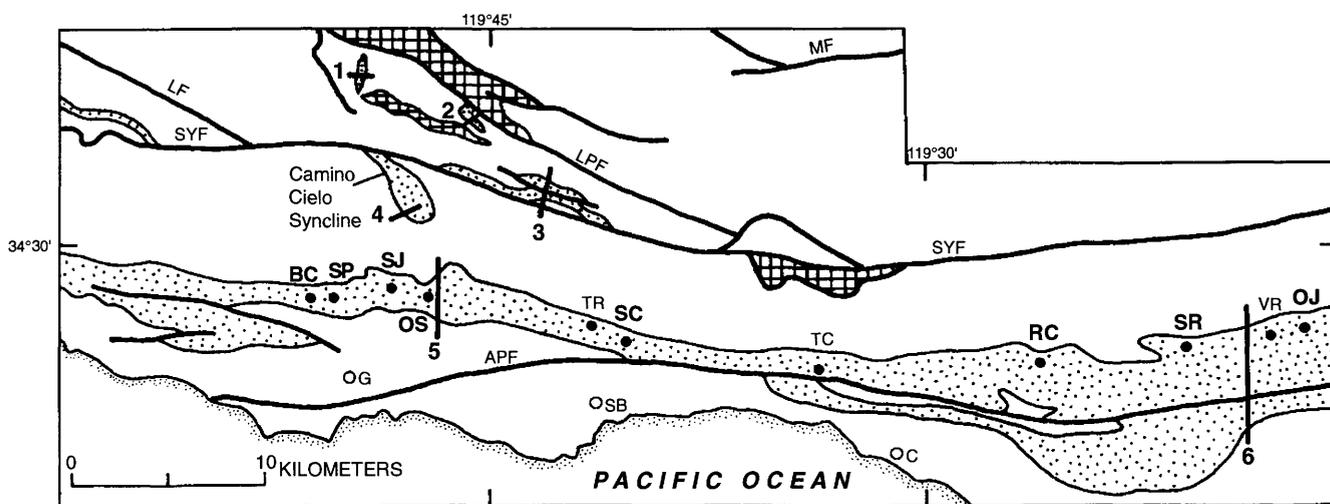
Figure 2. Map of study area showing geographic distribution of Sespe Formation (enclosed stippled areas) and locations of measured sections. Franciscan Complex indicated by box pattern. Location names (cities indicated by open circles): G, Goleta; SB, Santa Barbara; C, Carpinteria. Fault names: SYF, Santa Ynez Fault; LF, Lompoc Fault; LPF, Little Pine Fault; APF, Arroyo Parida Fault; MF, Munson Creek Fault. Localities discussed in text: Bartlett Canyon (BC); San Pedro Canyon (SP); San Jose Canyon (SJ); Old San Marcos Pass Road (OS); Tunnel Road (TR); Sycamore Canyon (SC); Toro Canyon (TC); Rincon Creek (RC); Superior Ridge (SR); Ventura River (VR); Ojai (OJ). Measured sections: 1, Loma Alta; 2, Oso Canyon; 3, Redrock Camp; 4, Camino Cielo; 5, San Marcos Pass Highway 154; 6, Lake Casitas.
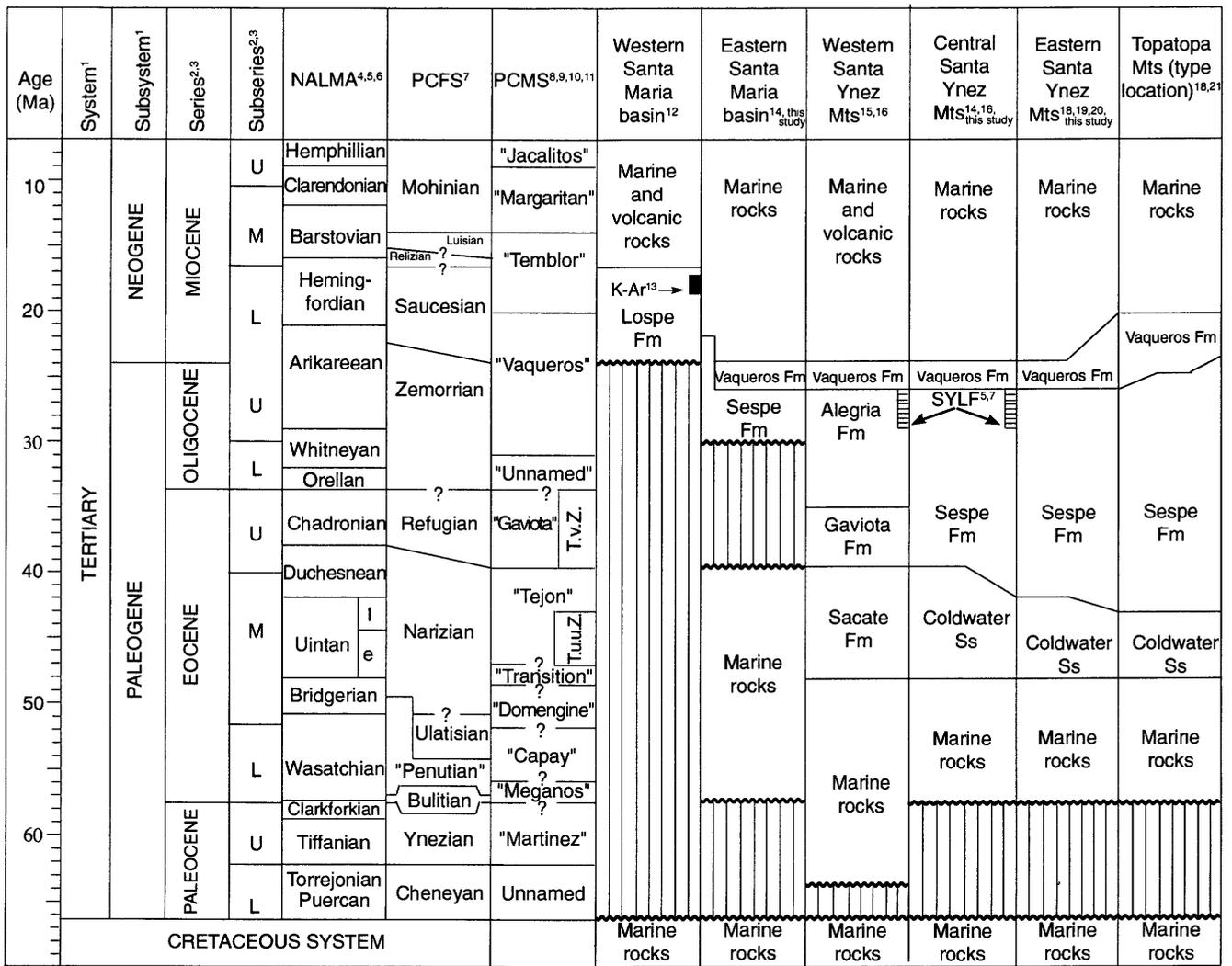
Figure 3. Correlation chart for Santa Maria basin, Santa Ynez Mountains and Topatopa Mountains, Calif. Modified from Weaver and others (1944) and Bishop and Davis (1984). NALMA, North American land mammal ages; e, early; l, late. PCFS, Pacific coast foraminiferal stages; PCMS, California molluscan stages and zones. T. u. u. Z., *Turritella uvasana uvasana* Zone; T. v. Z., *Turritella variata* Zone. Subseries: L, lower; M, middle; U, upper. K-Ar isotopic-age date (black bar). Inferred age range of land vertebrate faunas (ruled bar). SYLF, Santa Ynez local fauna, References: 1, Berggren and others (1985); 2, Berggren and others (1992);

3, Cande and Kent (1992); 4, Woodburne (1987); 5, Swisher and Prothero (1990); 6, Prothero and Swisher (1992); 7, Bartow (1991); 8, Clark and Vokes (1936); 9, Addicott (1970, 1973); 10, Weaver and Kleinpell (1963); 11, Givens and Kennedy (1979); 12, Woodring and Bramlette (1950); 13, Stanley and others (1991); 14, Dibblee (1966); 15, Dibblee (1950); 16, Kleinpell and Weaver (1963); 17, Lander (1983); 18, Bailey (1947); 19, Vedder (1972); 20, Hornaday (1970); 21, Eschner (1969). Note that biochronologic units are used in place of land mammal stages because chronostratigraphic units are not yet fully defined.

Figure 4. Stratigraphic nomenclature used previously in Santa Ynez Mountains, Calif. For facies descriptions see lithofacies in table 1.

sharp, it seems to have been drawn consistently at the top of the uppermost red stratum and (or) at the base of the lowest stratum containing marine macrofossils (Kew, 1924; Bailey, 1947; Dibblee, 1950, 1966; Lian, 1952, 1954; Kleinpell and Weaver, 1963; Edwards, 1971; McCracken, 1972).

The middle part of the Sespe Formation in the Simi area (fig. 1) has yielded terrestrial vertebrate faunas of late Uintan ("Uinta C") and Duchesnean ages (Stock, 1932, 1948; Durham and others, 1954; Golz, 1976; Golz and Lillegraven, 1977; Kelley, 1990). Both faunas were formerly considered late Eocene (see Stock, 1932, 1948), but now they are considered to be late middle and late Eocene in age (fig. 3) (Woodburne, 1987; Swisher and Prothero, 1990; Prothero and Swisher, 1992). Oligocene faunas of Whitneyan and Arikareean ages are also present upsection in the Simi area (Stock, 1948; Savage and others, 1954; Lander, 1983), and an ash bed in the upper part of the Sespe Formation (South Mountain area, fig. 1) has been dated as 27.8±0.28 Ma (late Oligocene) by the K-Ar method (Mason and Swisher, 1988). The uppermost part of the Sespe Formation interfingers with the marine lower Miocene Vaqueros Formation in the Big Mountain area north of Simi (fig. 1) (Taylor, 1983) and in the Santa Monica (Yerkes and Campbell, 1979) and Santa Ana Mountains (Schoellhamer and others, 1981) (not shown in fig. 1). The Vaqueros Formation contains mollusks assignable to the *Turritella inezana inezana* Zone of Loel and Corey (1932) and a foraminiferal fauna indicative of a latest Zemorrian or early Saucesian age (Sonneman, 1956; Edwards, 1971; Yerkes and Campbell, 1979; Blundell, 1981; Schoellhamer and others, 1981; Blake, 1983). This evidence indicates that complete sections of the Sespe Formation range in age from late middle Eocene to early Miocene.

The Sespe Formation overlies the type Coldwater Sandstone in the Topatopa Mountains east of the study area (fig. 1), where Kew (1924) and Eschner (1969) described the contact between them as unconformable. However, if the contact is drawn instead at the base of the "transitional beds," it is conformable (Eldridge and Arnold, 1907). The Coldwater Sandstone contains few fossils in this area, but the presence of *Turritella uvasana uvasana* Conrad (Kew, 1924) assigns it to the basal part of the "Tejon stage" of Clark and Vokes (1936). The age of these fossils, and those of the Uintan vertebrates in the Sespe Formation of the Simi area just to the south, suggests that the lower part of the Sespe Formation in the type area is of late middle Eocene age (fig. 3). In the eastern Santa Ynez Mountains near the Lake Casitas locality (fig. 1), Bailey (1947) reported *T. uvasana uvasana* Conrad from the upper part of the Coldwater Sandstone, but Blaisdell (1953) considered the *T. uvasana uvasana* Zone to lie in the underlying Cozy Dell Shale. A middle Eocene age is consistent with the presence of a "transition stage" molluscan fauna in the

Coldwater Sandstone (Givens, 1974) and the presence of Uintan vertebrates in the lower part of the Sespe Formation (Lindsay, 1968) near Pine Mountain just to the north. In the central Santa Ynez Mountains near the San Marcos Pass locality studied (fig. 1), the Coldwater Sandstone is characterized by a younger molluscan fauna assigned to the *Turritella schencki delaguerrae* Zone, and basal Sespe strata grade laterally westward into the Refugian (late Eocene) Gaviota Formation (Weaver and Kleinpell, 1963; Dibblee, 1966). The above relations indicate that the Coldwater-Sespe contact decreases in age westward across the Santa Ynez-Topatopa Mountains. In the western Santa Ynez Mountains west of the study area, the Coldwater Sandstone grades laterally westward and pinches out into contemporaneous strata of the Narizian Sacate Formation (Weaver and Kleinpell, 1963; Hornaday and Phillips, 1972; O'Brien, 1973), and the Sespe Formation grades laterally westward into the Alegria Formation (Bailey, 1947; Weaver and Kleinpell, 1963; Dibblee, 1966) of Refugian and Zemorrian age (Kleinpell and Weaver, 1963; Weaver and Kleinpell, 1963; Hornaday and Phillips, 1972; O'Brien, 1973).

The Sespe Formation is overlain by the Vaqueros Formation throughout the Santa Ynez Mountains. The sharp contact on the south side of this range (Dibblee, 1950, 1966; Weaver and Kleinpell, 1963) has been interpreted as an erosional disconformity created by the onlap of the Vaqueros Formation (Edwards, 1971; Rigsby, 1989). The Sespe-Vaqueros contact is unconformable in the Santa Ynez Mountains as far east as the Ojai area (fig. 2) near Lake Casitas but gradational and conformable in the Topatopa Mountains to the east (Loel and Corey, 1932; Bailey, 1947; Eschner, 1969; Edwards, 1971). Molluscan macrofossils and foraminifers indicate that the lower part of the Vaqueros Formation is of Zemorrian (late Oligocene) age in the Santa Ynez Mountains and near Pine Mountain; the Zemorrian-Saucesian boundary falls in the overlying marine section (Weaver and Kleinpell, 1963; Dibblee, 1966; Edwards, 1971; Squires and Fritsche, 1978). In the Topatopa Mountains, the Sespe and Vaqueros Formations are conformable or interfinger, and both upper and lower zones of the "Vaqueros stage" are present, suggesting that the Zemorrian-Saucesian boundary falls within the Vaqueros Formation (Loel and Corey, 1932; Eschner, 1969; Edwards, 1971). An early Miocene age for the uppermost Sespe Formation in this area is consistent with that found farther south as noted above. Thus, the Sespe-Vaqueros contact apparently decreases in age eastward across the Santa Ynez-Topatopa Mountains.

In contrast to data from complete sections of the Sespe Formation, the above data show that only the upper Eocene to Oligocene part of the formation is present in the areas of the Santa Ynez Mountains covered by this study. Some confusion has resulted from the fact that the lower contact of the Sespe Formation was defined inconsistently (fig. 4). For example, Lander (1983) reported that the fossil remains

of an oreodont (*Sespia nitida* Leidy) of presumed Oligocene age were found in "basal" strata of the Sespe Formation that were thought to grade laterally into marine Eocene rocks. This problem appears to have been resolved and is discussed below. There are other discrepancies in the literature regarding the age relations between the upper part of the Sespe Formation and the Vaqueros Formation. In particular, the correlation chart of Bishop and Davis (1984) suggests that deposition of the Vaqueros Formation in the Santa Ynez Mountains began about 31 Ma. This appears to be inconsistent with the Arikareean (Oligocene) age of *Sespia nitida* Leidy, which is found in the upper part of the underlying Sespe Formation (fig. 3) in both the central and western Santa Ynez Mountains (Weaver and Kleinpell, 1963; Lander, 1983). This stratigraphic problem will also be discussed below.

The Sespe Formation in the eastern Santa Maria basin is unfossiliferous and its age is poorly constrained because the lower contact is unconformable (Dibblee, 1966; Schussler, 1981). In the Redrock Camp area (fig. 2), it is overlain conformably by the Vaqueros Formation which contains Zemorrian mollusks (Dibblee, 1966). An Sr isotopic age of 20±2 Ma for the Vaqueros Formation in this general vicinity has been reported (Rigsby, 1989); however, the reliability of this date is questionable (Rigsby, oral commun., 1993). In the Oso Canyon and Loma Alta areas (fig. 2), the Sespe Formation is overlain by the early or middle Miocene "Temblor" sandstone of Dibblee (1966). A bentonite bed in the "Temblor" sandstone of this general vicinity correlates with the Tranquillon volcanics farther west (Dibblee, 1966), isotopically dated at 16.5±0.6 to 17.5±1.2 Ma[4] (Turner, 1970). Dibblee (1966) describes the Sespe-"Temblor" contact at Loma Alta and Oso Canyon as gradational and conformable and suggests that the Sespe Formation at these localities may be a nonmarine facies of the Vaqueros Formation. However, his cross section shows the "Temblor" sandstone unconformably overlapping the Rincon, Vaqueros, and Sespe Formations near Oso Canyon. The Sespe Formation in the eastern Santa Maria basin is lithologically similar to the Lospe Formation mapped in the Santa Maria basin farther west (Woodring and Bramlette, 1950; Dibblee, 1950). Isotopic and paleontologic age control there indicates that the type Lospe Formation is entirely of early Miocene age (Stanley and others, 1991). This evidence, and the fact that Sespe-like strata are conformable beneath the "Temblor" sandstone in some places, suggests that the Sespe Formation may be much younger in the eastern Santa Maria basin than in the Santa Ynez Mountains, at least locally.

Cartwright (1928) and Gianella (1928) made early sedimentologic studies and inferred that the Franciscan Complex was the source of certain heavy minerals and conglomerate clasts in the Sespe Formation of the Ojai area (fig. 2). Reed (1929, 1933) also recognized a Franciscan component but noted that the bulk of the detritus composing the Sespe Formation in the Santa Ynez-Topatopa Mountains was of granitic provenance and probably derived from the Mojave Desert region. He criticized earlier interpretations (for example, Kew, 1924; Reinhart, 1928) that the Sespe Formation was composed principally of alluvial fan deposits. He concluded that some alluvial fan sedimentation did occur locally in the eastern Santa Maria basin but that the Sespe Formation was deposited mainly as the delta of a very large river. The interpretation of Bailey (1947) essentially concurs with that of Reed (1929) except that he favors a nearby source for the granitic detritus. Dibblee's (1950, 1966) conclusions also agree with those of Reed (1929), and although he does not say this in his published reports, he considers the Mojave Desert to have been the source for certain types of non-Franciscan clasts (Dibblee, oral commun., 1986). Flemal (1966) resurrected the alluvial fan model of Sespe deposition. Later interpretations by McCracken (1972), Bohannon (1976), Anderson (1980), Black (1982), and Nilsen (1984, 1987) all tend to emphasize the alluvial fan aspects of the conglomeratic strata composing the lower and middle parts of the Sespe Formation. In the alluvial fan scenario, Sespe conglomerate clasts are considered to be locally derived. This implies that where there is a mixture of Franciscan-type clasts with other types such as granite, the former are viewed as being first cycle and the latter as being recycled from older conglomeratic strata. Despite the discrepancies regarding the alluvial fan versus fluvial-deltaic origin of the rudaceous strata, there seems to be general agreement in the literature that the arenaceous strata composing the upper part of the Sespe Formation were deposited by meandering streams.

Stratigraphic relations show that during the early Cenozoic, the region corresponding to the Santa Maria basin was a structural high, which was called the "San Rafael uplift" by Reed and Hollister (1936). The upsection appearance of Franciscan detritus in the Sespe Formation of the Santa Ynez Mountains was interpreted to be the result of tectonism ("Ynezan orogeny"[5] of Dibblee, 1950), which affected the San Rafael uplift during late Eocene and Oligocene time (Reed and Hollister, 1936; Dibblee, 1950, 1966, 1977; Weaver and Kleinpell, 1963). The Ynezan orogeny is usually attributed to domal uplift in the older literature; however, more recent models often show, or at least imply, that Sespe conglomerates were deposited as alluvial fans in a rift basin bounded on the north by an active Santa Ynez Fault (McCracken, 1972; Bohannon, 1976; Anderson, 1980; Black, 1982; Nilsen, 1984, 1987). In contrast, Namson (1987) inferred from structural evi-

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[4]K-Ar isotopic ages corrected using the critical tables of Dalrymple (1979).

[5]Dibblee (1966,1987) refers to this later as the "Ynezian orogeny," however, his original term "Ynezan" is given precedence in this report.

dence and Sespe conglomerate deposition that the Santa Ynez Fault originated as a south-dipping reverse fault during Oligocene time. Early indications were that the San Andreas transform fault system originated between 38 and 29 Ma (Atwater, 1970); hence, it has been suspected that the Ynezan orogeny was the result of plate tectonic interactions (Nilsen, 1984, 1987; Glazner and Loomis, 1984).

Reverse or thrust faulting has been taking place on the major faults of the study area since the Pliocene or Pleistocene (Dibblee, 1950, 1966, 1987; Yerkes and Lee, 1979; Namson and Davis, 1990; Hill, 1990). However, the presence of offset streams and canyons and the geometric orientation of fold axes suggest that a left-lateral component of displacement also has occurred recently on the Santa Ynez Fault (Page and others, 1951; Dibblee, 1950, 1966). The Santa Ynez Fault is believed to splay westward into another left-lateral strand (Santa Ynez River Fault of Sylvester and Darrow, 1979) called the Lompoc Fault in this paper (fig. 1). McCracken (1972) noted that although Franciscan detritus is present in the Sespe Formation as far east as the Ojai area, the easternmost outcrops of Franciscan source rocks are far to the west (fig. 2). Thus, he suggested that about 35 km of left-lateral separation occurred on the Santa Ynez Fault after the Oligocene. McCulloh (1981) has indicated that a laumonite isograd also appears to be offset along this same fault in a left-lateral sense by 37 km. Other estimates of the amount of offset are contradictory (Sylvester and Darrow, 1979).

Plate tectonic models and the recognition of possible large offsets on faults of the San Andreas system have fueled much additional speculation and debate over the possibility of right-lateral offsets on faults of the Coast Ranges, such as the Little Pine Fault, as well as on the Santa Ynez Fault. There is some evidence that the Lompoc-Santa Ynez Fault joins the Hosgri Fault offshore and that large-scale right-lateral separation has occurred on this fault system (Schmitka, 1973; Hall, 1978; Dickinson, 1983). It also has been suggested that the Hosgri-Lompoc-Santa Ynez Fault once joined the San Gabriel Fault to the east and was therefore a through-going element of the San Andreas Fault system (Hall, 1978, 1981). Other workers, however, find evidence for only limited offset on the Hosgri Fault (Sedlock and Hamilton, 1991), and previous mapping shows that the Santa Ynez Fault dies out both to the east and west (Dibblee, 1966, 1987), with negligible horizontal separation at the eastern termination in the Topatopa Mountains (Gordon, 1978).

Paleomagnetic declinations in Miocene rocks of the Santa Ynez Mountains appear to be deflected clockwise by about 90° in contrast with those in the Santa Maria basin, which are essentially undeflected (Hornafius, 1985; Hornafius and others, 1986; Luyendyk and Hornafius, 1987). This contrast in paleomagnetic declinations has been explained by a model in which the entire western Transverse Ranges province was rotated tectonically as a single

rigid block (Luyendyk and others, 1980, 1985). Paleomagnetic data from the Sespe Formation in the Santa Ynez Mountains support this interpretation (Liddicoat, 1990). All of the above studies also documented anomalously flat paleomagnetic inclinations. These inclinations were originally thought to indicate that the rotated single rigid block had also been tectonically transported northward a great distance (Crouch, 1979). In current "suspect terrane" models, however, the Santa Maria basin and Santa Ynez-Topatopa Mountains are part of the Santa Lucia-Orocopia allochthon, which was accreted to southern California near the beginning of the Eocene (Champion and others, 1984, 1986; Howell and others, 1987). The inclination data from Miocene rocks are thought to lack the precision necessary to document large-scale northward transport (Luyendyk and others, 1985; Nicholson and others, 1992).

METHODS OF STUDY

Lithofacies descriptions and paleoenvironmental interpretations in this study are based on characterization of six measured sections (fig. 2; localities 1 through 6). Sedimentary rocks were assigned colors according to the Munsell system and classified using the nomenclature of Folk (1974) and Dunham (1962). Other sedimentologic features were described using the terminology of Ingram (1954), Allen (1963), Campbell (1967), and Reineck and Singh (1973). Paleocurrent measurements were corrected stereographically for tectonic tilt about a single horizontal axis and analyzed statistically using Tukey's chi-square test (Middleton, 1965, 1967). Using this method, statistically significant differences indicate nonrandom distributions of paleocurrent vectors—that is, a preferred orientation for a vector mean. Statistically significant differences are indicated where the calculated χ^2 value exceeds the tabulated value. Use of the terms "alluvial fan," "fan delta," "braid delta," and "braided river" follows that of McPherson and others (1988); "depositional sequence" is used in the sense of Vail and others (1977).

The proportions of different clast types composing conglomerates were established by counting all clasts 3 cm or larger of a given lithology present in a one square meter area of outcrop. Thus, the values are reported as number percentages (Howard, 1993). In counting, the term "metavolcanic" was used collectively for any predominantly aphanitic metaigneous rock, and "granitoid" was used for any species of leucocratic plutonic rock. All crystalline rocks containing visible chlorite, epidote, or garnet were termed "gneiss," regardless of whether foliation was present. A total of 31 conglomerate clasts from the Sespe Formation (appendix), along with selected clasts from older rock units, were examined in thin sections stained for K-feldspar. Modal analysis of granitoid clasts is based on counts of 100 points per slide. The varietal study of metavolcanic clasts is

based on samples of 50 to 150 specimens collected from a 2-m^2 area of outcrop. Mafic clast types, which are commonly partially weathered, were excluded and only the hard silicified clast types were used. Silicified clasts were classified macroscopically as pyroclastic, porphyritic, and aphanitic (felsite) after sawing open each specimen. Green clasts are 5Y or greener; red clasts are 2.5 YR or redder; brown clasts have hues of 5 YR, 10 YR, or 2.5 Y; and gray clasts have a chroma of 2 or less. Those clasts classified as andesite were required to have a chroma of 2 or less and no quartz or K-feldspar phenocrysts. Statistically significant differences in the proportions of metavolcanic clast varieties were assessed by the method of Dixon and Massey (1957). Thus, confidence intervals were calculated to test the null hypothesis that two proportions (p_1 and p_2) are the same according to the relation:

$$p_1 - p_2 \pm t \sqrt{p_1(1-p_1)/n_1 + p_2(1-p_2)/n_2}$$

where t is the tabulated value of Student's t-distribution for $\alpha/2$ level of significance and n_1+n_2-2 degrees of freedom. If such an interval covers zero, the hypothesis is accepted; otherwise, it is rejected (Howard, 1993).

Clast shape measurements were made using only quartzite and metavolcanic clasts; thus the effects of weathering in outcrop or during temporary alluvial storage during deposition (Bradley, 1970) were minimized. At each sampling station, a set of 20 to 40 clasts with intermediate-axis diameters of 32 to 128 mm was obtained by choosing only clasts lacking an obvious planar fabric. The lower size limit of 32 mm was implemented because clasts smaller than this were not always common. Because of the natural size distribution of the conglomerates, the typical sample contained roughly twice as many clasts in the -5ϕ to -6ϕ (32 to 64 mm) grades as in the -6ϕ to -7ϕ (64 to 128 mm) grades. Clasts were sampled intact (using a rock hammer and a chisel) by removing all of a desired type from a 1-m^2 area of outcrop (enlarged slightly as necessary). Submarine fan samples were collected from a single thick, massive, or internally disorganized bed classified as facies A1.1 using the system of Pickering and others (1986), except for that in the Cozy Dell Shale which belongs to facies A1.2. In order to avoid operator bias during measuring, clast samples from different sites were assigned randomized lab numbers and combined. Clast shape distributions were characterized on the basis of oblate-prolate index (OPI) and modified Wentworth roundness (R_w) using the methods of Dobkins and Folk (1970). Thus, $R_w=r/R$, where r is the radius of the sharpest developed corner and R is the radius of the largest inscribed circle. OPI is defined as:

$$10\left(\frac{L-I}{L-S} - 0.5\right) / S/L,$$

where L, I and S are the long, intermediate and short axes, respectively. Axial ratios were calculated according to Zingg (1935) and measured using the long (L), intermediate (I), and short (S) axes as defined by Krumbein (1941). Maximum projection sphericity (ψ_p) was calculated according to the relation:

$$\psi_p = \sqrt[3]{S^2/LI}$$

(Sneed and Folk, 1958). All clasts were broken open prior to disposal to confirm hand specimen identifications. Using the chi-square test, selected samples were analyzed for "goodness of fit" to a normal distribution (see Davis, 1986, for further explanation). Few of these samples were found to have shape index distributions that differed significantly from normality (Howard, 1992). Hence, the variances and means of the shape indices were analyzed using the F-test and t-test, respectively, to test for statistical significance (Davis, 1986). The t-statistic was calculated as:

$$t = \frac{\bar{x}_1 - \bar{x}_2}{S_e},$$

where $\bar{x}_1$ and $\bar{x}_2$ are the means of two samples and S_e is the standard error (see Davis, 1986, for further explanation). If the calculated t exceeds the tabulated value, the hypothesis $\bar{x}_1 = \bar{x}_2$ is rejected and a statistically significant difference is identified. The associated F-statistic was calculated as:

$$F = \frac{s_1^2}{s_2^2},$$

where s_1^2 and s_2^2 are the variances associated with means $\bar{x}_1$ and $\bar{x}_2$. If the calculated value of F exceeds the tabulated value, the hypothesis that the variances are the same is rejected and the t-test cannot be applied. The clast shape terminology used in this paper follows that of Barrett (1980); thus, "shape" refers collectively to all aspects of external morphology, "roundness" is a two-dimensional measure of the sharpness of particle corners, and "form" refers to the overall three-dimensional morphology defined in terms of the relative lengths of particle axes.

Older conglomerate rock units studied (fig. 1) are all of submarine fan origin and include the Jurassic and Cretaceous Espada Formation (McLean and others, 1977; MacKinnon, 1978; Nelson, 1979) in the southern Coast Ranges (locality MC), the Cretaceous Jalama Formation (Rust, 1966; Walker 1975; Krause, 1986) in the Santa Ynez-Topatopa Mountains (localities OC and RS), the Cretaceous and Paleocene San Francisquito Formation

(Sage, 1973; Kooser, 1982) near the San Gabriel Mountains (locality FC), and the Eocene Cozy Dell Shale (Weaver and Kleinpell, 1963) in the Santa Ynez Mountains (locality WC). With the exception of the Cozy Dell Shale, all of the conglomerates sampled are clast supported and lack mud matrix.

SEDIMENTOLOGY OF THE SESPE FORMATION

Stratigraphy and Paleoenvironments

Seven lithofacies are recognized in the Sespe Formation of the study area (table 1). Their stratigraphic distribution in the six measured sections studied is shown schematically in figure 5, except for lithofacies B and E, which are either too thin to show or not present in these particular sections. Lithofacies A and C correspond to the Hay Hill and Calvary units of Lian (1952), respectively. Lithofacies B, D, E, F, and G have not been previously designated.

Lithofacies A is characterized by features found in modern braided stream environments (Miall, 1977), but lateral westward gradation into fossiliferous deltaic deposits of the Gaviota Formation (O'Brien, 1973) suggests deposition as a braid delta. Lithofacies B was observed only in the San Pedro Canyon area (fig. 2) just west of San Marcos Pass where it is interstratified with lithofacies A. These strata resemble wave-worked conglomerates elsewhere (Clifton, 1973; Nemec and Steel, 1984), and similar deposits in the Los Angeles basin area contain clasts with both beach and fluvial shape characteristics suggesting deposition as river mouth bars (Howard, 1992). Lithofacies E is also only found interstratified with lithofacies A and crops out along Old San Marcos Pass Road (fig. 2). It includes crevasse-splay deposits suggesting deposition in a deltaic floodplain or interdistributary lake (Coleman and Prior, 1982). Lithofacies C makes up the bulk of the Sespe Formation in the study area (fig. 5), except in the Camino Cielo area (fig. 2, section 4). It consists of repetitious sequences of normally graded conglomeratic sandstone and lithic-rich sandstone. These strata resemble those deposited episodically by modern braided rivers during waning flood stage (Nemec and Steel, 1984). The matrix-supported conglomerate of lithofacies G and the coarse boulder conglomerate of lithofacies F found in the Camino Cielo section resemble strata of debris flow and proximal braided-fluvial origin, respectively. Such deposits are typically found in modern alluvial fans (Bull, 1977; Nilsen, 1982). The upper part of the Sespe Formation is composed of lithofacies D, except in the Loma Alta section. The association of upward-fining sandstone-siltstone point-bar and levee deposits with overbank mudrock and crevasse-splay deposits in this facies indicates that it was deposited by meandering streams (Allen, 1965; Ray, 1976; Cant, 1982).

Lithologic correlations of the six measured sections (fig. 5) are shown in figure 6. The sections studied are much thinner north of the Santa Ynez Fault and range from only 35 m at Loma Alta to 50 m in Oso Canyon and to 100 m at Redrock Camp. South of the Santa Ynez Fault, the Sespe Formation ranges in thickness from about 400 m at Camino Cielo to 1,100 and 1,700 m, respectively, at the San Marcos Pass (State Highway 154) and Lake Casitas localities. Only in the southern part of the Santa Ynez Mountains is it possible to continuously trace each lithofacies from one place to another. The other sections have been physically isolated from each other by erosion; hence, the correlations shown in figure 6 are based on lithologic similarities. Lithofacies F and G in the Camino Cielo section are correlated with lithofacies C elsewhere on the basis of similarities in conglomerate clast assemblages (discussed below).

Lithofacies A is a distinctive sequence of coarse-grained pinkish-gray sandstone and interbedded red mudrock, which is transitional between typical strata of the Coldwater Sandstone below and red beds of the Sespe Formation above. Lithofacies A sandstones are much coarser grained and tend to be more quartz rich than those of the Coldwater Sandstone. Arenites in the latter are usually buff colored, medium to fine grained, and very feldspathic; they are obviously of granitic provenance and occasionally contain small pebbles of fine-grained granite. Some red mudstone interbeds are present in the upper part of the Coldwater Sandstone below the "transitional beds" of lithofacies A. I have traced lithofacies A continuously from the San Marcos Pass area to Toro Canyon (fig. 7), and Lian (1952) reported that this lithofacies (his Hay Hill unit) eventually pinches out farther to the east near Rincon Creek (fig. 2). Dickinson and Leventhal (1987, 1988) described a somewhat similar uranium-bearing facies in the Superior Ridge area (fig. 2) just west of Lake Casitas; however, these outcrops were not visited during this study. Lithofacies A is definitely absent in the sections examined at Lake Casitas (figs. 5 and 6) and along the Ventura River near Ojai. A sequence of "transitional beds" that includes some red mudstone is present in the Ventura River section (fig. 2), but this is unlike lithofacies A lithologically. These "transitional beds" have been referred to as the upper part of the Coldwater Sandstone by previous workers (for example, Weaver and Kleinpell, 1963; Moser and Frizzell, 1982) who described the Coldwater-Sespe contact as interfingering in this area. The Ventura River section appears to resemble more closely the "transitional beds" of the type section of Sespe Formation, but a thorough investigation of this correlation problem was beyond the scope of this study.

As discussed earlier, the "transitional beds" of lithofacies A were called Coldwater Sandstone by McCracken (1972) despite the fact that they were mapped previously as the Sespe Formation by Lian (1952, 1954) and Dibblee

Table 1. Lithofacies of the Sespe Formation in the eastern Santa Maria basin and vicinity

Lithofacies	Depositional mechanism	Paleoenvironment	Description
A: Trough cross-bedded pebbly sandstone and sandstone	Mainly flash flood, partly tidal deposition	Distal braided fluvial and delta distributary channel	Medium to very coarse grained, thick-bedded, pinkish-gray or light-gray arkose and pebbly arkose. Large-scale trough, festoon and channel-fill crossbedding abundant. Normally graded, thin-bedded sandstone. Minor conglomerate and red or gray shale. Unfossiliferous.
B: Massive sheet conglomerate	Wave worked	River-mouth bar or beachface	Sheet-like, massive to crudely stratified, clast-supported, very densely packed pebble-cobble conglomerate. Interbedded with sheet-like or lenticular, well-sorted, medium to coarse-grained arkose. Strong bedding segregation and lateral continuity. Unfossiliferous.
C: Graded sandstone/conglomerate	Flash flood	Distal braided fluvial	Stacked sheet-like bodies composed of interbedded red, gray, green, or brown, massive, densely packed, clast-supported, lenticular pebble-cobble conglomerate and normally graded conglomeratic arkose and feldspathic litharenite. Scattered paleochannels. Minor mudstone. Very rare terrestrial vertebrate fossils.
D: Upward-fining sandstone/ mudstone	Lateral accretion of point bar/levee deposits. Vertical accretion of overbank crevasse-splay deposits	Meandering fluvial	Repeated upward-fining sequences of red, brown, or gray arkose and feldspathic litharenite and siltstone alternating with mainly red, massive to weakly fissile claystone with scattered sandstone and massive limestone lenticles. Small-scale trough crosbedding; parting lineation and parallel lamination common. Very rare terrestrial vertebrate fossils.
E: Calcareous claystone	Lacustrine delta fill	Interdistributary lake	Massive to weakly fissile, dark reddish-brown calcareous claystone with scattered light gray calcareous arkose and pebbly arkose interbeds. Unfossiliferous.
F: Boulder conglomerate	Flash flood	Proximal braided fluvial or alluvial fan	Interbedded massive or crudely stratified and normally graded, very thick bedded,very poorly sorted, clast-supported, dark greenish-brown bouldery pebble-cobble conglomerate with scattered thin, very lenticular reddish-brown siltstone and pebbly mudstone. Unfossiliferous.
G: Pebbly mudstone	Debris flow	Proximal braided fluvial or alluvial fan	Interbedded massive and normally graded thin, sheet-like to lenticular, reddish- or greenish-brown, matrix-supported pebbly mudstone and mudstone. Locally cobbly and bouldery. Unfossiliferous.
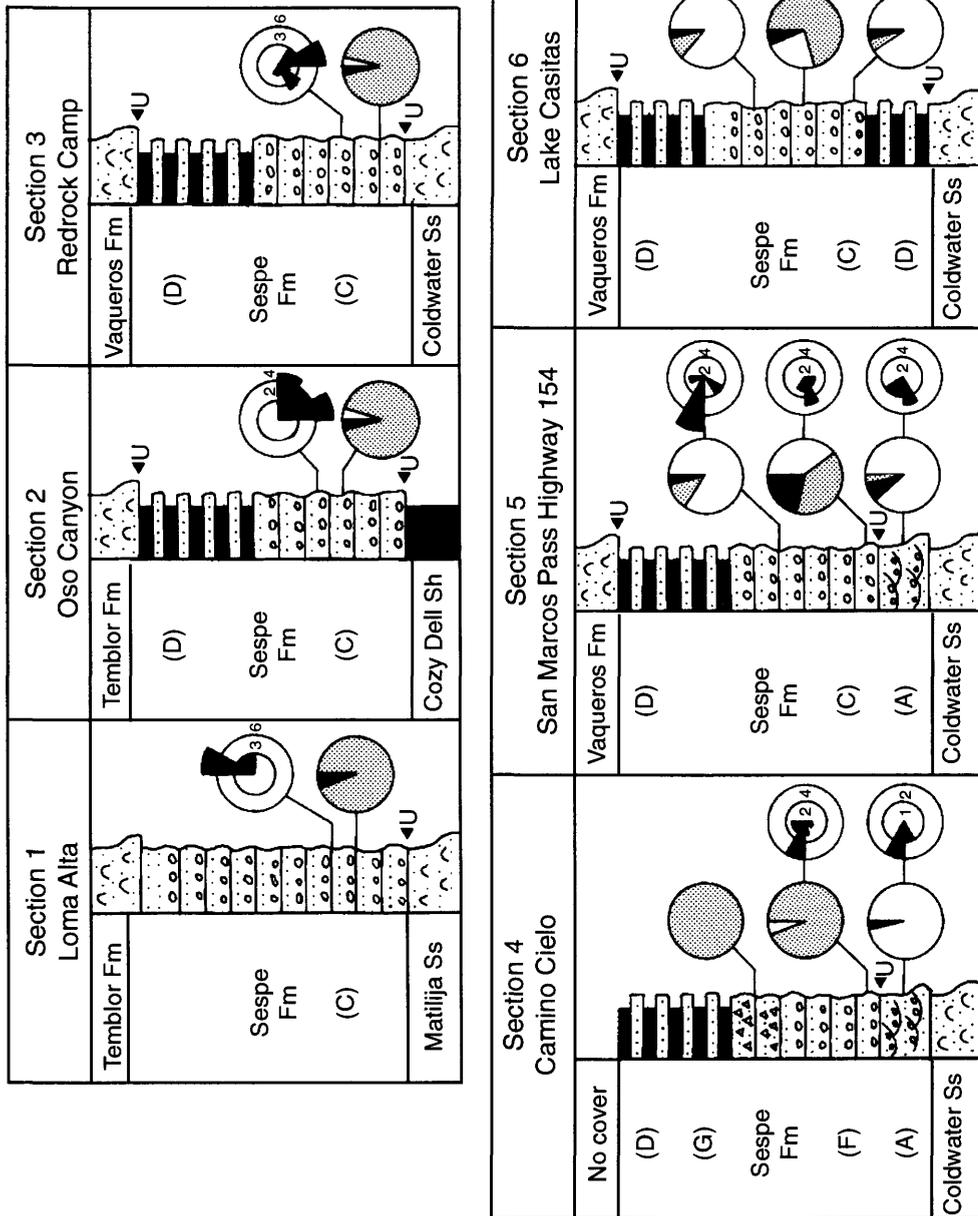
Figure 5. Schematic stratigraphic sections (not to scale) of Sespe Formation in study area. See figure 2 for section locations; lithofacies types (letters in parentheses) are described in table 1. Paleocurrent data are selected from tables 2 and 3. Pie diagrams indicate relative proportions of conglomerate clast assemblages. Rose diagrams indicate number of paleocurrent measurements within a directional interval; modern magnetic north is at top of diagram, and numbers refer to radii of inner and outer circles in number of paleocurrent measurements.
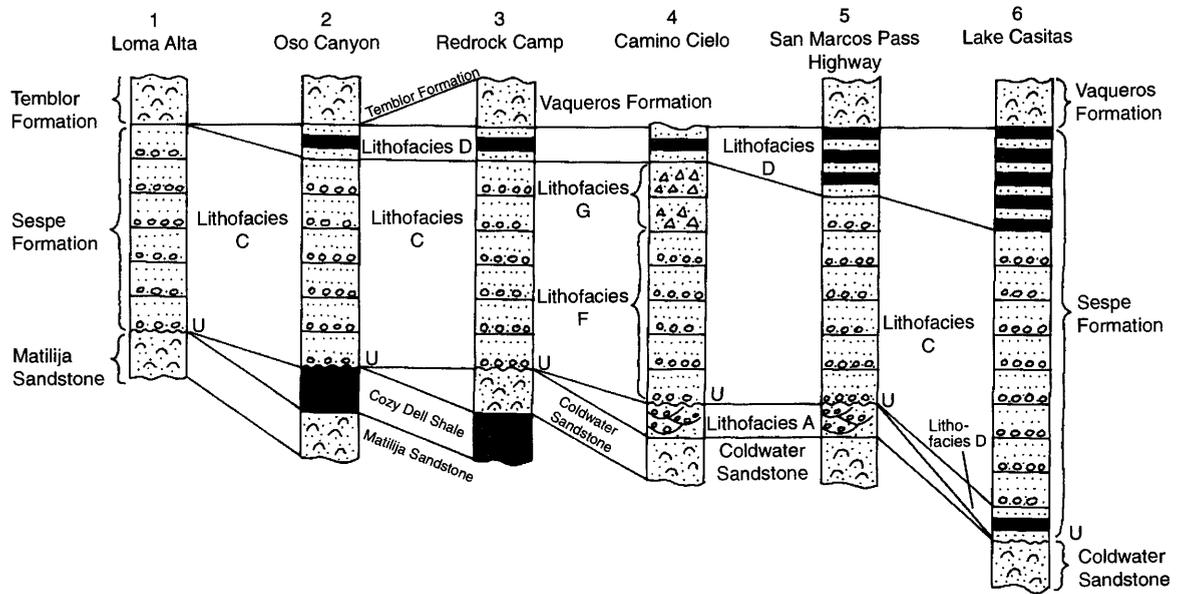
Figure 6. Lithologic correlations of Sespe Formation in study area (not drawn to scale). Lithologic symbols shown in figure 5. See figure 2 for section locations and table 1 for lithofacies description.
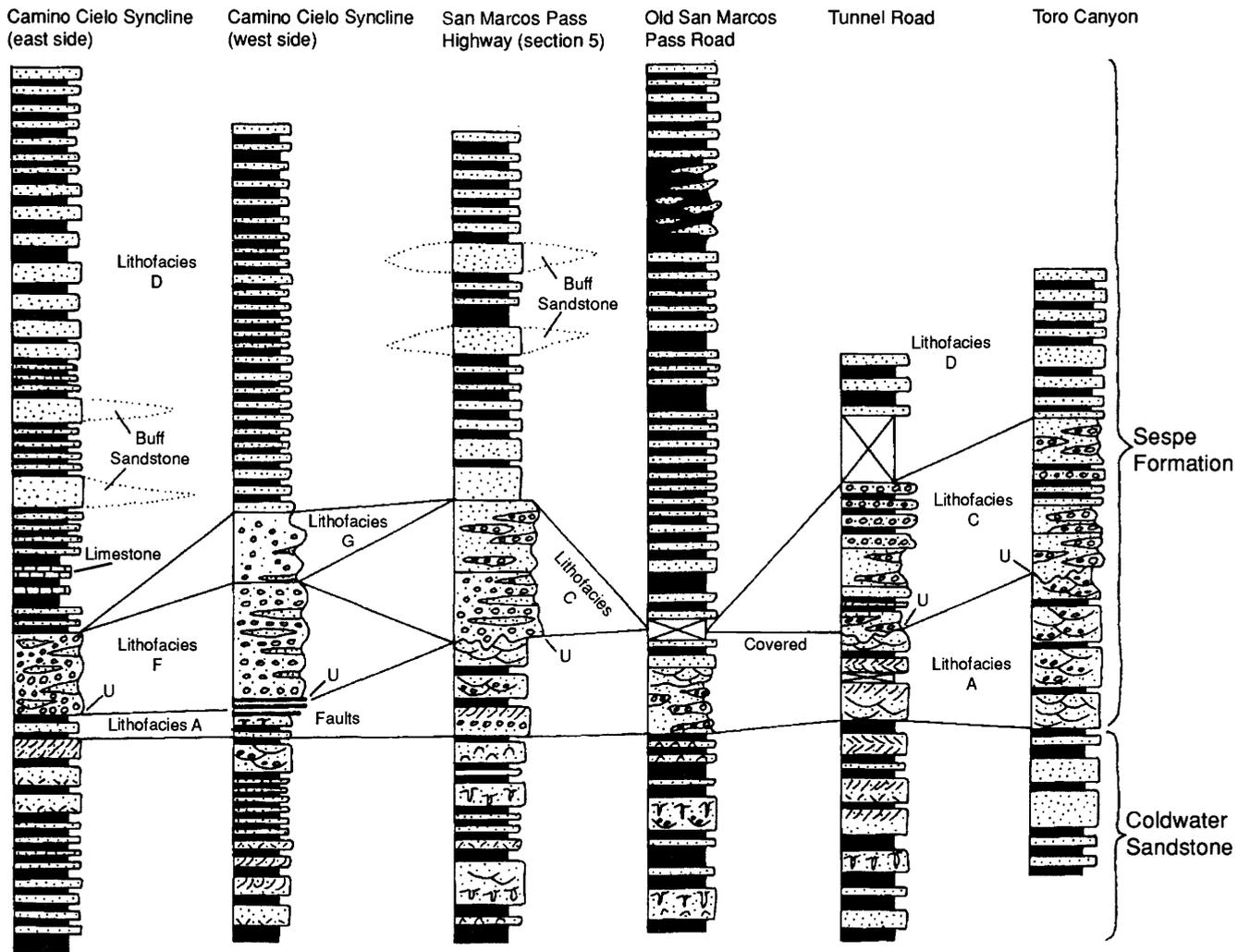
Figure 7. Schematic stratigraphic sections (not to scale) of Sespe Formation in southern Santa Ynez Mountains showing lithologic correlations. See figures 1 and 2 for site locations; lithofacies are described in table 1.
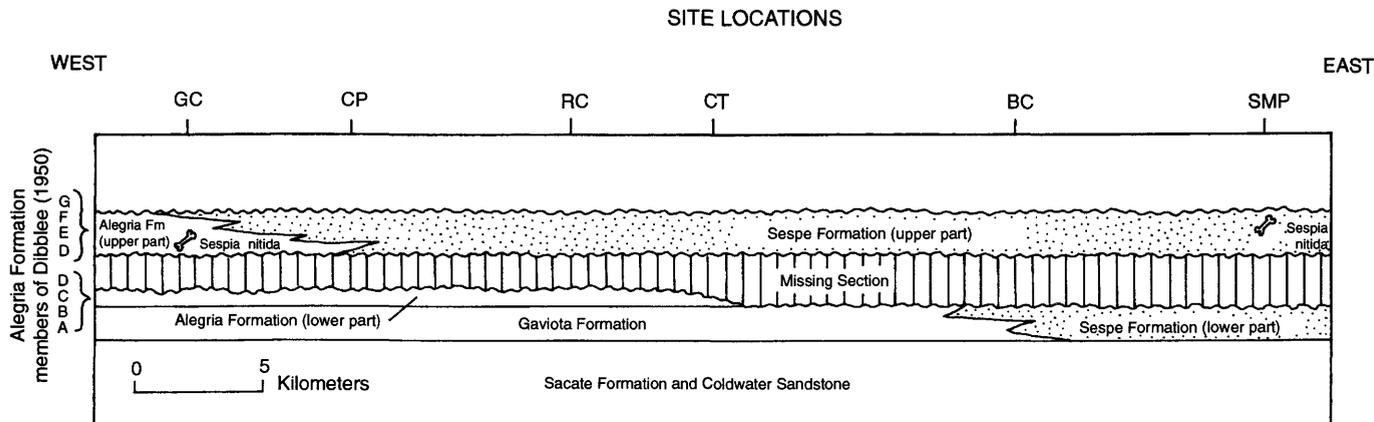
Figure 8. Facies relations between Sespe Formation and laterally equivalent marine units in the central and western Santa Ynez Mountains, Calif. Modified from Hornaday and Phillips (1972), as discussed in text. See figure 1 for site locations: GC, Gaviota Canyon; CP, Canada de la Posa; RC, Canada del Refugio; CT, Canada del Capitan; BC, Bartlett Canyon; SMP, San Marcos Pass State Highway 154.

Explanation

Covered section

Conglomerate

Sandstone

Conglomeratic sandstone

Trough crossbedded sandstone

Mudrock

Rip-up clasts

Shell debris

Burrows

Planar Crossbedding

– – –   Inferred contact

Lens-shaped body

∪   Unconformity
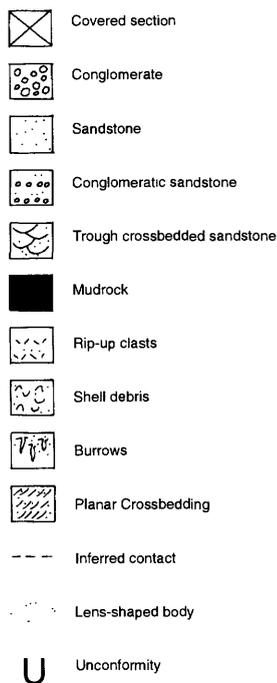
Figure 7.—Continued.

(1966). Further justifications for including lithofacies A as part of the Sespe Formation are the facts that (1) strata of similar lithology are found elsewhere in the Sespe Formation (Howard, 1987, 1989) but have not been recognized in the underlying Coldwater Sandstone (Dibblee, 1966; Vedder, 1972; McCracken 1972) and (2) these strata grade laterally westward into the Gaviota Formation, which overlies and is younger than the Coldwater Sandstone (Weaver and Kleinpell, 1963; Dibblee, 1966). Defining the lower contact at the base of lithofacies A includes many strata in the Sespe Formation that are not red but excludes those red beds found in the upper part of the Coldwater Sandstone. This is perhaps a violation of the priority of Watts' (1897) original usage of the unit as the Sespe brownstone. However, stratigraphers have agreed that priority should not prevent a more exact lithologic definition if the original definition is not everywhere applicable and if a change in the terminology does not require a new geographic term to be established (North American Commission on Stratigraphic Nomenclature, 1983). The presence or absence of red beds is an invalid criterion for a formal definition of the lower contact of the Sespe Formation because red beds are not a characteristic feature of the Sespe Formation in all locations where it has been mapped. Additional studies of the "transitional beds" at the type locality are needed to determine what lithologic criteria should be used to formally define the Sespe Formation.

In the Santa Ynez Mountains west of San Marcos Pass (fig. 1), the Sespe Formation grades laterally westward into the marine Gaviota and Alegria Formations (fig. 8). Nonmarine strata of the basal part of the Sespe Formation can be walked bed by bed across Bartlett Canyon into laterally equivalent marine rocks of the Gaviota Formation (Weaver and Kleinpell, 1963; Dibblee, 1950, 1966; Weaver, oral commun., 1985). West of Canada de la Posa (fig. 1), progressively higher beds of the Sespe Formation grade laterally westward into the marine Alegria Formation until the Sespe Formation pinches out a short distance west of Gaviota Canyon (Dibblee, 1950; Weaver and Kleinpell, 1963). In the western Santa Ynez Mountains west of Arroyo Bulito (fig. 1), the Vaqueros Formation rests unconformably on the Alegria Formation. From this point, the Vaqueros Formation overlaps the Alegria and Sespe Formations eastward

across the Santa Ynez Mountains. Thus, lateral facies changes involving marine and nonmarine strata and age data already discussed show that the upper Eocene to lower Miocene section of the Santa Ynez Mountains is a sedimentary offlap-onlap sequence. These relations have been known for many years, but it usually has been assumed that sedimentation was more or less continuous.

However, during the recuration of the California Institute of Technology collection of vertebrate fossils from the Sespe Formation by the Los Angeles County Museum, it was discovered that several specimens had been found in the Santa Ynez Mountains which were not described in the literature (Lander, oral commun., 1986). Hence, Lander (1983) reported that a fossil oreodont (*Sespia nitida* Leidy) was found at two locations. One locality is in the upper part of the Sespe Formation (lithofacies D of this report) at Sycamore Canyon north of Santa Barbara (fig. 2) and the other is at the base of Lian's Calvary unit (lithofacies C of this report) along Maria Ygnacio Creek (fig. 1, locality SMP). The latter specimen was apparently collected by D.W. Weaver and D.E. Savage (Weaver, oral commun., 1985) along San Marcos Pass highway (Lander, oral commun., 1986). This oreodont species is generally thought to be of Arikareean (Oligocene) age, and isotopic dates from rocks elsewhere in the western United States suggest an age of 26 to 30 Ma (Lander, 1983; Swisher and Prothero, 1990). These data indicate that at San Marcos Pass the oreodont-bearing red beds of lithofacies C are of late Oligocene age and that these red beds lie unconformably on pinkish-gray strata of lithofacies A that grade laterally westward into the Gaviota Formation of late Eocene age (fig. 8). The unconformity separating these two units (figs. 5 and 6) is, therefore, a significant hiatus representing much or all of early Oligocene time.

As noted earlier, confusion over the position of the boundary between the Coldwater Sandstone and Sespe Formation has resulted in errors in the interpretation of age relations. Thus, by drawing the lower boundary of the Sespe Formation at the base of the red bed interval instead of at the base of the "transitional beds," Lander (1983) misinterpreted the oreodont-bearing red beds to be of late Eocene age because "basal Sespe strata" had been reported to grade laterally westward into the Gaviota Formation of Refugian age. A reconnaissance of San Jose, San Pedro, and Bartlett Canyons (fig. 2) during this study showed that it is only the "transitional beds" (lithofacies A) that grade laterally westward into the Gaviota Formation. Strata directly above the "transitional beds" in these canyons are physically correlative with the red bed conglomerates farther east at San Marcos Pass that contain the late Oligocene oreodont remains.

Sespia nitida Leidy also has been discovered in the Alegria Formation farther west in Gaviota Canyon (fig. 1). Weaver and Kleinpell (1963) indicated that the fossil was found in a lower Zemorrian time-rock (chronostratigraphic) interval, presumably corresponding with members D through F of Dibblee (1950). Lander (1983), however, apparently miscorrelated the oreodont-bearing beds with members A and B of Refugian (late Eocene) age. Dibblee (1950) and Weaver and Kleinpell (1963) noted that the lower members of the Alegria Formation (up through the basal part of member D) contain a Refugian molluscan fauna identical to that in the Gaviota Formation. Lander (1983) recognized further that planktonic foraminifers belonging to zone P17 and nannoplankton belonging to zones NP 19 and 20 assign a late Eocene age to Alegria strata just below member C. These data suggest that the same hiatus found at San Marcos Pass within the Sespe Formation is also present within the Alegria Formation at Gaviota Canyon. If so, some Alegria strata of late Oligocene age must lie unconformably on other Alegria strata of late Eocene age (fig. 8). This modified view of lateral facies relations is interesting because it suggests that the Sespe Formation and its marine lateral equivalents were deposited rapidly during two distinct episodes rather than as one long continuous event spanning all of Oligocene time as previously supposed.

Paleocurrent Analysis

Paleocurrent data for braided-stream deposits of the Sespe Formation north of the Santa Ynez Fault are shown in table 2 and figure 5. The modes (class intervals containing the most observations) and calculated vector means suggest that paleocurrents flowed primarily toward the west and south, except at Loma Alta where paleoflow was directed northward. The chi-square test indicates that one vector mean from the lower part of the Sespe Formation at Redrock Camp is not significant, a determination which suggests that no preferred orientation is present. Nevertheless, the data from Oso Canyon and Redrock Camp seem to show an upsection shift from west to south and perhaps a change in paleoslope direction over time. South of the Santa Ynez Fault, paleocurrent modes and vector means (table 3; fig. 5) are directed mainly toward the west and southwest, but a southward component of transport is also present. The data for lithofacies A at the Camino Cielo locality, and for lithofacies C at the San Marcos Pass highway locality fail the chi-square test for statistical significance. This failure suggests that a preferred orientation is not present; hence, these data should be regarded as inconclusive. At the San Marcos Pass highway locality, clasts are very prolate, making imbrication difficult to find, but there is a well defined east-west-trending paleochannel cut on top of lithofacies A. Similar features are common at the Lake Casitas locality.

Conglomerate Clast Assemblages

The clast types characterizing Sespe conglomerates in the study area are classified into three assemblages (table 4) that are interpreted as being genetically distinct with regard to provenance. Assemblage 1 includes two varieties of chert

Table 2. Paleocurrent data for lithofacies C of the Sespe Formation in the eastern Santa Maria basin

[All data are for pebble imbrication relative to modern magnetic north and corrected for bedding tilt. χ^2, test statistic; α, level of significance. See figure 2 for site locations]

Parameter	Loma Alta	Oso Canyon Lower[1]	Oso Canyon Upper	Redrock Camp Lower	Redrock Camp Lower	Redrock Camp Upper
Arithmetic mean(°)	24.1	254.6	135.1	286.5	233.6	184.2
Standard deviation(°)	20.7	50.1	22.2	32.5	61.6	46.3
Vector mean(°)	27.2	256.4	137.1	288.2	229.0	187.5
χ^2 (calculated)	[3]21.3	[2]7.7	[3]24.5	[3]11.6	1.8	[3]10.9
Modal class(°)	0-29	240-269	150-179	240-269	240-269	150-179
No. of measurements	12	7	15	8	8	20

[1]Refers to stratigraphic position.
[2]Significant difference (α=0.05; tabulated χ^2=5.99).
[3]Highly significant difference (α=0.01; tabulated χ^2=9.21).

Table 3. Paleocurrent data for the Sespe Formation in the Santa Ynez Mountains

[All data are for pebble imbrication relative to modern magnetic north and corrected for bedding tilt. χ^2, test statistic; α, level of significance. See Figure 2 for site locations]

Parameter	East[1] (F)[2]	West (A)	Camino Cielo West (F)	Camino Cielo Lower[3] (A)	San Marcos Pass Highway Lower (C)	San Marcos Pass Highway Middle (C)	San Marcos Pass Highway Upper (C)	Lake Casitas Lower (C)	Lake Casitas Middle (C)	Lake Casitas Upper (C)
Arithmetic mean (°)	179.5	272.7	279.1	277.7	284.7	261.3	255.0	252.5	271.4	214.4
Standard deviation (°)	38.3	21.2	40.3	29.6	98.7	41.8	83.0	43.5	36.0	19.4
Vector mean (°)	182.2	275.0	283.8	280.5	319.2	250.1	285.0	251.5	272.3	217.2
χ^2 (calculated)	[4]8.5	5.9	[5]9.6	[5]11.6	[5]10.5	5.5	[5]9.8	[5]13.1	[5]18.3	[5]32.6
Modal interval (°)	150-179	270-299	270-299	240-269	330-359	240-269	270-299	210-239	240-269	210-239
No. of measurements	6	3	8	7	14	7	10	10	14	18

[1] Indicates east or west limb of syncline (see fig. 2).
[2] Indicates lithofacies type (see table 1).
[3] Indicates stratigraphic position.
[4] Significant difference (α=0.05; tabulated χ^2=5.99).
[5] Highly significant difference (α=0.01; tabulated χ^2=9.21).

clasts (not differentiated in table 4). The most abundant type ranges in size from granule to boulder and is typically red, green, or gray, angular, and very thin bedded or laminated. The other type is black, well rounded, pebble sized, and rare. The lithic-arenite clasts resemble graywacke in hand specimen but contain less than 15 percent matrix. They apparently lack detrital K-feldspar, although only one clast was studied in thin section. The clasts counted as metabasalt are mainly aphanitic, but some of vesicular or diabasic texture are also present. Two of the three examined in thin section have basaltic texture and contain the greenschist-facies minerals chlorite, albite, epidote, and actinolite. The other specimen has ophitic texture and is of basaltic composition. The typical

clasts of ultramafic rock are partially serpentinized and found in association with jasper-calcite rock, glaucophane schist or chlorite schist, and vein quartz. A clast of leucogabbro examined in thin section contains uralitic hornblende and plagioclase with the composition of andesine-labradorite (An_{35-55}). Other gabbro-like hand specimens are apparently of dioritic composition, and some are quartz or garnet bearing.

Assemblage 2 includes several lithologic varieties of feldspathic sandstone and limestone, none of which were examined in thin section. The sandstone clasts are mainly buff-colored arkosic arenite containing biotite and (or) muscovite and, locally, whole or broken molluscan fossils. Clasts of subarkosic arenite and quartz arenite are also

Table 4. Percentages of clasts in Sespe conglomerates of the eastern Santa Maria basin area

[Tr, trace amount—<1.0 percent; facies are lithofacies in table 1. See figure 2 for site locations]

				Camino Cielo			San Marcos Pass Highway			Lake Casita		
Clast type	Loma Alta	Oso Canyon	Redrock Camp	Facies A	Facies F	Facies G	Facies A	Facies C	Facies C	Facies C	Facies C	Facies C
Assemblage 1												
Chert	2	46	3	0	2	10	0	32	10	4	45	8
Lithic arenite	98	29	96	0	0	0	0	4	1	1	25	9
Metabasalt	0	3	1	0	6	13	0	tr	0	0	0	0
Ultramafic	tr	0	0	0	tr	0	0	tr	0	tr	tr	tr
Leucogabbro	0	tr	0	0	76	77	0	0	0	0	0	0
Glaucophane schist	tr	1	0	0	0	0	0	0	0	0	0	0
Jasper-calcite rock	tr	9	tr	0	0	0	0	3	0	0	tr	1
Vesicular basalt	0	0	0	3	0	0	0	0	0	0	0	0
Quartz diorite	0	0	0	0	2	0	0	0	0	0	0	0
Vein quartz	tr	tr	tr	0	0	0	0	0	0	0	0	0
Assemblage 2												
Arkosic arenite	0	2	tr	3	0	0	10	9	1	tr	4	4
Limestone	tr	1	tr	0	0	0	1	12	2	2	2	8
Assemblage 3												
Quartzite	0	9	tr	45	8	0	50	27	42	33	4	17
Felsic metaporphyry	0	tr	tr	28	0	0	22	4	23	18	5	16
Felsic metatuff	0	0	tr	3	0	0	7	tr	11	4	tr	3
Metaandesite	0	tr	tr	0	0	0	0	0	0	tr	tr	tr
Granitoid	0	tr	tr	18	6	0	10	2	3	10	1	10
Gneiss	0	tr	tr	tr	0	0	0	6	7	28	14	27
Sample size	122	186	117	122	143	244	184	404	296	227	113	162

included. The limestone clasts are mainly grayish white, very fine grained with a saccharoidal texture, massive, and unfossiliferous. One specimen was found that was partially dolomitized and thinly laminated with black chert. The other common clast type counted as limestone is medium- to coarse-grained calcarenite containing sand- and granule-sized lithic fragments of red and green chert and jasper.

The quartzite clasts in assemblage 3 are mainly gray and brown; however, distinctive maroon and black varieties are also common, along with a rare ochre-colored type. They range collectively from granular to vitreous and porcelaneous and from very fine grained to pebbly. Only one specimen was examined in thin section. This is a black, partially recrystallized specularitic variety that contains irregularly distributed clots of chlorite. The granitoid clasts in assemblage 3 vary widely in grain size and are typically biotite bearing and of midcrustal origin; granulitic types were not observed. One garnet-bearing lithology counted

as gneiss in the field was found in thin section to have a granitoid texture; thus of the nine clasts examined in thin section, four are biotite granite, three are garnet-biotite granite, and two are alkali-feldspar granites. Biotite in these samples is usually partially chloritized and plagioclase is commonly sericitized. Microcline and striped perthite are also common constituents. Most of the gneiss clasts appear to be of granitic composition and of green-schist-facies or lower amphibolite-facies grade. Some specimens are apparently of granodioritic or dioritic composition. One of the few samples studied in thin section is plagioclase rich and contains a strongly developed granophyric texture. One clast of silicified ultramylonite was found in lithofacies A at the Camino Cielo locality. Some of the garnet-bearing clasts are massive or weakly foliated and thus impossible to differentiate from granitoid clasts.

The silicified metavolcanic clasts of assemblage 3 are mainly of felsic composition, but andesitic varieties are also

Table 5. Maximum observed clast size (centimeters) in Sespe conglomerates of the eastern Santa Maria basin area

[See figure 2 for site locations. ---, not found]

				Site location					
	Loma	Oso	Redrock	Camino Cielo		San Marcos Pass Highway		Lake Casitas	
Clast type	Alta	Canyon	Camp	Facies A	Facies F	Facies G	Facies A	Facies C	Facies C
Assemblage 1									
Chert	60	30	26	---	7	5	---	23	18
Lithic arenite	18	20	25	---	---	---	---	4	6
Metabasalt[1]	---	3	3	---	20	3	---	18	---
Ultramafic	10	---	---	---	10	---	---	5	5
Leucogabbro	---	17	---	---	38	8	---	---	---
Glaucophane schist	12	8	---	---	---	---	---	---	---
Jasper-calcite rock	90	13	4	---	---	---	---	30	30
Vesicular basalt	---	---	---	5	---	---	---	---	---
Quartz diorite	---	---	---	---	13.5	---	---	---	---
Vein quartz	3	3	3	---	---	---	---	---	---
Assemblage 2									
Arkosic arenite	---	15	75	10	---	---	6	38	30
Limestone	3	20	5	---	---	---	4	13	30
Assemblage 3									
Quartzite	---	40	10	25	22	---	35	60	35
Felsic metaporphyry	---	5	5	28	---	---	15	23	25
Felsic metatuff	---	---	4	15	---	---	6	15	32
Metaandesite	---	15	1	---	---	---	---	---	18
Granitoid[2]	---	20	40	18	10	---	5	18	34
Gneiss	---	16	6	8	---	---	---	14	45

[1] Includes metadiabase.
[2] Includes pegmatite.

present. Both clast types contain secondary minerals suggesting a greenschist-facies grade of metamorphism. The felsic types range from those with obvious pyroclastic texture (volcanic breccia, welded tuff, lapilli-ash tuff) to porphyritic and aphanitic types. The porphyritic types are most common and typically contain a quartz- and K-feldspar-rich microgranular groundmass. Two porphyritic varieties are easily differentiated—one type containing quartz, sanidine, and plagioclase phenocrysts and the other only feldspar phenocrysts. Plagioclase phenocrysts are typically altered to sericite, and mafics are sparse having been replaced by sphene, epidote, or piemontite. The felsic types appear to be mainly of rhyolitic composition and vary in color with groundmass composition. Those clasts containing abundant epidote are green; red clasts are hematitic; dark-greenish-

gray clasts are chloritic; and brown clasts are goethitic or have heavily albitized plagioclase as a major groundmass component. The metaandesite clasts are dark colored (low chroma), contain abundant mafic phenocrysts, and have lath-shaped plagioclase phenocrysts that are usually saussuritized or albitized. Felted (pilotaxitic) texture, which is rare in the felsic types, is common in the metaandesitic clasts.

Assemblage 1 predominates north of the Santa Ynez Fault (fig. 5; table 4). Thus, Sespe conglomerate at Loma Alta and Redrock Camp is greenish gray owing to the abundance of lithic arenite clasts, but that at Oso Canyon is redder owing to the abundance of chert clasts of this color. The largest clasts observed (table 5) are boulders of red chert and jasper-calcite rock at Loma Alta and large boulders of oyster-bearing arkosic arenite at Redrock

Table 6. Upsection changes in Sespe conglomerate clast suites

[See figure 2 for site locations. Tr, trace amount]

			Site location					
			San Marcos Pass Highway				Camino Cielo	
Clast type	Facies A (1)[1]	Facies A (2)	Facies A (3)	Facies A (4)	Facies A	Facies F	Facies F	Facies G
				Assemblage 1				
Chert	6	4	0	3	0	2	5	10
Lithic arenite	0	0	0	3	0	0	0	0
Metabasalt	0	0	0	2	0	6	4	13
Ultramafic	0	0	0	0	0	tr	0	0
Leucogabbro	0	0	0	0	0	76	85	77
Glaucophane schist	0	0	0	0	0	0	0	0
Jasper-calcite rock	0	0	0	0	0	0	0	0
Vesicular basalt	0	0	0	0	3	0	0	0
Quartz diorite	0	0	0	0	0	2	5	0
				Assemblage 2				
Arkosic arenite	22	1	10	3	3	0	0	0
Limestone	0	0	1	3	0	0	0	0
				Assemblage 3				
Quartzite	49	59	50	59	45	8	1	0
Felsic metaporphyry	5	7	22	2	28	0	tr	0
Felsic metatuff	4	16	7	11	3	0	0	0
Metaandesite	0	0	0	4	0	0	0	0
Granitoid	14	12	10	7	18	6	0	0
Gneiss	0	1	0	3	tr	0	0	0
Sample size	122	158	184	201	122	143	169	244

[1]Numbers in parentheses indicate ascending stratigraphic order.

Camp. Cobbles and small boulders of assemblage 3 are occasionally found at Redrock Camp and Oso Canyon, but no clasts of this assemblage are present at Loma Alta.

Localities south of the Santa Ynez Fault (Camino Cielo, San Marcos Pass, Lake Casitas) have a much more complex distribution of clasts, although those of assemblage 3 generally predominate (fig. 5; table 4). Lithofacies A at Camino Cielo and San Marcos Pass contains mainly assemblage 3 clasts. However, a few clast types from assemblages 1 and 2 appear and increase in abundance upsection in lithofacies A at San Marcos Pass (table 6). Boulders of oyster-bearing arkosic arenite 45 cm in diameter also are present in lithofacies A at San Jose Canyon (fig. 2). The conspicuous intraformational unconformity between lithofacies A and C along San Marcos Pass highway and between lithofacies A and F at Camino Cielo (fig. 5) corresponds with an upsection provenance change. At the San Marcos Pass highway locality, red beds above the unconformity contain abundant clasts of red chert, but at Camino Cielo boulder-sized leucogabbro and diorite clasts abruptly appear above the erosion surface (tables 4 through 6). Clasts of leucogabbro or diorite are not found anywhere at San Marcos Pass, but some clasts of these types up to 6 cm in size are present in basal beds of lithofacies C in San Jose and San Pedro Canyons to the west (fig. 2). In the Lake Casitas section and near Ojai (fig. 2), lithofacies A is absent and interbedded conglomerate lenses in the basal part of the Sespe Formation (lithofacies

D) are composed almost entirely of assemblage 1 (Howard, 1989). Upsection, conglomerates of lithofacies C are rich in either assemblage 1 or 3, and alternate from bed to bed (fig. 5; table 4). In the San Marcos Pass and Lake Casitas areas (table 5) and elsewhere in the central and eastern Santa Ynez Mountains, some clasts of assemblage 3 range up to boulder size, but the average clast is generally less than 10 cm in size. Figure 9 shows that there is a systematic westward decrease in the maximum sizes of granitoid and gneiss clasts across the Santa Ynez-Topatopa Mountains. The difference in the size ratio of softer granitoid/gneiss to harder quartzite lithologic types of clasts in figure 9 presumably reflects the effects of abrasion with increasing transport distance.

Metavolcanic Clast Varieties

Silicified metavolcanic rocks are common as clasts in conglomerates of Cretaceous and Paleogene age in and around the study area. Because they are exceptionally durable and resistant to chemical weathering, it is theoretically possible for such clasts to have been recycled during deposition of the Sespe Formation. The metavolcanic clast suites of the Jurassic and Cretaceous Espada Formation in the San Rafael Mountains and the Cretaceous and Paleocene San Francisquito Formation near the San Gabriel

Figure 9. Ratios of maximum granite/gneiss clast size to maximum quartzite clast size versus distance west of the San Gabriel Fault. Granitoid-quartzite ratio (solid line); Gneiss-quartzite ratio (dashed line). See figure 1 for site locations: Sespe Creek (dot), Lake Casitas (square), Montecito (triangle), Gibraltar Road (diamond), San Marcos Pass State Highway 154 (circle), Glen Annie Canyon (open square). Data from McCracken, 1972.

Mountains (fig. 1, localities MC and FC) are compared in table 7 with clast suites of Sespe conglomerate. Espada conglomerate is characterized by large quantities of metaandesite and dark gray metarhyolitic porphyry. Both varieties are uncommon in Sespe conglomerate at Camino Cielo (locality EC) and San Marcos Pass (locality SMP) in the central Santa Ynez Mountains. The Sespe suite includes abundant brown, goethitic metavolcanic clasts, which form only a minor component of the Espada clast suite. When the proportions (p_1 and p_2) of these varieties in Espada and Sespe sample pairs are compared, the differences are statistically significant at the 95 percent level (table 8). San Francisquito conglomerate contains more clasts of metaandesite and pyroclastic rocks and less of brown metaporphyry than Sespe conglomerate near Lake Casitas (locality LC) in the eastern Santa Ynez Mountains (table 7). The differences in these proportions are also statistically significant (table 8). These data suggest that metavolcanic clasts in Sespe Formation conglomerates at Camino Cielo and San Marcos Pass were not recycled from the Espada Formation. Similarly, the San Francisquito Formation does not appear to have supplied Sespe conglomerates at Lake Casitas with such clasts.

Conglomerate Clast Morphology

In table 9, clast shape data for rock units at the locations shown in figure 1 are summarized. For quartzite clasts, the values of R_w (roundness) are greater at the San Marcos Pass (SMP) and Gibraltar Road (GR) Sespe conglomerate localities than at the Lake Casitas locality (LC) farther east. The values of S/L and S/I are lower at the San Marcos Pass locality than at the Gibraltar Road and Lake Casitas localities. The results of t-tests when mean values at these locations are compared show which of the differences are statistically significant (table 10). These data suggest that Sespe quartzite clasts become rounder and flatter westward across the Santa Ynez Mountains. In the case of metavolcanic clasts, the same apparently holds for roundness; however, the other shape indices are highly variable, perhaps because of the effects of texture on shape (Drake, 1970; Howard, 1992).

Table 9 also shows that the values of the I/L index are generally greater and OPI values lower for the pre-Sespe submarine fan deposits studied (localities MC, OC, RS, FC, and WC) than for braided-stream deposits of the Sespe Formation (localities LC, GR, and SMP). In the eastern Santa Ynez Mountains, I/L and S/L values are significantly greater and ψ_p (see definitions in methods section) and OPI values are lower (table 11) for Jalama conglomerate at Rose Canyon (OC) than for Sespe conglomerate at Lake Casitas (LC) when quartzite clasts are compared. However, the comparison of ψ_p using quartzite clasts failed the F-test; thus, the t-test result should be
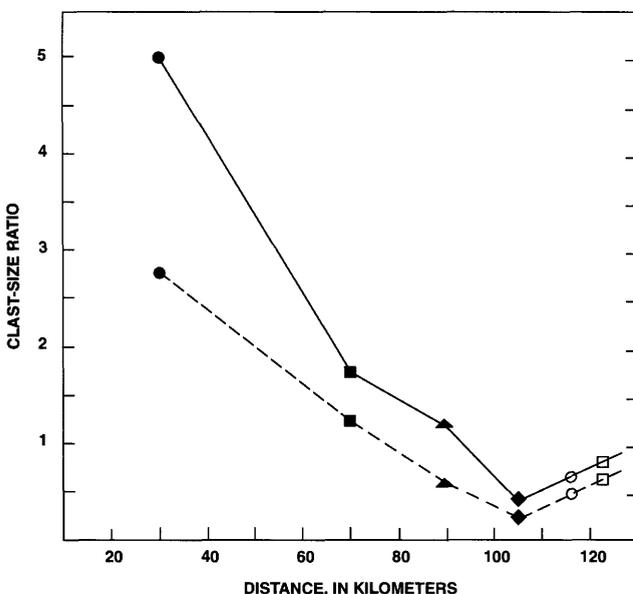
Table 7. Analysis of metavolcanic clast suites in Cretaceous and Paleogene conglomerates by percentage

[See figure 1 for site locations: MC, Manzana Creek; FC, Fish Creek; EC, east Camino Cielo; SMP, San Marcos Pass State Highway 154; LC, Lake Casitas. R, red; G, green; B, brown; D, gray; q, quartz phenocrysts; f, feldspar phenocrysts. Letters in parentheses indicate facies types (see table 1)]

Clast variety	Espada Formation	San Francisquito Formation	Sespe Formation		
	MC	FC	EC(A)	SMP(C)	LC(C)
Andesitic suite					
Porphyry and pyroclastic............37	37	0	3	6
Rhyolitic suite					
Pyroclastic:					
Welded tuff..............................5	5	11	29	5
Lapilli-ash tuff........................3	22	4	0	2
Volcanic breccia....................0	0	0	0	0
Porphyry:					
Rqf...0	0	2	2	0
Rf...0	0	3	0	3
Gqf..12	4	13	7	2
Gf...7	5	7	6	11
Bqf...7	4	26	19	3
Bf...7	23	26	26	40
Dqf..20	0	3	0	0
Felsite:					
R...0	0	0	0	0
G...0	0	0	0	0
B...2	0	0	6	27
D...0	0	0	0	0
Sample size..............................67	74	113	135	89

regarded as inconclusive. San Francisquito conglomerate near the San Gabriel Fault (FC) also differs significantly from Sespe conglomerate at Lake Casitas (LC) in terms of I/L and OPI for the quartzite clasts. In the central Santa Ynez Mountains, S/L, S/I and ψ_p values for metavolcanic clasts in Sespe conglomerate at Gibraltar Road (GR) are significantly lower (tables 9 and 11) than those in the Espada Formation (MC) and Cozy Dell Shale (WC). Quartzite and metavolcanic clasts in Sespe conglomerate at Gibraltar Road are also significantly rounder than those in the nearby Jalama Formation at Romero Saddle (RS). Farther west at San Marcos Pass, I/L values for quartzite and metavolcanic clasts in Sespe conglomerate (SM) are significantly lower than those in Espada (MC), Jalama (RS), and Cozy Dell (WC) conglomerates. All of these comparisons passed the F-test, with the exception of that using I/L values and quartzite clasts in Sespe and Cozy

Dell conglomerates, indicating that statistical comparisons using the t-test are valid. These relations suggest that quartzite and (or) metavolcanic clasts from the San Francisquito and Jalama Formations were not recycled to form conglomerates of the Sespe Formation at Lake Casitas. These same clast types in the Sespe Formation at San Marcos Pass and Gibraltar Road also do not appear to have been derived from nearby sources in the Jalama, Espada, or Cozy Dell rock units.

GEOLOGIC HISTORY OF THE STUDY AREA

Sespe Conglomerate Provenance

The rock types found as clasts in assemblage 1 (tables 4 through 6) generally crop out today as bedrock of

Table 8. Tests for statistically significant differences in metavolcanic clast suites of Sespe and pre-Sespe conglomerates

[See figure 1 for site locations: MC, Manzana Creek (Espada Formation); FC, Fish Creek (San Francisquito Formation); EC, east Camino Cielo (Sespe Formation); SMP, San Marcos Pass State Highway 154 (Sespe Formation); LC, Lake Casitas (Sespe Formation)]

Sample pairs	Confidence intervals $(p_1 - p_2)$	Hypothesis $p_1 = p_2$
Pyroclastic rocks		
MC versus EC	$-0.022 < p_1 - p_2 < 0.162$	Accept
MC versus SMP	$0.110 < p_1 - p_2 < 0.310$	Reject
FC versus LC	$0.085 < p_1 - p_2 < 0.314$	Reject
Andesite		
MC versus EC	$0.254 < p_1 - p_2 < 0.486$	Reject
MC versus SMP	$0.221 < p_1 - p_2 < 0.459$	Reject
FC versus LC	$0.189 < p_1 - p_2 < 0.431$	Reject
Red felsite and porphyry		
MC versus EC	$0.001 < p_1 - p_2 < 0.090$	Reject
MC versus SMP	$-0.0004 < p_1 - p_2 < 0.044$	Accept
FC versus LC	$-0.0005 < p_1 - p_2 < 0.065$	Accept
Green felsite and porphyry		
MC versus EC	$-0.109 < p_1 - p_2 < 0.129$	Accept
MC versus SMP	$-0.050 < p_1 - p_2 < 0.170$	Accept
FC versus LC	$-0.056 < p_1 - p_2 < 0.135$	Accept
Brown felsite and porphyry		
MC versus EC	$0.233 < p_1 - p_2 < 0.487$	Reject
MC versus SMP	$0.228 < p_1 - p_2 < 0.472$	Reject
FC versus LC	$0.291 < p_1 - p_2 < 0.569$	Reject
Gray felsite and porphyry		
MC versus EC	$0.069 < p_1 - p_2 < 0.271$	Reject
MC versus SMP	$0.104 < p_1 - p_2 < 0.296$	Reject
FC versus LC		Accept

Table 9. Morphologic indices of clasts in Sespe and pre-Sespe conglomerates

[Modified from Howard, 1992. See figure 1 for sample locations: MC, Manzana Creek (lower section); OC, Rose Canyon; RS, Romero Saddle; FC, Fish Creek; WC, West Camino Cielo; LC, Lake Casitas (midsection); GR, Gibraltar Road (midsection); SMP, San Marcos Pass State Highway 154 (midsection). M, metavolcanic; Q, quartzite. Morphologic indices: S/L, short/long axes; I/L, intermediate/long axes; S/I, short/intermediate axes; R_w, modified Wentworth roundness; ψ_p, maximum projection sphericity; OPI, oblate-prolate index; n, sample size; part, indicates that data applies to only part of the formation]

Rock unit	Age	Map symbol	Clast type	Morphologic indices						
				S/L	I/L	S/I	R_w	ψ_p	OPI	n
Espada Formation (part).	Cretaceous	MC	M	0.55	0.78	0.71	0.68	0.73	-0.54	66
Jalama Formation	Cretaceous	OC	Q	0.61	0.81	0.76	0.54	0.66	-0.42	20
			M	0.54	0.74	0.71	0.61	0.74	-0.66	14
Jalama Formation	Cretaceous	RS	Q	0.57	0.80	0.72	0.55	0.72	-0.47	27
			M	0.52	0.77	0.70	0.52	0.71	-0.01	16
San Francisquito Formation (part).	Paleocene	FC	Q	0.59	0.81	0.74	0.54	0.76	-0.24	22
			M	0.53	0.78	0.68	0.53	0.72	-0.68	20
Cozy Dell Shale	Eocene	WC	Q	0.59	0.80	0.74	0.63	0.77	-0.47	24
			M	0.57	0.78	0.74	0.61	0.75	0.19	19
Sespe Formation (part).	Oligocene	LC	Q	0.56	0.73	0.76	0.55	0.76	2.12	27
			M	0.54	0.73	0.73	0.54	0.73	1.48	16
Sespe Formation (part).	Oligocene	GR	Q	0.61	0.78	0.78	0.70	0.78	1.05	20
			M	0.48	0.79	0.62	0.65	0.68	-1.20	15
Sespe Formation (part).	Oligocene	SMP	Q	0.50	0.72	0.70	0.64	0.70	1.16	23
			M	0.54	0.52	0.76	0.64	0.73	0.42	23

Table 10. Calculated t-values for tests of statistically significant differences in clast shapes of Sespe conglomerates (see text for further explanation)

[See figure 1 for locations: GR, Gibraltar Road; SMP, San Marcos Pass State Highway 154; LC, Lake Casitas. M, metavolcanic; Q, quartzite. Morphologic indices: S/L, short/long axes; I/L, intermediate/long axes; S/I, short/intermediate axes; R_w, modified Wentworth roundness; ψ_p, maximum projection sphericity; OPI, oblate-prolate index]

Sample pairs	Clast type	Morphologic indices					
		S/L	I/L	S/I	R_w	ψ_p	OPI
GR versus SMP	Q	[3]2.99	1.41	1.91	1.45	[3]2.96	-0.07
	M	[1]-1.76	[3]6.15	[3]-3.34	0.2	[1]-1.85	-1.15
LC versus GR	Q	-1.43	-1.46	-0.56	[3]-4.28	-0.81	1.0
	M	1.41	-1.52	[2]2.25	[1]-1.99	1.60	[1]1.80
LC versus SMP	Q	[2]2.11	0.25	1.57	[2]-2.54	[3]2.83	0.69
	M	0	[3]5.55	-0.66	[1]-1.70	0	0.82

[1]Probably significant difference ($\alpha=0.10$).
[2]Significant difference ($\alpha=0.05$).
[3]Highly significant difference ($\alpha=0.01$).

Table 11. Calculated *t*-values for tests of statistically significant differences in clast shapes of Sespe and pre-Sespe conglomerates (see text for further explanation)

[See figure 1 for locations: GR, Gibraltar Road; SMP, San Marcos Pass State Highway 154; LC, Lake Casitas. M, metavolcanic; Q, quartzite. Morphologic indices: S/L, short/long axes; I/L, intermediate/long axes; S/I, short/intermediate axes; R_w, modified Wentworth roundness; ψ_p, maximum Projection sphericity; OPI, oblate-prolate index]

Sample pairs	Clast type	Morphologic indices					
		S/L	I/L	S/I	Rw	$\hat{A}$p	OPI
Eastern Santa Ynez Mountains							
LC versus OC ....Q		[1]1.69	[2]-2.18	0.0	0.28	[3,4]3.03	[2]-2.01
	M	0.0	-0.14	0.41	-1.11	-0.04	1.52
LC versus FC ....Q		-1.04	[2]2.59	0.58	0.35	0	[2]2.03
	M	0.23	-1.43	1.07	0.19	0	0.57
Central Santa Ynez Mountains							
GR versus MC .... M		[2]-2.12	0.30	[2]-2.45	-0.72	[2]-2.23	-0.59
GR versus RS ....Q		1.1	0.59	1.47	[3]3.43	[1]1.92	1.37
	M	-1.31	0.51	-1.63	[2]2.66	-1.11	-0.74
GR versus WC ....Q		0.51	-0.70	1.05	1.57	0.33	1.52
	M	[2]-2.42	0.24	[2]-2.42	1.01	[2]-2.30	-0.82
SMP versus MC .... M		-0.35	[3]-8.95	1.59	-1.04	0	1.02
SMP versus RS ....Q		[2]-2.33	[2]-2.0	-0.47	[2]2.10	-0.72	1.15
	M	0.62	[3]-6.61	1.33	[2]2.18	0.71	0.31
SMP versus WC ....Q		[3]-2.78	[2,4]-2.18	-0.98	0.23	[2]-2.64	1.18
	M	-0.84	[3]-6.73	0.45	0.63	-0.68	0.16

[1]Probably significant difference ($\alpha=0.10$).
[2]Significant difference ($\alpha=0.05$).
[3]Highly significant difference ($\alpha=0.01$).
[4]Did not pass *F*-test.

the Franciscan Complex (fig. 2), and the Franciscan component of Sespe conglomerate is easily recognized (Dibblee, 1966; McCracken, 1972; Anderson, 1980; Nilsen, 1984). The Franciscan Complex is largely mélange composed of a strongly deformed mass of sandstone and mudrock with large ophiolitic fragments scattered throughout (Page, 1981). Within the sedimentary sequence are lenticles of thin-bedded red and green chert and occasional conglomerate lenses containing well-rounded clasts of black chert and mafic metavolcanic rocks (Dibblee, 1966). The mafic masses are often strongly sheared and composed of metabasalt and ultramafic rocks. The metabasalts are more or less metamorphosed and usually aphanitic, although vesicular and amygdaloidal textures may be present (Bailey and others, 1964). The ultramafic rocks are usually serpentinized and sometimes altered to glauco-

phane schist or chlorite schist and jasper-calcite rock. Unlike the overlying Upper Cretaceous and Paleogene strata, Franciscan sandstones lack detrital K-feldspar (Dickinson and others, 1982), and although they are commonly called "graywacke," those in the Santa Maria basin vicinity typically contain less than 15 percent matrix (McLean, 1991).

Thus, Sespe conglomerates in the study area north of the Santa Ynez Fault (figs. 5 and 10) were probably derived mainly from nearby Franciscan sources. The Franciscan Complex lies unconformably beneath Miocene strata of the Santa Maria basin (Dibblee, 1950, 1966; Hall, 1978; McLean, 1991) and could have been eroded to form Sespe conglomerate at Loma Alta (fig. 10, locality 1). Franciscan bedrock north of the Little Pine Fault could have supplied Sespe conglomerate in Oso Canyon (fig. 10, locality 2)

with clasts; however, suitable source rocks do not crop out today northeast of Redrock Camp (fig. 10, locality 3). Clasts of assemblage 1 south of the Santa Ynez Fault (figs. 5 and 10) were also derived from Franciscan sources to the north because no sources are present to the east. South-directed paleocurrent vectors documented in table 3 and elsewhere (McCracken, 1972; Black, 1982; Howard, 1989) support this interpretation. The leucogabbro and diorite clasts found at Camino Cielo (fig. 10, locality 4) also prob-

ably are of Franciscan origin on the basis of their size, shape, abundance, and association with other clasts which are clearly of Franciscan origin. Leucogabbro and diorite bedrock do not crop out in the study area today, but quartz-bearing diorite and gabbro are found elsewhere within the Franciscan Complex (Bailey and others, 1964). Rocks of similar lithology are also found as part of the Coast Range ophiolite in the Santa Maria basin (Hopson and others, 1981) and as the Willows Plutonic Complex on Santa Cruz

EXPLANATION

Franciscan basement exposed

Sedimentary cover; no Franciscan exposed

Sespe conglomerates of Franciscan provenance

Sespe conglomerates of Franciscan and Mojave Desert provenance

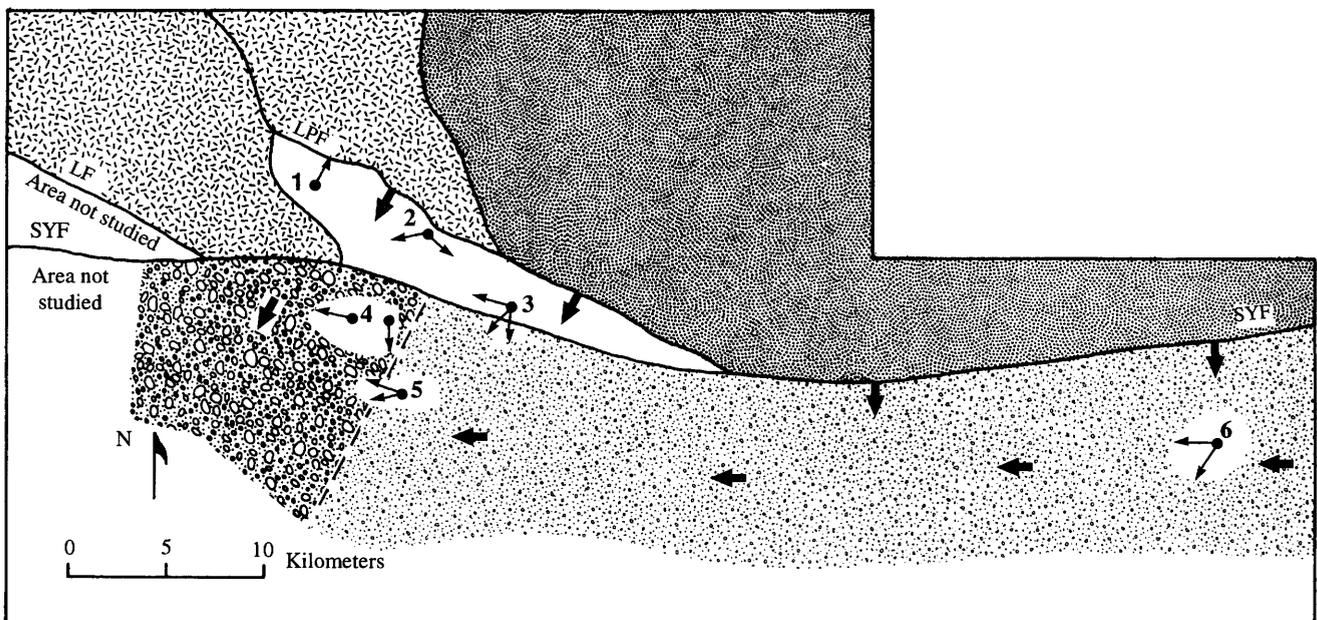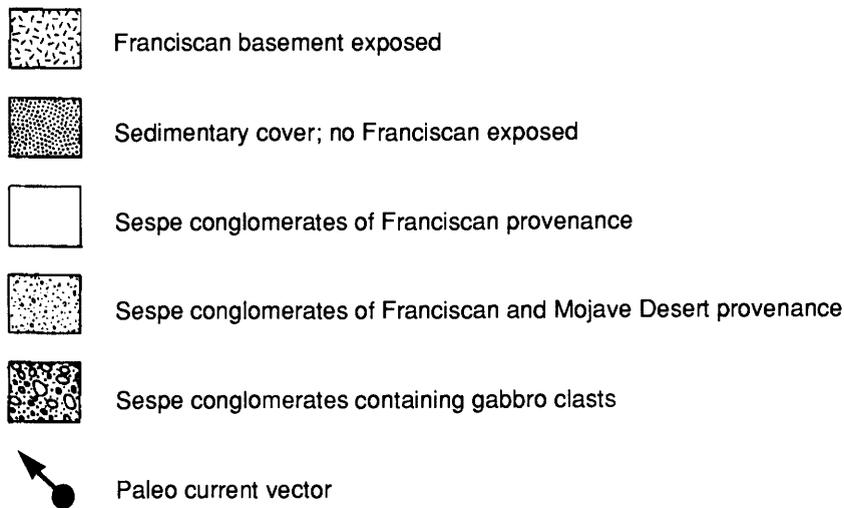Sespe conglomerates containing gabbro clasts

Paleo current vector

Figure 10. Inferred sediment dispersal directions (large arrows) in study area. Clasts of Franciscan provenance were derived from north, where no Franciscan sources are exposed today. Gabbro clasts were derived from nearby sources to northeast, where no source is exposed today. Clasts of Mojave Desert provenance were derived from the east. Observed paleocurrent vectors (small arrows) are based on data in Tables 2 and 3. Dashed line marks eastern limit of gabbro clasts. Fault names: SYF, Santa Ynez Fault; LF, Lompoc Fault; LPF, Little Pine Fault. Measured sections: 1, Loma Alta; 2, Oso Canyon; 3, Redrock Camp; 4, Camino Cielo; 5, San Marcos Pass Highway 154; 6, Lake Casitas.

Island (Sorensen, 1985) (fig. 1). They most likely represent late stage differentiates and could conceivably be found as intrusions anywhere within Franciscan mélange. Somewhat similar gabbro clasts are present in conglomerates of the Cozy Dell Shale in the Santa Ynez Mountains (fig. 1, locality WC); however, a thin section showed these to be olivine bearing and more mafic than the leucogabbro clasts in Sespe conglomerates.

The arkosic sandstone clast types in assemblage 2 (tables 4 through 6) have been attributed previously to unspecified sources in the Upper Cretaceous to Paleogene sedimentary sequence of the region (McCracken, 1972; Anderson, 1980). The underlying Coldwater Sandstone, which is characteristically micaceous arkosic arenite with abundant oyster reefs in the upper part (Dibblee, 1966; McCracken, 1972; O'Brien, 1973), is the most probable source. Weaver and Kleinpell (1963) noted that molluscan fossils in these sandstone clasts are similar to those in the underlying Gaviota Formation and Coldwater Sandstone. Sespe conglomerates unconformably overlap progressively older Coldwater strata northward from the south side to the north side of the Santa Ynez Mountains (Dibblee, 1966), and large boulders of oyster-bearing arkosic arenite and intact fossil fragments are found in the Redrock Camp section. Smaller clasts of this type are also common in the lower part of lithofacies C between San Marcos Pass and Lake Casitas (table 5).

McCracken (1972) suggested that the limestone clasts in assemblage 2 were derived from the Sierra Blanca Limestone, which crops out just north of the Santa Ynez Fault at the base of the Paleogene sequence (Dibblee, 1966). Several lithofacies have been identified in the Sierra Blanca Limestone including sedimentary breccia, algal-laminated lime mudstone, coral- or foraminifer-rich lime mudstone, and calcarenite containing red and green chert pebbles of Franciscan provenance (Walker, 1950; Page and others, 1951; Dibblee, 1966; Schroeter, 1972). Although the calcarenite lithology of the Sierra Blanca Limestone does strongly resemble some clasts found in Sespe conglomerate, a source in the Sierra Blanca Limestone is unlikely because the other lithofacies are not represented. The typical Sespe limestone clasts are massive and unfossiliferous. A much more probable source of the limestone clasts is the upper part of the Coldwater Sandstone, which contains massive calcareous nodules and highly calcareous oyster reefs as well as calcarenite with lithics of Franciscan provenance. Massive calcareous concretions are found in the Franciscan Complex in the study area (Dibblee, 1966; Bailey and others, 1964; Schussler, 1981) and apparently furnished Sespe conglomerate at Loma Alta with limestone clasts.

As noted above, there is disagreement in the literature regarding the provenance of the suite of rocks composing assemblage 3. In the alluvial fan depositional model for the Sespe Formation, these clasts are regarded as reworked from local rock units, whereas in the fluvial-deltaic model, they are first cycle and derived from more distant sources. Recycling of material from conglomeratic deposits should result in an abundance of sedimentary clasts, and although this is observed locally in the Pine Mountain area, the rarity of such clasts in the Los Angeles basin area implies a first-cycle origin (Howard, 1989; Howard and Lowry, 1994). Thus, with the possible exception of certain varieties of granitoid and gneiss also present in the coastal batholithic belt of southern California, first-cycle assemblage 3-type clasts must have come from sources at least as far east as the Mojave Desert region because no such bedrock presently crops out west of the San Andreas Fault (Woodford and others, 1968; Woodford and Gander, 1980). The closest possible source for the metavolcanic clasts is the Sidewinder volcanic series of Bowen (1954) in the western Mojave Desert near Victorville, California; however, specific sources for certain red metavolcanic-clast types also have been identified in northwestern Sonora, Mexico (Abbott and Smith, 1978, 1989). The presence of two-mica, garnet-bearing, and alkali-feldspar granite clasts in Sespe conglomerates suggests a provenance in the eastern Mojave Desert region (Howard, 1987), and certain types of quartzite clasts have been matched petrographically with bedrock sources in central Arizona (Howard and Lowry, 1994). These data suggest an Eocene-Oligocene paleogeography in which the Mojave-Sonora Desert region was an alluvial plain with long braided rivers draining westward from an uplifted hinterland in Arizona and northern Mexico.

Given the south- and west-directed Sespe paleocurrent vectors in the Santa Ynez Mountains (table 3; fig. 5) and assuming that recycling has taken place, possible conglomeratic source rocks include the Espada, Jalama, San Francisquito, and Cozy Dell rock units (fig. 1). However, tables 7 and 8 show that metavolcanic clast varieties in these rock units are significantly different from those in Sespe conglomerate. The morphologic indices of quartzite and metavolcanic clasts in the pre-Sespe conglomerates of submarine fan origin studied are also generally different from those in fluvial Sespe conglomerates (tables 9 and 11). The contrasts in clast shape cannot be attributed to abrasion during reworking because no differences in roundness are observed. It also seems unlikely that such clasts were reworked upsection in the Santa Ynez Mountains from sources in the underlying Jalama and Cozy Dell rock units because the Sespe Formation is part of a conformable stratigraphic succession. Given these data, sediment recycling was not the predominant mechanism that provided Sespe conglomerate with such clasts. A first-cycle, eastern provenance for assemblage 3 in the Santa Ynez-Topatopa Mountains is also suggested by the eastward coarsening of granitoid and gneiss clasts (fig. 9) and by the magmatic arc provenance of associated Sespe sandstones (McCracken, 1972; Howard, 1987). A west-directed sediment dispersal pattern is further suggested by the

westward increase in the proportions of flat quartzite and metavolcanic clasts (tables 9 and 10). Modern and ancient braided-stream deposits have been found to show a down-current increase in flat clasts owing to selective sorting by shape (Bluck, 1964, 1965; Bradley and others, 1972).

Sespe conglomerate in the Santa Ynez Mountains, therefore, appears to have a complex but largely first-cycle provenance (fig. 10). A Franciscan component was contributed from sources to the north, some clasts of assemblage 2 were eroded from the underlying upper part of the Coldwater Sandstone, and both suites were mixed with a large volume of assemblage 3 clasts derived from sources east of the San Andreas Fault, possibly including distant parts of the Mojave Desert region. Reworking of assemblage 3-type clasts probably did occur locally within the Sespe Formation. This is seen most clearly in the Camino Cielo section (fig. 5), where clasts of assemblage 3 were obviously cannibalized from lithofacies A below (table 6). The same mechanism may account for the small component of assemblage 3 clasts found locally at other sites north of the Santa Ynez Fault (fig. 5), although recycling of clasts from older conglomerates cannot be ruled out entirely.

Depositional History

Facies relations in the Santa Ynez Mountains illustrated in figure 8 and provenance interpretations discussed above suggest that the Sespe Formation was deposited as part of two depositional sequences separated by an intraformational erosional unconformity. The lower sequence (lithofacies A, B, and E) is a late Eocene phase of braided-river/braid-delta deposition in an alluvial coastal-plain setting. The sediments deposited were probably of first-cycle origin and derived mainly from sources in the Mojave Desert region east of the San Andreas Fault. However, the northward thinning of the Gaviota Formation (Dibblee, 1950, 1966; O'Brien, 1973) and the upsection appearance of clasts of Franciscan Complex and Coldwater Sandstone provenance in lithofacies A (table 6) suggest that a paleotopographic high ("San Rafael uplift") in the Santa Maria basin became exposed to subaerial erosion during late Eocene time. This paleogeography was even more strongly defined when the upper sequence (lithofacies C, D, F, and G) was deposited during late Oligocene time. In the area of the present Santa Ynez Mountains, first-cycle sediment of Mojave Desert provenance was deposited on an alluvial plain, probably by a long west-flowing braided river. However, in the area of the present Santa Maria basin, another alluvial plain was drained by much shorter south- or southwest-flowing braided tributary streams that carried considerable detritus of Franciscan origin. This sediment dispersal pattern is inferred from the facts that in the eastern and central Santa Ynez Mountains paleocurrent vectors are both

south and west directed and that clast suites were derived from both Franciscan and Mojave sources (in the Lake Casitas section these two clast suites alternate from bed to bed (table 4), but elsewhere they are intermixed). The upper depositional sequence was deposited after a significant hiatus and grades upsection from braided to meandering fluvial sedimentation, suggesting some paleohydrological response to changing tectonic or climatic conditions.

In the eastern Santa Maria basin, only the upper depositional sequence is present, as shown by the unconformable northward overlap and Franciscan provenance of Sespe conglomerates. The upsection change from braided-to meandering-stream deposition in the Oso Canyon and Redrock Camp sections suggests a correlation with the upper depositional sequence found in the Santa Ynez Mountains to the south. This correlation implies that the upsection shift from west- to south-directed paleocurrent vectors noted above was due to a gradual change in paleoslope direction during the late Oligocene. The significance of the north-directed vector mean at Loma Alta is unclear. If this vector represents deposition contemporaneous with those sections described above, it implies that a complex paleotopography was present. On the other hand, if Sespe conglomerate at Loma Alta is equivalent to the Lospe Formation of early Miocene age, this vector may represent a drainage reversal associated with a third, possibly unrelated, regional depositional sequence.

In figure 11, I have integrated sedimentary facies relations in the study area and in the Santa Ynez Mountains farther to the west (fig. 8) with those in the Topatopa Mountains and the Los Angeles basin area to the east. Although the exact sequence stratigraphy is unknown because of poor internal age control, the reconstruction suggests that the lower part of the Sespe Formation and laterally equivalent marine strata (fig. 11, time lines T1 and T2) represent progradational (regressive) sedimentation from the late middle to late Eocene. This interpretation is supported by the fact that conglomerates in the lower part of the Sespe Formation of the Simi Valley area, which lie stratigraphically below rocks containing late Uintan fossil vertebrates, were probably deposited in a beach setting (Howard, 1992) and therefore define the location of a paleoshoreline of late middle Eocene age. As discussed above, the Coldwater-Sespe contact decreases in age westward across the Santa Ynez-Topatopa Mountains, and another paleoshoreline of late Eocene age is defined by the westward gradation of fluvial Sespe strata into deltaic Gaviota and Alegria strata west of San Marcos Pass. The westernmost extent of this offlap sequence is uncertain because it is apparently truncated by the inferred erosional unconformity within the Alegria Formation.

The basis for showing the intraformational erosional unconformity within the Sespe Formation is the conspicuous erosion surface at the San Marcos Pass highway locality (fig. 8), where late Oligocene oreodont-bearing red

beds lie directly on a thin sequence of lighter colored late Eocene strata of lithofacies A. An unknown but substantial amount of relief must have been developed on lithofacies A in this particular area because nearby, along Old San Marcos Pass Road, lithofacies D rests directly on a much thicker section of lithofacies A (fig. 7). The same unconformity appears to be present to the north at the Camino Cielo locality, where lithofacies F overlies lithofacies A; lithofacies F unconformably overlaps progressively older rock units northward from this location (fig. 5). The unconformity is also interpreted as extending westward to Gaviota Canyon, but the amount of relief in the intervening area has not been assessed; therefore, the erosion surface is shown schematically in figures 8 and 11. The unconformity probably extends eastward to Rincon Creek where lithofacies A was eroded away completely, and it apparently corresponds with the Coldwater-Sespe contact (in other words, the "transitional beds"-Sespe contact) in the Lake Casitas (figs. 5 and 6) and Ventura River sections. The latter situation is inferred from the fact that the lowest beds at the base of the thick sequence of Sespe red beds at Lake Casitas and near Ojai are composed almost entirely of assemblage 1 clasts. The same unconformity is shown extending eastward beyond the study area and into the Topatopa Mountains on the basis of (1) sedimentologic evidence for two depositional sequences in the Sespe Formation of the Los Angeles basin area (Howard, 1989; Howard and Lowry, 1994), (2) magnetostratigraphic data from a section of the Sespe Formation in the Simi Valley area that suggest that much or all of the lower Oligocene section is missing (Prothero and others, 1992), and (3) studies in areas adjacent to the Transverse Ranges that

show the presence of a regional Oligocene unconformity (Davis and Lagoe, 1988).

Stratigraphically above the unconformity, the reconstruction shows the upper part of the Sespe Formation and laterally equivalent marine strata (fig. 11, time lines T3 and T4) as composing a retrogradational (transgressive) episode of sedimentation that occurred during the late Oligocene and early Miocene. The lower part of the upper depositional sequence of the Sespe Formation is interpreted as grading laterally westward into the upper part of the Alegria Formation, and the intraformational unconformity is shown extending westward into the Alegria Formation, as inferred from relations in the Gaviota Canyon area (fig. 8). In the westernmost Santa Ynez Mountains, the unconformity is inferred to lie at the base of the Vaqueros Formation, which rests with slight (<15°) angular discordance on Alegria strata of late Eocene age west of Cojo Canyon (fig. 1). The Vaqueros Formation in this area is an upward-deepening sequence that includes coarse proximal fan-delta deposits of Franciscan provenance in its lower part and distal fan-delta and strandline deposits in its upper part (Rigsby, 1989). Associated paleocurrent vectors are east-directed indicating a paleotopographic high at the west end of the Santa Ynez Mountains. These coarse fan-delta deposits are inferred to be time-stratigraphically equivalent with the deltaic Alegria Formation and in turn with the Franciscan-rich braided fluvial Sespe Formation conglomerates, which lie above the intraformational unconformity farther east (fig. 11). This interpretation resolves the apparent stratigraphic discrepancy between the 30 to 26 Ma age of the Sespe Formation (based on the oreodont *Sespia* sp.) and the older ages of 31 to 29 Ma reported for the Vaqueros Formation
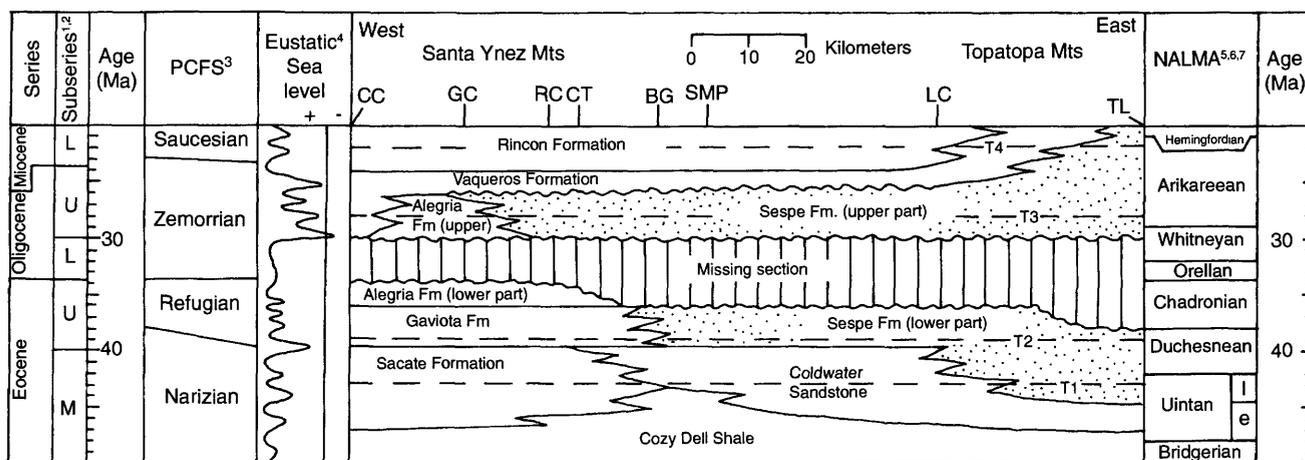

Figure 11. Facies relations between the Sespe Formation and laterally equivalent units in Santa Ynez-Topatopa Mountains, Calif. Modified from Hornaday and Phillips (1972), as discussed in text. See figure 1 for site locations: CC, Canada del Cojo; GC, Gaviota Canyon; RC, Canada del Refugio; CT, Canada del Capitan; BC, Bartlett Canyon; SMP, San Marcos Pass State Highway 154; LC, Lake Casitas; TL, Sespe Creek. Time lines T1 through T4 illustrate offlap-onlap relations. PCFS, Pacific coast foraminiferal stages; NALMA, North American land mammal ages. L, lower; M, middle; U, upper; e, early; l, late. References for chronologic framework: 1, Berggren and others (1992); 2, Cande and Kent (1992); 3, Bartow (1991); 4, Haq and others (1988); 5, Woodburne (1987); 6, Swisher and Prothero (1990); 7, Prothero and Swisher (1992).

(Bishop and Davis, 1984; Rigsby, 1989) at the west end of the Santa Ynez Mountains. The average age of the Vaqueros Formation in the western and central Santa Ynez Mountains farther to the east appears to be about 26 Ma (Rigsby, 1989), an age which is consistent with the time-transgressive onlap of the shallow marine facies of the Vaqueros Formation. This reconstruction is further supported by the facts that (1) the Vaqueros Formation is locally conformable on the Alegria Formation between Arroyo Bulito and Canada del Agua Caliente (fig. 1) in the western Santa Ynez Mountains (Weaver and Kleinpell, 1963), (2) in the northern part of the western Santa Ynez Mountains (Alisal Creek area, fig. 1), the Sespe Formation is unconformable on Gaviota and older strata but grades conformably upward into the Vaqueros Formation (Dibblee, 1950), and (3) the uppermost part of the Sespe Formation near Gaviota Canyon is overlapped by the Vaqueros Formation instead of grading laterally into the Alegria Formation (Dibblee, 1966).

The irregularity of the erosion surface below the upper sequence and the presence of exposed Franciscan Complex both to the north and west of the Santa Ynez Mountains suggest that a complex paleotopography was present during deposition of the upper depositional sequence of the Sespe Formation. These deposits evidently reflect basin backfilling following a major episode of base-level lowering; however, in the Santa Maria basin and other areas nearest the Franciscan sources, these deposits are, nevertheless, relatively thinly and evenly bedded, suggesting deposition on a rather broad alluvial plain. The upper depositional sequence probably represents aggradation below and around paleotopographic highs of exposed Franciscan Complex bedrock. Although there are alluvial-fan-like deposits locally, Sespe sedimentation probably did not involve discrete fan-shaped geomorphic features.

The upsection gradation from braided- to meandering-stream deposits is virtually a basinwide feature (Howard, 1987, 1989); thus the upper depositional sequence of the Sespe Formation appears to cover a significant geographic area. The widespread geographic and stratigraphic distribution of *Sespia* sp. in the upper depositional sequence and the age of the overlying Vaqueros Formation suggest further that sedimentation occurred rather rapidly (from approximately 30 to 26 Ma) in the Santa Ynez Mountains. Farther east in the Topatopa Mountains and Los Angeles basin area, sedimentation persisted into early Miocene time as shown by a paleoshoreline defined by interfingering between the Sespe and Vaqueros Formations. A characteristic feature in the study area is the presence of abundant Franciscan detritus, but in the Topatopa Mountains and Los Angeles basin area there is an abundance of granitic detritus in the upper depositional sequence of the Sespe Formation (Howard, 1989).

The two depositional sequences or cycles in the Sespe Formation appear to be preserved most completely near Simi (fig. 1) in the vicinity of the Los Angeles basin. Each cycle there is a couplet of braided and meandering fluvial deposits (Howard, 1989; Howard and Lowry, 1994). The presence of grazing, leaf-eating, and arboreal types of land vertebrate fossils (Savage and others, 1954) and the remains of palm tree fronds and fern (?) spores (Protzman, 1960) shows that vegetation was abundant when the meandering stream facies were being deposited. Flemal (1966), McCracken (1972), and Peterson and Abbott (1979), however, cited sedimentologic evidence in favor of semiarid or drier conditions during deposition of the braided stream facies. Thus, climatic conditions may have fluctuated markedly as the Sespe Formation was deposited. Frederiksen (1991) recognized four pulses of climatic deterioration during the middle Eocene to earliest Oligocene. This deterioration of global climate may have peaked during the middle Oligocene, as shown by profound sea level drop (fig. 11).

The actual time span represented by the lower depositional sequence in the Sespe Formation is poorly constrained; thus, correlation with the global sea level curve (fig. 11) is highly speculative. May and Warme (1987) presented evidence that eustasy played a significant role during Paleogene sedimentation in southern California, despite a tectonically active environment. However, Miall (1992) recently questioned the validity of such correlations with the eustatic curve. Figure 11 also suggests that the late middle to late Eocene progradational episode of Sespe sedimentation spanned several such postulated episodes of global sea level change. The outbuilding of the Sespe coastal plain, therefore, probably reflects high terrigenous sediment influx from the Mojave Desert and is unrelated to eustasy. Uplift in the hinterland of the Sespe fluvial system caused by the Laramide orogeny may have been a contributing factor as shown by reset K-Ar ages in the Sonora Desert region (Abbott and Smith, 1989). Frakes and Kemp (1972) suggested that seasonal monsoon conditions may have existed in the southwestern United States during the Eocene. This also may have contributed to the development of large west-flowing river systems despite the fact that conditions may have been arid in coastal California.

The age and geographic extent of the intraformational unconformity and the upper depositional sequence in the Sespe Formation are also poorly defined. The presence of the oreodont *Sespia nitida* Leidy suggests that sedimentation resumed about 30 Ma, and if the unconformity is truly a basinwide feature, which remains to be proven, base-level fall may have resulted from the very drastic eustatic sea level drop that occurred at about this same time (Haq and others, 1988). Eustatic climatic control of the upper depositional sequence is suggested by the fact that deposition coincides with a period of rising or globally high sea level (fig. 11) and by the upsection change from braided to meandering deposition. A similar alteration in river morphology is recorded in the upper Pleisto-

cene to lower Holocene sedimentary sequences associated with many of the world's rivers; this alteration in morphology is thought to have developed in response to climatic change following the last major episode of glaciation (Schumm, 1968; Jackson, 1978; Baker, 1978). The coarse Sespe braided stream alluvium was possibly deposited initially under conditions of fading glacial aridity during the early stages of rising sea level, and then meandering stream sediments were deposited in response to more humid conditions, rising sea level, and decreasing river gradient.

Tectonic History

An angular unconformity is present locally beneath the Sespe Formation north of the Santa Ynez Fault. Below the unconformity in the eastern Santa Maria basin, the youngest rock unit deformed is apparently the Coldwater Sandstone (Dibblee, 1966); however, 28 km to the west in the northern part of the western Santa Ynez Mountains (Alisal Creek area, fig. 1), the Refugian Gaviota Formation is the youngest unit folded, and the Sespe Formation lies below Zemorrian strata of the Vaqueros Formation (Dibblee, 1950; Weaver and Kleinpell, 1963). Thus, the episode of base-level drop, which caused both erosion of the San Rafael uplift and the upsection appearance of abundant Franciscan detritus in the Sespe Formation, has been attributed to tectonism during the late Eocene and Oligocene Ynezan orogeny (Dibblee, 1950). The apparent absence of lower Oligocene rocks in the Santa Ynez Mountains may be exclusively the result of nondeposition owing to regional uplift during the Ynezan orogeny. However, the absence of an angular unconformity in the Santa Ynez-Topatopa Mountains south of the Santa Ynez Fault and in the Los Angeles basin area suggests that the Ynezan deformation was a local event. The apparent absence of the lower Oligocene section over such a wide geographic area probably indicates that eustasy contributed significantly to the episode of regional base-level lowering that occurred during deposition of the Sespe Formation.

Given the sediment dispersal pattern observed in the Sespe Formation, it seems unlikely that the Ynezan orogeny was caused by extensional deformation and that the Santa Ynez Fault was an active normal fault during Sespe deposition. By analogy with alluvial fans formed along active normal faults in modern settings (Bull, 1977; Nilsen, 1982), syntectonic Sespe conglomerates would be markedly coarser and thicker on the downthrown side of such a fault, and paleotransport indicators should define a paleoslope dipping in a direction perpendicular to the fault trace. However, the fan-like deposits at Camino Cielo neither thicken nor coarsen northward, and the absence of leucogabbro clasts at San Marcos Pass (fig. 10) implies a paleoslope dipping toward the southwest—that is, in a

direction oblique to the trend of the Santa Ynez Fault. Furthermore, the folding or anticlinal warping of the Santa Maria basin suggests that the Ynezan orogeny reflects contractional rather than extensional strain. It is possible that reverse dip-slip separation was actively occurring on the Santa Ynez Fault during the Ynezan orogeny as postulated by Namson (1987). In his model, the Santa Ynez Fault developed as a backthrust, while the San Rafael uplift was rising vertically along a south-verging thrust fault. However, his reconstruction seems to require that the Santa Ynez Mountains tectonic block (on the hanging wall of the Santa Ynez Fault) be structurally and topographically higher than the San Rafael uplift. This is at odds with the late Oligocene paleogeographic reconstruction presented in this paper and with the unconformable northward overlap of Sespe Formation conglomerates. In terms of timing, it appears that the Santa Ynez Fault could have developed by Namson's (1987) proposed mechanism long after the end of Sespe deposition.

If any of the major faults in the study area were vertically active during Sespe sedimentation, relief must have been much lower than at present. Franciscan clasts 2 to 4 m in size are common in modern streams of the study area, but even at Loma Alta where the coarsest Sespe conglomerate was observed, Franciscan clasts are less than 1 m in size. As noted above, the Sespe Formation is conformable beneath the late early Miocene "Temblor" sandstone (Dibblee, 1966) at this locality and therefore may be correlative with the Lospe Formation of early Miocene age. If so, the Sespe Formation conglomerates in this particular case may reflect greater topographic relief created by the Lompocan orogeny, an early Miocene event which followed the Ynezan episode of deformation (Dibblee, 1950). This would account for the apparent drainage reversal and the coarse nature of the detritus at the Loma Alta locality.

McCracken (1972) noted that Franciscan detritus is found in Sespe Formation conglomerates south of the Santa Ynez Fault as far east as Ojai, yet the present exposures of Franciscan bedrock on the north side of the fault are much farther west (fig. 2). This implies that approximately 35 km of left-lateral separation has occurred along the Santa Ynez Fault since the Oligocene. Horizontal offset along the Santa Ynez Fault is also suggested by the fact that no bedrock source is present northeast of the proximal, leucogabbro-bearing conglomerate at Camino Cielo (fig. 10) and by the juxtaposition of this conglomerate facies with Sespe conglomerate at Redrock Camp, which has a significantly different assemblage of clasts. Lateral displacement on the Little Pine Fault also may be indicated by the absence of a suitable Franciscan bedrock source northeast of the Redrock Camp section. Crustal shortening has no doubt contributed to the apparent juxtaposition of contrasting lithofacies, and it is possible that Franciscan basement sources once exposed are now buried beneath hanging wall rocks of

the Santa Ynez and Little Pine Faults. Thus, the sense, magnitude, and timing of such lateral separations (if they exist) cannot be determined without additional subsurface information. Such data still may not provide a conclusive answer because it is also possible that fault slivers of Franciscan basement once present may have been eroded away entirely. Dibblee (1987) pointed out that severe constraints are placed on the amount of possible displacement by the negligible offset observed at the eastern termination of the Santa Ynez Fault.

Previous authors have concluded that the paleogeography of the western Transverse Ranges is explained better by using the tectonic rotation model of Luyendyk and others (1980, 1985), than by using sedimentologic data alone. For example, it has been suggested that paleocurrent vectors derived from Cretaceous and Paleogene deep-sea fan deposits in the Santa Ynez Mountains (Krause, 1986; Thompson, 1988) and on the Channel Islands (Kamerling and Luyendyk, 1985) appear to have been rotated. However, it should be noted that sediment transport is typically transverse in submarine fans; hence, lateral facies relations are perhaps more reliable indicators of regional paleoslope. In the Eocene section of the Santa Ynez Mountains where facies relations are reasonably well defined (for example, Van de Kamp and others, 1974; Thompson, 1988), they indicate an east-to-west transition from shallow to deep marine facies, a facies transition which one would expect in a basin that has not been rotated. The same pattern is evident in figures 8 and 11. A similar argument that has been made in favor of the rotation model is that deposition of the Sespe Formation as alluvial fans in a rift basin is explained better if the basin were originally oriented north to south (Luyendyk and others, 1985). The results of this study, however, suggest that the Sespe Formation was deposited in a forearc basin setting and not in a rift basin. These observations do not disprove the Luyendyk and others (1980,1985) rotation model; they only argue that tectonic rotation is not necessary to explain the sedimentologic data. On the basis of this rotation model, it has been postulated that the forearc basin sediments of the Santa Ynez Mountains were deposited in a trough which was originally oriented north to south (Hornafius and others, 1986). This basin geometry is reasonable and applies equally well for the Sespe Formation, but because sedimentologic data lack an absolute frame of reference, they cannot be used to either prove or disprove the rotation model.

In early plate tectonic models, the East Pacific Rise was thought to have collided with a subduction zone in southern California sometime between 38 and 29 Ma (Atwater, 1970; Atwater and Molnar, 1973). Consequently, the upper Eocene part of the Sespe Formation has been interpreted as the final phase of forearc basin sedimentation, and rifting was presumed to be the dominant factor that controlled deposition during the Oligocene (Nilsen, 1984, 1987). It is now recognized that regional extension

began later (about 25 Ma) and then peaked during the early Miocene (from 23 to 18 Ma) (Spencer and Reynolds, 1989; Dokka and Glazner, 1982; Bartley and Glazner, 1991). This orogenic activity, which included detachment faulting in the Mojave Desert region, is now thought to have preceded strike-slip movements associated with the San Andreas system, and subduction-related volcanic and forearc basin sedimentation are believed to have continued into the early Miocene in coastal southern California (Tennyson, 1989; Cole and Basu, 1992). If these interpretations are correct, then the Ynezan orogeny was unrelated to the major change of tectonic regimes associated with ridge-trench collision as previously thought. These data imply that large-scale strike-slip movements are unlikely to have occurred during Sespe deposition; indeed, all of the major faults in the study area may in fact greatly postdate Sespe deposition.

CONCLUSIONS

The results of this study support the conclusion drawn long ago by Reed (1929) that the Sespe Formation conglomerates of the Santa Ynez-Topatopa Mountains were deposited mainly as part of the delta of a very large river and not as alluvial fans. Evidence has been presented here that suggests the Sespe Formation has a complex but mainly first-cycle provenance, including both a Mojave Desert suite of conglomerate clasts derived from distant sources and a Franciscan component derived from nearby sources. The former suite of clasts was probably derived in part from very distant sources to the east. This suggests an Eocene-Oligocene paleogeography in which the Mojave Desert region was a broad alluvial plain with long braided rivers draining westward from a hinterland farther east. The coastal part of this braid plain was bordered by a piedmont area developed around a Franciscan source terrane. Thus, the Franciscan clast assemblage is associated with alluvial-fan-like deposits in the eastern Santa Maria basin, but lateral facies relations define a relatively broad braid plain aggrading below and around paleotopographic highs of exposed Franciscan Complex bedrock; Sespe sedimentation probably did not involve discrete fan-shaped geomorphic features in this area.

A major outcome of this study is the recognition of an intraformational erosional unconformity of possible regional extent within the Sespe and Alegria Formations. This implies that these rock units were deposited as two discrete depositional sequences rather than as one long continuous event. The relations between sequence stratigraphy, tectonism, and eustasy remain to be fully determined; however, regional considerations suggest that climatic and eustatic effects were much more important than previously thought. The apparent absence of the lower Oligocene section probably reflects a combination of tectonism (Ynezan orogeny) and eustasy locally in the

vicinity of the Santa Maria basin; elsewhere, it may be exclusively the result of a middle Oligocene global sea level drop. Additional studies are needed to better define this unconformity and to evaluate its geologic significance. The most important area to investigate further is probably the western Santa Ynez Mountains, where it may be possible to combine a careful study of the marine-nonmarine lateral facies relations with detailed paleontological and magnetostratigraphic studies of the Alegria Formation.

Another significant result of this study is the conclusion that the overall pattern of sediment dispersion suggests an absence of active faulting in the region studied at least until the end of Sespe-Vaqueros deposition during the early Miocene. Deformation associated with the Ynezan orogeny can be explained simply by anticlinal warping or broad folding of the San Rafael uplift. Thus, the alluvial-fan/rift-basin scenario often found in the literature is unsupported by the results of this study. Sedimentation is interpreted to have occurred in a forearc-basin setting, an interpretation which is consistent with accumulating regional evidence indicating that subduction persisted along the continental margin in southern California until the early Miocene. Horizontal offsets along the Santa Ynez and Little Pine Faults after Sespe deposition are suggested by geologic relations in the study area, but the juxtaposition of contrasting lithofacies is probably partly the effect of crustal shortening. Sediment-source mismatches also are present but are difficult to evaluate without additional information on the subsurface distribution of Franciscan bedrock. Lateral offsets have been predicted by and incorporated into models involving large-scale tectonic rotations; however, the sense, timing, and magnitude of such movements are still unconstrained by independent geologic evidence.

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Appendix. List of conglomerate clasts from the Sespe Formation studied in thin section

[See figures 1 and 2 for sample locations. S. Y. Mts., Santa Ynez Mountains; metavolc., metavolcanic]

Sample number	Lithofacies	Sample location	Lithology
17-1	C	Sespe Creek, Topatopa Mts.	biotite-granite
17-2	C	Sespe Creek, Topatopa Mts	alkali feldspar-granite
16-1	C	Lake Casitas, Santa Ynez Mts.	pebbly sandstone
16-2	C	Lake Casitas, Santa Ynez Mts.	biotite-granite
16-3	C	Lake Casitas, Santa Ynez Mts.	garnet-biotite-granite
16-4	C	Lake Casitas, Santa Ynez Mts.	garnet-biotite-granite
16-5	C	Lake Casitas, Santa Ynez Mts.	biotite-granite
15-2	A	Toro Canyon, Santa Ynez Mts.	biotite-granite
15-3	A	Toro Canyon, Santa Ynez Mts.	biotite-granite
13-2	A	Old San Marcos Pass Rd., S. Y. Mts.	biotite-granitic gneiss
13-3	A	Old San Marcos Pass Rd., S. Y. Mts.	alkali feldspar-granite
1-2	A	Camino Cielo, Santa Ynez Mts.	red felsic metavolc.
1-3	A	Camino Cielo, Santa Ynez Mts.	felsic metatuff
1-4	A	Camino Cielo, Santa Ynez Mts.	brown felsic metavolc.
1-5	A	Camino Cielo, Santa Ynez Mts.	brown felsic metavolc.
1-6	A	Camino Cielo, Santa Ynez Mts.	dioritic gneiss
1-7	A	Camino Cielo, Santa Ynez Mts.	green felsic metavolc.
1-9	A	Camino Cielo, Santa Ynez Mts.	black quartzite
1-10	F	Camino Cielo, Santa Ynez Mts.	leucogabbro
1-12	A	Camino Cielo, Santa Ynez Mts.	green felsic metavolc.
1-13	A	Camino Cielo, Santa Ynez Mts.	red felsic metavolc.
1-14	A	Camino Cielo, Santa Ynez Mts.	green felsic metavolc.
1-15	A	Camino Cielo, Santa Ynez Mts.	brown felsic metavolc.
1-16	A	Camino Cielo, Santa Ynez Mts.	metaandesite
1-17	A	Camino Cielo, Santa Ynez Mts.	metaandesite
1-18	A	Camino Cielo, Santa Ynez Mts.	metaandesite
1-19	F	Camino Cielo, Santa Ynez Mts.	metabasalt
1-20	F	Camino Cielo, Santa Ynez Mts.	metabasalt
1-21	F	Camino Cielo, Santa Ynez Mts.	metabasalt
20-1	C	Redrock Camp, Santa Maria basin	lithic arenite
20-1	C	Redrock Camp, Santa Maria basin	leucogabbro

Chapter I

Reconnaissance Bulk-Rock and Clay Mineralogies of Argillaceous Great Valley and Franciscan Strata, Santa Maria Basin Province, California

By RICHARD M. POLLASTRO, HUGH MCLEAN, and LAURA L. ZINK

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FIGURES

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Reconnaissance Bulk-Rock and Clay Mineralogies of Argillaceous Great Valley and Franciscan Strata, Santa Maria Basin Province, California

By Richard M. Pollastro, Hugh McLean, *and* Laura L. Zink

Abstract

X-ray powder diffraction (XRD) analyses of samples from the Santa Maria basin province, Calif., of argillaceous rocks from the lower part of the Great Valley sequence (Upper Jurassic to lower Upper Cretaceous) and mélange of the Franciscan assemblage are similar in general bulk-mineral composition, consisting mostly of quartz, feldspar, and clay minerals. Although quartz-feldspar-clay (QFC) contents of these argillaceous rocks generally overlap and have almost identical mean QFC compositions, they differ somewhat in the amount of potassium feldspar. The bulk-mineral relations of the argillaceous rocks are thus similar to those reported for interbedded sandstones.

Although the clay contents of Great Valley sequence and Franciscan assemblage argillites are quite similar in the Santa Maria basin, the clay mineral assemblages are somewhat different. Argillites in the lower and upper parts of the Great Valley sequence contain a similar, mostly illitic, clay-mineral suite of discrete illite, mixed-layer illite/smectite (I/S), and iron-rich chlorite, with rarer kaolinite and mixed-layer chlorite/smectite (C/S). In contrast, C/S is a common and abundant constituent of argillaceous samples from the Franciscan assemblage. Textural analysis suggests that some clay was changed or neoformed during burial diagenesis. The main difference in clay mineralogy where Great Valley argillites are more illitic and Franciscan argillites more chloritic in Santa Maria basin samples, however, probably reflects different primary source materials. A much greater contribution of mafic igneous detritus, probably altered ophiolite, was recognized for the argillaceous beds and mélange matrix of the Franciscan assemblage in the basin.

INTRODUCTION

Upper Jurassic to Upper Cretaceous submarine fan deposits composed of sandstone, shale, and conglomerate called the Great Valley sequence by Bailey and others (1964), and more recently called the Great Valley Group

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by Ingersoll (1990), are widely exposed in the southern Coast Ranges of California and in the basement of the Santa Maria basin (SMB). Petrographic and provenance studies of Great Valley sequence sandstones exposed along the west side of the Sacramento and San Joaquin Valleys (Ojakangas, 1968; Dickinson and Rich, 1972; Ingersoll, 1978) and of rocks of correlative age in the Coast Ranges that include basement rocks of the SMB (Gilbert and Dickinson, 1970; MacKinnon, 1978; Nelson, 1979; Gray, 1980; Toyne, 1987) generally recognize a number of sandstone petrofacies that reflect the progressive unroofing of a late Mesozoic volcanic/plutonic complex of batholithic proportions.

Sandstones of the lower part of the Great Valley sequence (lower Great Valley sequence petrofacies of McLean, 1991) in and adjacent to the Santa Maria basin range in age from Tithonian to approximately Cenomanian. Sandstones in the lower part of the Great Valley sequence are characterized by abundant chert and mafic volcanic rock fragments, subordinate quartz, plagioclase, and trace amounts of potassium feldspar (K-feldspar). Interbedded conglomerate is generally composed of well-rounded, dark-colored chert pebbles and mafic volcanic rock fragments (Seiders, 1983). Conversely, sandstones in the upper part of the Great Valley sequence (upper Great Valley sequence petrofacies of McLean, 1991) range in age from approximately Turonian to Maastrichtian and are characterized by abundant quartz, plagioclase, and K-feldspar (Lee-Wong and Howell, 1977; Dickinson and others, 1982). Rock fragments include an assortment of granitic, intermediate volcanic, and schistose metamorphic rock fragments. Upper Cretaceous conglomerate populations vary widely, although intermediate volcanic porphyries, quartzite, and granitoid clasts predominate (Howell and others, 1977).

Other lithotectonic elements of Mesozoic age in the SMB and vicinity include mélange of the Franciscan assemblage (also called Franciscan Complex by some workers) and blocks of ophiolite that locally are conformably overlain by the lower part of the Great Valley

sequence (Hsü, 1968; Hopson and Frano, 1977). Franciscan assemblage mélange contains a diverse assortment of chert, greenstone, sandstone, metagraywacke, and subordinate quantities of schist (including glaucophane schist) dispersed in sheared argillaceous matrix. The ophiolite suite includes a variety of ultramafic rocks, pillow lavas, and chert that locally (Point Sal and Stanley Mountain) grade upward into the lower part of the Great Valley sequence.

McLean (1991) found that "knockers" composed of sandstone and metagraywacke in Franciscan assemblage mélange around the margin of the Santa Maria basin uniformly lack K-feldspar and that their quartz-feldspar-lithic (QFL) detrital modes tend to overlap those of the sandstones of the upper and lower parts of the Great Valley sequence. The absence of K-feldspar noted by McLean (1991) is an important characteristic of Franciscan sandstones and serves as a valuable petrographic tool in differentiating sandstones of the lower and upper parts of the Great Valley sequence from those of the Franciscan assemblage. Furthermore, textural and mineralogical similarities suggest that Franciscan sandstones may represent variably metamorphosed Upper Jurassic to Upper Cretaceous Great Valley sequence rocks that were initially deposited on an active accretionary continental margin and subsequently incorporated into mélange.

Clay minerals are particularly useful in understanding both the depositional and diagenetic environments of sedimentary rocks and basins (Chamley, 1989). Many clay minerals are sensitive to the often changing geologic environment and are commonly good indicators of particular physical, chemical, or thermal conditions. Other clay minerals, however, are relatively stable under a wide range or variety of conditions, often with little or no change after sedimentary recycling. Of particular interest and importance are the mixed-layer clay minerals because they are commonly products of burial-diagenetic and hydrothermal reactions (Dunoyer de Segonzac, 1970; Hoffman and Hower, 1979). The most common mixed-layer clay minerals are illite/smectite (I/S) and chlorite/smectite (C/S); however, C/S is much less common and much less understood than I/S (Reynolds, 1988; Schiffman and Fridleifsson, 1991). For example, C/S is a common product of the alteration of basic igneous rocks, particularly ophiolite complexes (Evarts and Schiffman, 1983; Bevins and others, 1991). Thus, the presence of mafic-rock material in the various petrofacies may determine the occurrence of C/S in sandstones and(or) adjacent shales.

Several studies have addressed the petrologic aspects of sandstone and conglomerates of the upper and lower Great Valley and Franciscan petrofacies. Few studies, however, have addressed the composition of argillaceous interbeds within these units. We present here a reconnaissance study of the mineralogic composition of shale, argillaceous interbeds, and mélange matrix in the upper and lower Great Valley sequence and Franciscan sandstone knockers using X-ray powder diffraction (XRD). In particular, the clay-mineral assemblages of these argillaceous materials were studied to aid in distinguishing rocks of the lower and upper parts of the Great Valley sequence and Franciscan assemblage and their provenance.

Acknowledgments

METHODS

Bulk-rock and <2 μm clay mineralogies were determined by XRD on 20 shale samples from outcrops at various sites around the SMB's margin and 23 shale samples from cores of exploratory wells in the SMB. Cores were obtained from the repositories of Unocal Corporation (Brea, California) and the State of California (California State University, Bakersfield). Shale samples are grouped according to the three corresponding sandstone petrofacies, as defined by McLean (1991), and represent the upper and lower parts of the Great Valley sequence and the Franciscan assemblage. Samples were scrubbed to remove surficial contaminants, dried, and then ground to <35 mesh. Each sample was then split with a Jones splitter into two portions for (1) whole-rock XRD and (2) carbonate dissolution and clay mineral analysis. Carbonate was dissolved in 1N HCl and the residue filtered, washed, dried, and weighed (Pollastro, 1977). The weight percents of the acid-insoluble residue and (or) acid-soluble carbonate were then determined.

Qualitative identification and semiquantitative estimates of the minerals in bulk-rock samples were made using XRD analysis of randomly oriented powders that were ground to a maximum grain size of 44 μm (<325 mesh) and packed into aluminum specimen holders from the back side. A fine, random texture was imparted onto the surface to be irradiated in order to further disrupt any preferred orientation created while mounting the sample (see Schultz, 1978). Semiquantitative weight-percent values for total clay (phyllosilicates) and other minerals or mineral groups were calculated by comparison with several prepared mixtures of minerals having similar XRD character-

Table 1. Semiquantitative X-ray powder diffraction analyses of bulk-rock and <2 µm clay minerals in argillaceous samples from outcrop, Santa Maria basin, Calif.

[Data reported in weight percent. Carb., carbonate as calcite unless suffixed by D, dolomite; plag, plagioclase; K-spar, potassium feldspar; I/S, mixed-layer illite/smectite; chl, chlorite; kaol, kaolinite; C/S, corrensite; leaders (--), none detected; tr, trace. Suffix P in "other" category is pyrite]

| Sample No. | Location | Bulk-rock | | | | | | <2 µm clay | | | | |
		Quartz	Clay	Carb	Plag	K-spar	Other	Illite	I/S	Chl	Kaol	C/S
Upper part of the Great Valley sequence												
L87-3-1	Cuyama Gorge	15	75	--	7	3	--	45	24	31	--	--
987-2-1	Highway 46	18	60	tr	12	10	--	12	52	36	--	--
987-2-2	Highway 46	16	65	--	12	6	--	12	32	56	--	--
581-20-2B	Carrie Creek	18	56	--	22	4	--	8	89	3	--	--
586-7-3	Hi Mountain saddle	24	53	--	15	8	--	11	77	12	--	--
M377-17-11	Salsipeudes Creek	--	--	--	--	--	--	25	45	12	18	--
Lower part of the Great Valley sequence												
M585-17-1A	Rinconada Creek	--	--	--	--	--	--	51	44	5	--	--
586-7-1	Rinconada Creek	16	57	--	27	--	--	6	35	40	--	19
586-7-2	Hi Mountain Saddle	25	65	--	10	--	--	57	30	13	--	--
587-14-1	West Bald Mountain	25	66	--	9	1	--	--	--	--	--	--
L87-6	Canada Honda	17	77	--	6	2	--	21	71	4	4	--
L87-7	Canada Honda	12	62	3	15	--	--	12	50	38	--	--
L87-11	Canada Honda	12	76	--	11	--	1P	13	55	22	10	--
581-20-1B	Dear Trail Mine	25	68	--	7	--	--	36	55	4	5	--
987-2-4	Toro Formation	21	61	--	14	4	--	14	41	45	--	--
Franciscan assemblage												
987-2-3	Hwy 46	17	72	1	8	2	--	6	20	27	--	47
987-2-5	Vista Point	13	55	--	30	2	--	37	28	20	6	9
987-2-6	melange matrix	16	71	1	9	4	--	17	12	32	--	39
987-2-7	Highway 1, Gorda	21	71	4D	4	--	--	54	1	45	--	--
L87-2g	Cuyama Gorge	14	66	--	18	2	--	51	24	25	--	--

istics using the procedures outlined by Schultz (1964) and Hoffman (1976), as modified by Pollastro (1985). The semiquantitative relative weight percent values calculated for total carbonate minerals by XRD were then compared with those determined using chemical dissolution.

Oriented clay aggregates of the <2-µm and <0.25-µm (equivalent spherical diameter) fractions were prepared using a modified filter-membrane peel technique (Pollastro, 1982), similar to that described by Drever (1973). Semiquantitative XRD analysis of the clay-sized fractions were made using the method of Schultz (1964), as modified by Pollastro (1985). Ratios and ordering of I/S and C/S clay were determined on oriented, ethylene glycol-saturated specimens of both the <2-µm and <0.25-µm fractions using the methods of Reynolds and Hower (1970) and Moore and Reynolds (1989).

Selected samples from outcrops and well cores were also prepared for reconnaissance textural and petrologic analysis using the scanning electron microscope (SEM) to aid in interpreting detrital versus authigenic origin of individual mineral phases.

RESULTS

X-ray Powder Diffraction

Bulk-Rock Mineralogy

Semiquantitative bulk-rock XRD analysis of samples from the lower and upper parts of the Great Valley sequence and the Franciscan assemblage are listed and grouped in tables 1 and 2. Selected XRD profiles of shale from each of these three groups are also shown in figures 1 through 5. The bulk-rock XRD data are also simplified into three components, quartz, feldspar, and clay (QFC), which are normalized and plotted in the ternary diagram of figure 6; QFC compositions for each petrofacies are also shown by different symbols. Samples containing ≤5 weight percent carbonate were normalized to 100 percent after exclusion of the carbonate; three core samples having >5 weight percent carbonate are not included in figure 6.

Generally, QFC contents of the argillaceous rocks from the three petrofacies plot within the same compositional area of figure 6; all sample groups also have a similar

Table 2. Semiquantitative X-ray powder diffraction analyses of bulk-rock and <2 µm clay minerals in argillaceous samples from well-core samples, Santa Maria basin, Calif.

[Data reported in weight percent. Depth in feet. Carb, calcite; plag, plagioclase; K-spar, potassium feldspar; I/S, mixed-layer illite/smectite; chl, chlorite; kaol, kaolinite; C/S, corrensite; leaders (--), none detected; tr, trace. Suffixes in "other" category are P, pyrite and G, gypsum]

| Well Name | Depth | Bulk-rock | | | | | | <2 µm clay | | | | |
		Quartz	Clay	Carb	Plag	K-spar	Other	Illite	I/S	Chl	Kaol	C/S
Upper part of the Great Valley sequence												
Union Pezzoni 1-------	1,270	19	74	1	4	2	--	12	45	17	26	--
Union Pezzoni 1------	1,986-1,997	23	68	--	9	--	--	15	58	27		--
Union Moretti 1-------	2,370	19	56	15	11	1	--	9	58	33	--	--
Standard Sudden 1 -----	2,598	16	64	--	20	5	--	8	62	22	8	--
Continental Leroy 5--	2,630-2,638	19	67	4	9	--	1	16	51	28	--	5
	2,951-2,966	14	76	--	10	--	--	35	45	20	--	--
	3,207-3,222	19	60	--	16	5	--	21	56	23	--	--
	3,211-3,219	21	66	1	11	1	1	40	30	30	--	--
Union Stinson 1-------	5,126	9	78	--	8	--	3P, 3G	59	35	3	3	--
Texaco Petan 2---------	1,818-1,831	21	51	2	23	3	--	6	5	35	--	54
Lower part of the Great Valley sequence												
Murphey Sinclair 74-6	1,093	10	76	--	12	2	--	35	32	33	--	--
Welch Janeway 1------	5,694	18	55	3	23	1	--	17	17	45	--	21
Shell Barret 1 ---------	1,452	24	51	6	19	--	--	20	45	35	--	--
Amerada Tognazzi-----	8,602-8,618	--------no bulk sample------						21	33	46	--	--
Franciscan assemblage												
Superior Silva CH1 ---	916-922	16	69	1	14	--	--	25	25	50	--	--
Union W. Bradley 1---	2,792	18	55	--	25	2	--	28	32	40	--	--
	4,400	18	66	1	15	--	--	24	33	43	--	--
Superior Beck CH1----	1,717-1,721	14	40	5	36	--	--	3	--	45	--	52
Union Wineman 2-----	3,650-3,658	17	68	--	12	3	--	6	30	40	--	24
	3,658-3,665	21	71	1	4	3	--	2	6	15	--	77
Texaco Fulwider 1-----	3,458	24	57	9	8	2	--	2	7	44	--	48
Welch Janeway 1------	5,922	23	59	--	17	1	--	17	26	29	--	28
Union Phelan 1--------	3,591-3,600	19	66	--	10	2	3P	15	34	35	-	16
Chief Larson Brothers	2,784	17	75	--	5	3	--	19	48	21	12	--
Ericson Dart 1----------	3,200	16	66	3	12	--	1P	10	18	43	--	29

compositional range. Mean QFC compositions for the three groups are similar, having about 65 weight percent total clay (as phyllosilicates), 20 weight percent quartz, and 15 weight percent total feldspar, and thus can not be distinguished from one another on this basis (fig. 6). Type and amount of feldspar and phyllosilicates (tables 1 and 2), however, show some differences among the three groups.

Shale interbeds from the upper part of the Great Valley sequence contain the greatest amount of K-feldspar, having a mean of >3 weight percent; average K-feldspar contents for samples of shale from the lower part of the Great Valley sequence and the Franciscan assemblage are 0.8 and 1.2 weight percent, respectively. Similarly, petrographic studies by McLean (1991) found that sandstones of the upper part of the Great Valley sequence contain the greatest amount of K-feldspar by volume. Moreover, McLean (1991) found that the average volume percent K-feldspar was higher in the upper Great Valley sandstones

from outcrop (about 10 volume percent) than in well-core (about 7 volume percent) samples. Similarly, K-feldspar contents in upper Great Valley argillaceous rocks of this study average about 6 weight percent in samples from outcrop and <2 weight percent in those from well cores.

Carbonate is low to absent in most samples, averaging <2 weight percent of the bulk rock; it is present mostly as calcite in core samples (tables 1, 2; figs. 1, 2, 4). Plagioclase is more abundant than K-feldspar. Coarse micas (muscovite and(or) biotite; figures 1A, 2B) and chlorite (figs. 1 through 5) are also commonly present.

Clay Mineralogy

Qualitative and semiquantitative clay mineralogies of the <2-µm fraction are listed in tables 1 and 2. The clay data are also combined and plotted on the ternary diagram
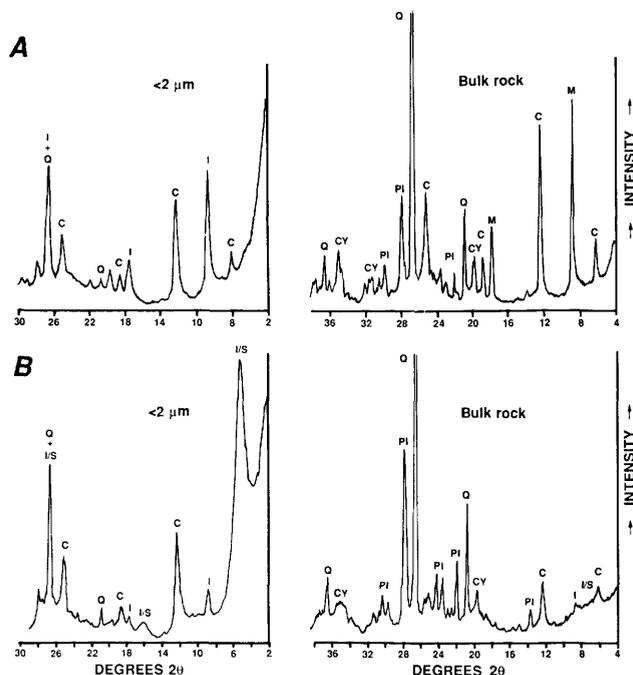
Figure 1. X-ray powder diffraction profiles of random-powder bulk rock (right) and oriented glycol-saturated <2 μm fractions (left) of argillaceous samples from upper part of Great Valley sequence. A, Outcrop sample L87-3-1 from Cuyama Gorge. B, Core sample from Standard Sudden well at depth of 2,599 ft. CuK$_\alpha$ radiation. Q, quartz; CY, total clay (combined clay peaks); M, coarse mica; C, chlorite; Pl, plagioclase; I, illite; I/S, illite/smectite.
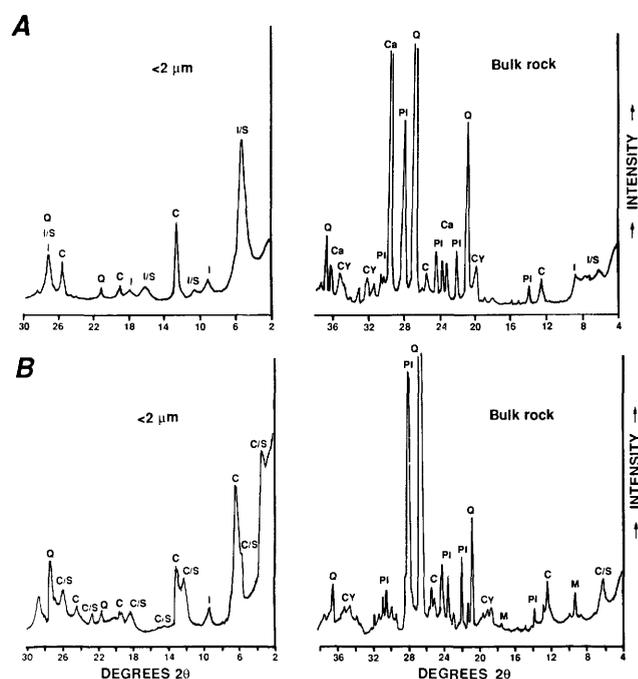

Figure 2. X-ray powder diffraction profiles of random-powder bulk rock (right) and oriented glycol-saturated <2 μm fractions (left) of argillaceous samples from core from upper part of Great Valley sequence. A, Union Moretti 1 well at 2,370 ft. B, Texaco Petan 2 well at 1,818 to 1,831 ft. CuK$_\alpha$ radiation. Q, quartz; CY, total clay; M, coarse mica; C, chlorite; Pl, plagioclase; I, illite; I/S, illite/smectite; Ca, calcite; C/S, chlorite/smectite (corrensite).

of figure 7. There is considerable variation in clay mineralogy for each of the three petrofacies from both outcrop and core samples; few diagnostic relations, however, are immediately evident from the data.

The clay minerals in these rocks are mainly discrete illite and chlorite, I/S, kaolinite, and C/S (mostly as corrensite). All samples contain illite and chlorite with varying amounts of I/S. All discrete chlorites have stronger (higher relative intensity) 002 and 004 basal reflections than 001, 003, and 005 reflections, indicating that they are iron rich (Brindley, 1961; Moore and Reynolds, 1989). The ratio and ordering of I/S varies from sample to sample. For example, upper Great Valley core samples shown in figures 1B and 2A and the Franciscan core sample shown in figure 4A contain mainly smectitic I/S (R0 I/S with 50 to 80 percent smectite layers), whereas the lower Great Valley core sample shown in figure 3B contains illitic, ordered I/S (R1 I/S with 70 to 75 percent illite layers). Additionally, a heterogeneous assemblage of I/S is commonly present within a single sample (fig. 4B).

Figure 7 compares the clay mineralogy of the <2-μm fraction for each sample and petrofacies. The categories forming the points of the ternary diagram of figure 7 were selected on the basis of mineralogical associations determined from XRD and textural observations from reconnaissance SEM analysis. On average, the clay mineralogy
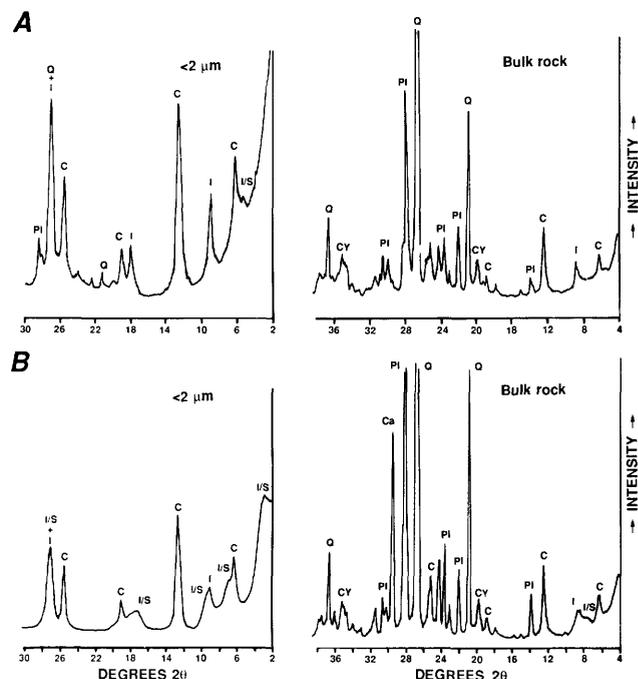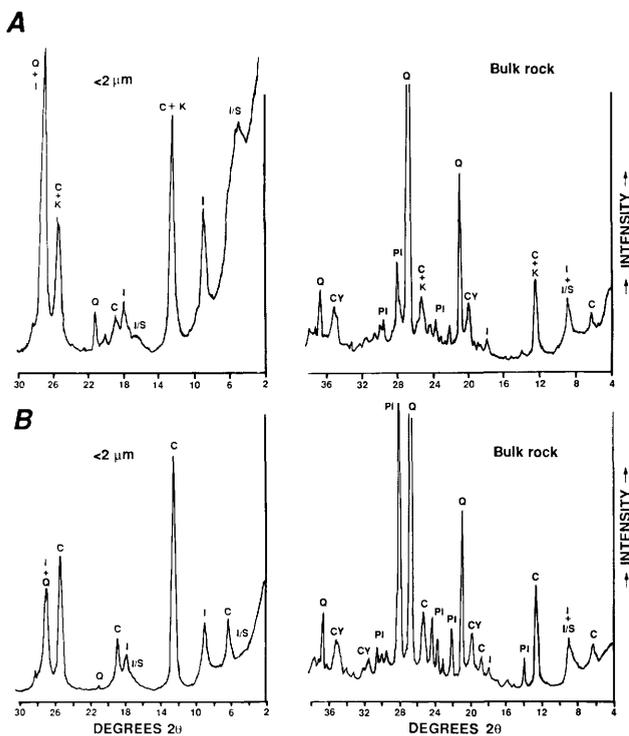

Figure 3. X-ray powder diffraction profiles of random-powder bulk rock (right) and oriented glycol-saturated <2 μm fractions (left) of argillaceous samples from lower part of Great Valley sequence. A, Outcrop sample 987-2-4 on Highway 46 (Toro? Formation). B, Core sample from Shell Barret 1 well at 1,452 to 1,464 ft. CuK$_\alpha$ radiation. Q, quartz; CY, total clay; M, coarse mica; C, chlorite; Pl, plagioclase; I, illite; I/S, illite/smectite.

Figure 4. X-ray powder diffraction profiles of random-powder bulk rock (right) and oriented glycol-saturated <2 μm fractions (left) of argillaceous samples from well cores in Franciscan assemblage. A, Chief Larson Brothers well at 2,784 to 2,804 ft. B, Superior Silva CH1 well at 916 to 922 ft. CuK$_\alpha$ radiation. Q, quartz; CY, total clay; M, coarse mica; C, chlorite; Pl, plagioclase; I, illite; I/S, illite/smectite.
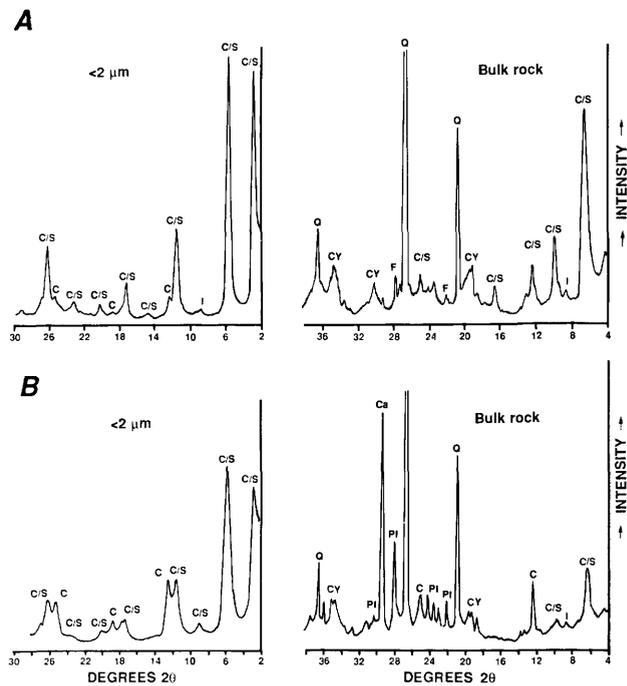
Figure 5. X-ray powder diffraction profiles of random-powder bulk rock (right) and oriented glycol-saturated <2 μm fractions (left) of argillaceous samples from a well core in Franciscan assemblage containing abundant chlorite/smectite (C/S) as corrensite. A,Union Wineman 2 at 3,658 to 3,665 ft. B, Texaco Fulwider 1 well at 3,458 ft. CuK$_\alpha$ radiation. Q, quartz; CY, total clay; M, coarse mica; C, chlorite; F, plagioclase and potassium feldspars; Pl, plagioclase; I, illite; Ca, calcite.
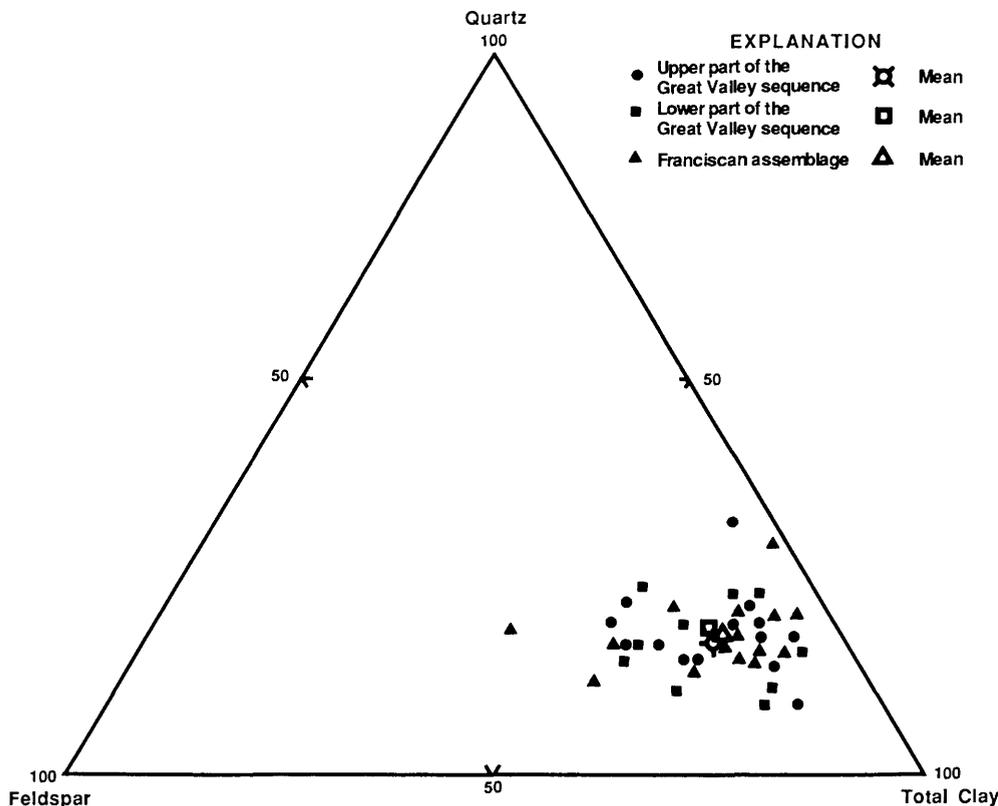

Figure 6. Ternary diagram showing bulk-rock quartz/feldspar/total-clay contents (in weight percent) of argillaceous outcrop and well-core samples from Franciscan assemblage and upper and lower parts of Great Valley sequence. Data normalized on carbonate-free basis from tables 1 and 2. Mean values for each group designated by open symbols.

of upper Great Valley and lower Great Valley shale is similar, and they have an overall average composition of about 55 relative weight percent illite plus chlorite and 45 weight percent I/S; kaolinite and C/S are less abundant. As mentioned previously, the I/S amount and ordering type varies greatly among samples (figs. 1, 4, and 7). Lower Great Valley shale samples commonly contain more illite and chlorite than those from upper Great Valley shale samples; upper Great Valley shale samples generally contain more I/S. Additionally, I/S in lower Great Valley argillites is commonly more illitic than that in samples from the upper part of the Great Valley sequence. An example of this relation is shown in the samples from the Hi Mountain Saddle outcrop (table 1). Note that upper Great Valley sample 586-7-3 contains 77 relative weight percent I/S as compared to 30 weight percent I/S in lower Great Valley sample 586-7-2. This relation is also shown by the abundance of solid circles in the upper portion of figure 7. Moreover, those samples with the least amount of I/S commonly contain ordered I/S (fig. 7).

Although some samples of argillite and shale from the Franciscan assemblage have a clay-mineral suite similar to those from the Great Valley sequence, more commonly the clay minerals in Franciscan assemblage samples are markedly different. This difference is shown by the individual sample and mean clay compositions plotted on figure 7. The primary difference in clay mineralogy in argillitic samples from the Franciscan assemblage compared to those from Great Valley sequence shales is the common presence and abundance of C/S (tables 1 and 2; fig. 5). An extreme case is demonstrated by the Franciscan core sample from the Union Wineman 2 well at depths of 3,658 to 3,665 ft (table 2; fig. 5A). This shale sample consists of 71 weight percent total clay with C/S (as corrensite) constituting 77 weight percent of the clay in the <2-μm fraction. Also, Franciscan samples are generally more chloritic, whereas Great Valley samples are more illitic. On weighted average, the <2-μm fraction of Franciscan shales is composed of about 58 percent chloritic clays (chlorite plus C/S), 41 percent illitic clays (illite plus I/S), and 1 percent kaolinite, as compared to the weighted average for shale samples from the Great Valley sequence, which average about 31 percent chloritic clays, 67 percent illitic clays, and 2 percent kaolinite. From this standpoint, the mean clay-mineral composition of shales from the upper and lower petrofacies of the Great Valley are nearly identical.
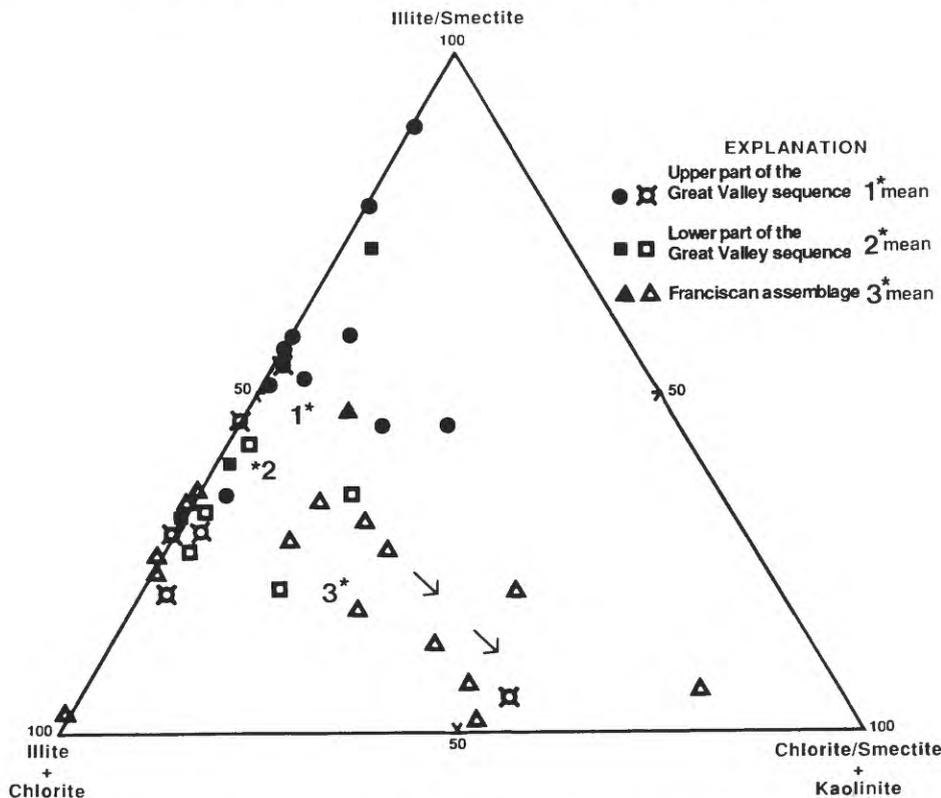

Figure 7. Ternary diagram showing clay-mineral contents (in weight percent) in <2 μm fraction of argillaceous outcrop and well-core samples from the Franciscan assemblage and upper and lower parts of Great Valley sequence. Data compiled from tables 1 and 2. Solid symbols indicate samples containing random interstratified illite/smectite (I/S), whereas open symbols are those samples containing only ordered I/S and (or) chlorite/smectite (C/S). Mean values for each group are designated by numbered asterisks. Arrows indicate general direction of increasing C/S with decreasing I/S.

Observations from SEM Analysis

SEM analysis of freshly broken surfaces of shale and argillite samples from the Great Valley sequence and Franciscan assemblage show both detrital and authigenic textures on and oblique to bedding planes. These observations apply to samples from both outcrop and well-core samples. For example, scanning electron micrographs of core samples of the Franciscan assemblage from the Union Wineman 2 well at depths ranging from 3,650 to 3,658 ft are shown in figures 8 and 9. Samples sometimes appear blocky or massive, especially when viewed perpendicular to bedding (fig. 8A). Layered and microgranular clay fabrics containing individual flakes and plates of clay are commonly observed (figs. 8 and 9), particularly when viewed parallel or oblique to bedding.

Of particular interest are textural characteristics indicative of origin for corrensite and I/S. The fine-grained nature of most samples, the dense, compacted, blocky-to-bedded and granular fabrics, and the anhedral, ragged, platy clay habit (fig. 8) suggest a detrital origin for most of the clay minerals in these argillaceous samples. Coarser, nonclay, detrital mineral grains are also commonly dispersed throughout most samples (figs. 8B, D).

Crystal habits indicative of authigenic clay-mineral formation were also commonly observed along bedding planes, microfractures, and in micropores. For example, some authigenic growth of I/S is evidenced by the presence of honeycomb- and flamelike structures on I/S substrates and along edges of I/S having a detrital cornflake habit (fig. 9A to C) in core samples from the upper part of the Great Valley sequence. Samples containing both I/S and correns-

200 µm

2 µm

10 µm
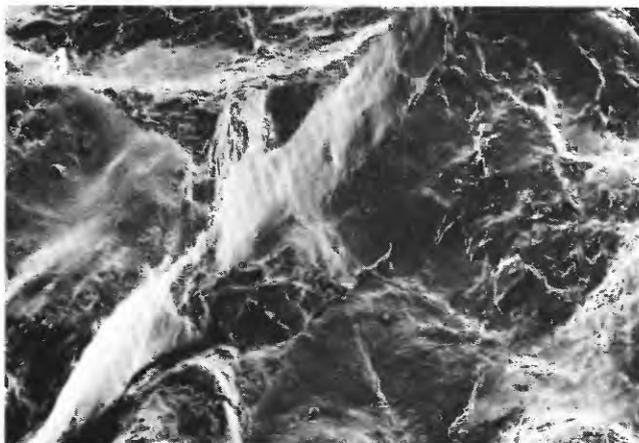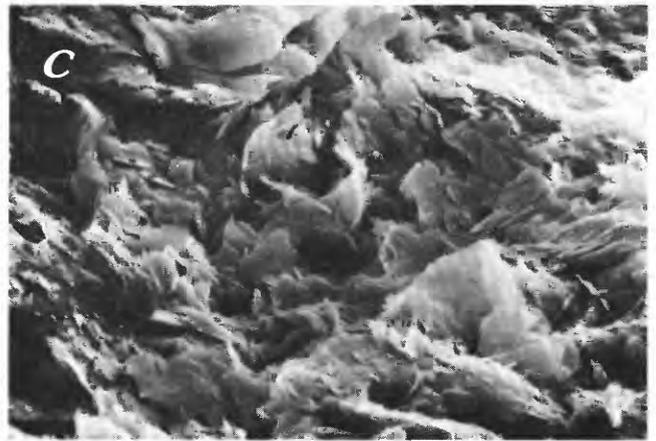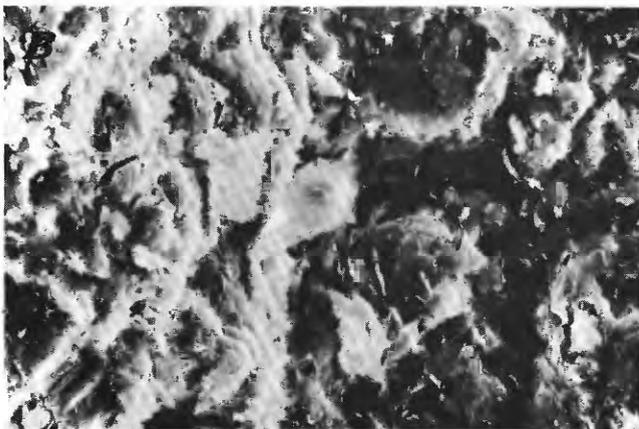
10 µm

Figure 8. Scanning electron micrographs of mostly detrital textures in argillaceous rocks from the Franciscan assemblage. A, Blocky or massive texture shown perpendicular to bedding in core sample from Union Wineman 2 well at 3,650 ft. B, Ragged platy to granular heterogeneous fabric of chloritic clay in core sample from Superior Beckett CH 1 well at 1,717 ft. C, Thin platy semibedded fabric of chloritic clay in Union Wineman 2 well at 3,650 ft. D, Flaky heterogeneous clay fabric with dispersed coarse mineral grains (arrows). Note authigenic clay coating on coarse grain at left.

ite also contained authigenic lath-, fiber-, and step-like overgrowths (fig. 9D). These overgrowth features are typical of authigenic-clay formation by burial diagenesis (Pollastro, 1985) and hydrothermal processes (Inoue and others, 1987). Coarser mineral grains and cements are also commonly observed in samples; these are sometimes partly altered or replaced by authigenic clay (fig. 8D).

The standardless, normalized, energy-dispersive chemistry of the corrensite- and chlorite-rich Franciscan mélange matrix outcrop sample 987-2-3 (table 1) and a scanning electron micrograph of the area of this sample analyzed are shown in table 3 and figure 10, respectively. These data support the interpretation from XRD analyses that chlorites are iron enriched (relative to magnesium).

DISCUSSION

Petrographic studies of the detrital modes in sandstones from the upper and lower parts of the Great Valley sequence found that upper Great Valley sandstones contain greater amounts of detrital quartz, K-feldspar, and biotite than sandstones from the lower part of the Great Valley sequence (Dickinson and others, 1982; Gray, 1980; McLean, 1991). Although detrital modes of individual lithic clasts or grains can not be determined by XRD analysis, bulk XRD compositions of argillites and shale determined on the bases of QFC contents are similar for all three groups shown in figure 6. XRD data does, however, show greater mean K-feldspar content in argillites designated as
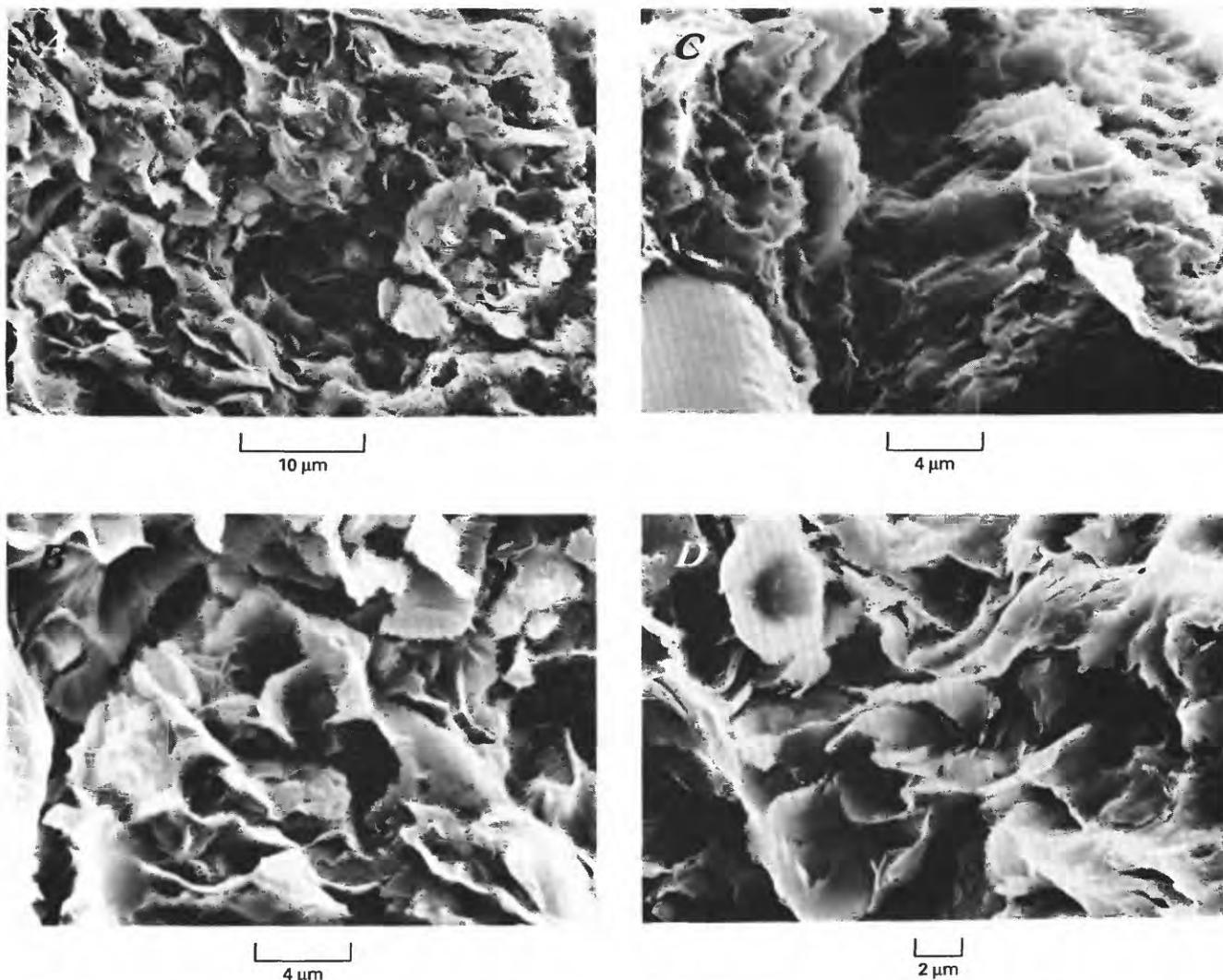
Figure 9. Scanning electron micrographs of authigenic clay textures in argillaceous rocks from upper Great Valley sequence (A through C) and Franciscan assemblage (D). A, B, Honeycomb, flamelike, and wrinkled-sheet overgrowths and coatings (arrows) in illite/smectite-rich microporous core sample from Union Moretti 1 well at 2,370 ft. C, Honeycomb- and flamelike textures in illite/smectite-rich core sample from Union Pezzoni 1 well at 1,986 ft. D, Lath-, fiber-, and steplike overgrowths in core from Union Wineman 2 well at 3,650 ft.

representing the upper part of the Great Valley sequence, particularly in those from outcrop.

It is important to note that there is a wide range in diagenetic histories for the samples studied because (1) the diversity in location of outcrops and cores in the SMB, (2) local variations in basin tectonics and burial histories, and (3) variation in present and (or) maximum burial depth of well-core samples. Thus, any interpretation of detrital clay mineralogy is complicated by the likely modification, transformation, or destruction of the original detrital assemblage (particularly clay minerals) and (or) the neoformation of coarse-grained cements and clay minerals as a consequence of burial.

The mean clay mineral compositions of the <2-μm fraction of argillites from the upper and lower parts of the Great Valley sequence are similar, with only slight differences in the amount and ordering type of I/S. Moreover, the lack of clay mineral data from interbedded sandstones

Table 3. Standardless energy-dispersive analysis normalized to 100 percent for area of sample 987-2-3 shown in figure 10

[Sample 967-2-3 listed in table 1]

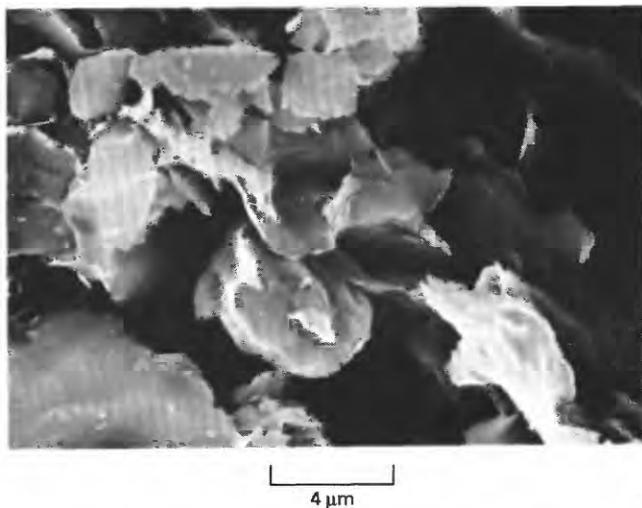| Oxide compound	Weight percent
Al_2O_3 -------------	10.0
SiO_2 -------------	45.0
K_2O -------------	.5
FeO -------------	30.5
CaO -------------	2.4
MgO -------------	11.9
Total ---------	100.0

4 μm

Figure 10. Scanning electron micrograph of chloritic clay minerals from mélange matrix of Franciscan assemblage from outcrop on Highway 46 (sample 987-2-3, table 1). Corresponding standardless energy-dispersive analysis normalized to 100 percent is listed in table 3.

and their association with the hydrothermally metamorphosed Point Sal ophiolite (Hopson and Frano, 1977) caused further problems in distinguishing between detrital and authigenic assemblages and interpreting the diagenetic grade of the chloritic clay minerals.

Illite and chlorite are present in all samples. Interbedded sandstones contain coarse white micas, biotite, and chlorite as primary grains or within rock fragments (McLean, 1991). Thus, much of the illite and chlorite may be recycled detrital material from older eroded sedimentary and(or) metamorphic rocks. Illite commonly is highly ordered (fig. 2A), as defined by the ratio of XRD peak height and width (Kubler, 1968), indicating that some illite may have formed at high (>200°C) temperatures, an indication which is consistent with high-diagenetic/low-metamorphic grades. In some samples, highly expandable I/S coexists with illite of high crystallinity. These coexisting phases indicate a wide contrast in thermal histories (Hoffman and Hower, 1979) and strongly suggest a recycled origin for the well-ordered illite in many of the samples (Pollastro, 1990; 1993). Moreover, textural observations from SEM also suggest a detrital origin for most of the clay.

Some discrete clay-sized chlorite may have formed from the alteration of basaltic or ophiolitic rocks and rock fragments (Schiffman and Fridleifsson, 1991) prior to deposition. Mafic volcanic rock fragments are particularly common in Franciscan graywackes and mélange (Dickinson and others, 1982), as well as in some sandstones from the lower part of the Great Valley sequence (MacKinnon, 1978). Also, some authigenic chlorite and illite may have formed from diagenetic clay-mineral reactions in those samples that have undergone burial to depths >4,000 m and (or) temperatures in excess of 200°C (Hower, 1981).

I/S is found in all samples in varying amounts, often varying in illite/smectite ratio and ordering types. These differences in I/S are commonly due to variations in (1) detrital source areas and materials, (2) burial histories, and (3) diagenetic origins of I/S (Hoffman and Hower, 1979; Rettke, 1981; Pollastro, 1990; 1993). Upper Great Valley argillites tend to contain a greater amount of I/S with higher smectite content, whereas argillites from the lower part of the Great Valley sequence have a higher illite content and more illitic I/S. The general relation shown in figure 7, where samples containing the least amount of I/S contain ordered I/S, suggests that illitization has occurred in these rocks through forming illite layers in I/S and some discrete illite, in part, by cannibalization of earlier smectite or I/S (Pollastro, 1985). The coexistence of C/S (mostly corrensite) with illitic ordered I/S provides dual clay geothermometers suggesting paleoburial temperatures in excess of 100°C (Hoffman and Hower, 1979). Additionally, figure 7 shows that there are progressively decreasing amounts of I/S, along with illite and chlorite, with increasing amounts of C/S (plus kaolinite), an observation which suggests that some C/S may form at the expense of I/S.

However, these relations are difficult to test because the formation of C/S may be strongly dependent on precursor source materials and not on postdepositional conditions, as will be discussed below.

Mixed-Layer Chlorite/Smectite (Corrensite)

C/S, identified here mostly as corrensite, is common in shale samples from the Franciscan assemblage; it is also present in some shale samples from the Great Valley sequence. Some XRD profiles containing C/S, however, may also be interpreted as mixed-layer chlorite/corrensite (Shau and others, 1990). Although corrensite occurs in diverse geologic settings (Reynolds, 1988), it is often a product of the hydrothermal alteration of mafic rocks (Furbish, 1975; Kristmannsdottir, 1975, 1979; Brigatti and Poppi, 1984; Evarts and Schiffman, 1983; Bettison and Schiffman, 1988; Bettison-Varga and others, 1991; Shau and others, 1990; Schiffman and Fridleifsson, 1991). Corrensite is also commonly formed in burial-diagenetic settings (Hoffman and Hower, 1979; Reynolds, 1988).

Roberson (1988), Bettison and Schiffman (1988), and Bettison-Varga and others (1991) have identified and described corrensite in hydrothermally altered basaltic rocks of the Jurassic Point Sal and Coast Range ophiolites in California. Similarly, Evarts and Schiffman (1983) have described the formation of smectite, random-ordered (R0) C/S, corrensite, and discrete chlorite clay minerals from the hydrothermal metamorphism of the Del Puerto ophiolite (Evarts, 1977) in central California. Evarts and Schiffman (1983) attributed the formation of R0 C/S, corrensite, and chlorite to the interaction between the ophiolitic rocks and heated seawater shortly after formation and emplacement of the ophiolite in the submarine environment. They concluded that any effects of diagenetic or burial metamorphic overprint on submarine hydrothermal metamorphism of the Del Puerto ophiolite was negligible. A detailed petrographic, XRD, and chemical study by Bettison and Schiffman (1988) revealed that phyllosilicates (smectite, C/S, and chlorite) of the Point Sal ophiolite are similar to the composition and zonations of those from the Del Puerto remnant studied by Evarts and Schiffman (1983). Moreover, oxygen-isotope analyses suggest that phyllosilicates were formed during the initial alteration of Point Sal volcanic rocks when water/rock ratios and pCO_2 were high.

The origin of the corrensite, as well as some smectite and chlorite, is therefore strongly dependent upon the amount of detrital input and degree of alteration of volcanogenic and ophiolitic material deposited within these argillaceous sediments. McLean (1991) reported that sandstones in the lower part of the Great Valley sequence contain rock fragments consisting mainly of aphanitic and basaltic andesite in an altered glassy groundmass of semi-

opaque clay minerals. Although no XRD analyses were performed on the clay minerals in these sandstones, textural descriptions suggest that some clays may have formed by diagenetic processes. Moreover, SEM analysis has shown that some authigenic clay formed in these rocks as overgrowths on precursor clays and by the alteration or replacement of coarser mineral grains as well. However, unless extensive diagenetic modification has occurred, significant contributions of fine-grained altered ophiolitic material may be necessary to explain the abundance of corrensite in samples such as that found in the Union Wineman 2 well at depths of 3,658 to 3,665 ft (table 2: fig. 5A).

McLean (1991) also reported the highest (3.5 volume percent) epidote content in sandstones from well samples of the lower part of the Great Valley sequence. Argillaceous samples from the lower part of the Great Valley sequence in this study also have the highest chlorite content (40 weight percent) (table 2; fig. 7). High chlorite content is consistent with an early "epidote zone" grade of metamorphism (Evarts and Schiffman, 1983) and may be indicative of the original igneous detritus. The data of our study, however, can only suggest origins for specific clay minerals through relations such as those mentioned above. Detailed studies of both sandstone and shale integrating SEM, XRD, transmission electron microscopy, and electron microprobe analyses, similar to those performed by Shau and others (1990) and Bettison-Varga and others (1991), are necessary before the origins of specific clay minerals in the argillaceous rocks of the SMB that we studied can be confidently determined.

SUMMARY

XRD mineralogy of argillaceous samples from outcrop and well-core samples of the Franciscan assemblage and the upper and lower parts of the Great Valley sequence in the SMB shows that the main constituents are quartz, feldspar, and clay minerals with much lesser amounts of carbonates, pyrite, and gypsum. Although the total quartz-feldspar-clay contents of these three rock groups overlap and all three have almost identical mean compositions, they differ somewhat in the amount of K-feldspar and in clay mineralogy. Similarly, McLean (1991) found that the quartz-feldspar-lithic detrital modes for sandstones from these three rock groups in the SMB overlapped and that variations in the amount of K-feldspar served as a petrographic tool for differentiating rocks in the lower and upper parts of the Great Valley sequence from those of the Franciscan assemblage.

The three rock groups in this study contain discrete illite and chlorite, as well as I/S; kaolinite occurs less commonly in several samples. Most discrete chlorite is iron rich. C/S (mainly corrensite) is abundant in most of

the Franciscan samples but also is present in some argillites of the Great Valley sequence. The ratio and ordering of I/S varies from sample to sample and is sometimes heterogeneous within a single sample, suggesting a variety of origins. The diversity in location of outcrops and cores, core-sample depths, and local variations in tectonic and burial histories in the SMB among the samples studied, however, complicates any interpretation of either detrital or diagenetic origins for these clay minerals.

Argillaceous rocks in the upper and lower parts of the Great Valley sequence are composed of a similar mostly illitic clay-mineral suite consisting mostly of discrete illite, I/S, and chlorite; upper Great Valley samples, however, are commonly richer in I/S with a higher smectite content. In contrast, samples from the Franciscan assemblage commonly contain more chloritic clays. In particular, the common presence and abundance of C/S as corrensite in these argillaceous rocks indicates a different source material than for similar rocks from the Great Valley sequence. Although C/S is commonly a product of burial diagenesis, it is also ubiquitous in hydrothermally altered basaltic rocks and has been reported as a common constituent of the Point Sal and Del Puerto ophiolites in the western part of southern California. Thus, C/S may have been derived from hydrothermally altered detritus and (or) the burial diagenesis of basic igneous rocks. Similarly, the presence and abundance of smectitic and chloritic clay minerals in these argillaceous rocks are dependent on the amount and degree of alteration of original detrital volcanogenic material.

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