<u>OVERVIEW</u>

Late Cenozoic tectonics of the central and southern Coast Ranges of California

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ABSTRACT

The central and southern Coast Ranges of California coincide with the broad Pacific–North American plate boundary. The ranges formed during the transform regime, but show little direct mechanical relation to strike-slip faulting. After late Miocene deformation, two recent generations of range building occurred: (1) folding and thrusting, beginning ca. 3.5 Ma and increasing at 0.4 Ma, and (2) subsequent late Quaternary uplift of the ranges. The ranges rose synchronously along the central California margin and are still rising; their long axes are quasiparallel to the plate boundary and strike-slip faults. The upper crustal internal and marginal structures of the ranges are contractional, dominated by folds and thrusts resulting from the convergent component of plate motion. Newly constructed transects using seismic reflection and refraction, plus gravity and magnetic studies, reveal lower crustal basement(s) at depths of 10–22 km. The upper surface of the basement and Moho show no effect of the folding and thrusting observed in the upper crust. We conclude that horizontal shortening is accommodated at depth by slip on subhorizontal detachments, and by ductile shear and thickening. The ranges are marked by high heat flow; weak rocks of the Franciscan subduction complex; high fluid pressure; bounding high-angle reverse, strike-slip, or thrust faults; and uplift at a rate of 1 mm/yr beginning about 0.40 Ma. Transverse compression manifested in folding within the Coast Ranges is ascribed in large part to the well-established change in plate motions at about 3.5 Ma.

INTRODUCTION

The California Coast Ranges province encompasses a system of elongate mountains and intervening valleys collectively extending southeastward from the latitude of Cape Mendocino (or beyond) to the Transverse Ranges. This paper deals with the portion of the province that lies southeast of San Francisco, a subprovince (Fig. 1) that we call the central and southern Coast Ranges, encompassing the Diablo Range, East Bay Hills, Santa Cruz Mountains, Gabilan Range, Santa Lucia Range, La Panza Range, San Rafael Range, Sierra Madre Range, Caliente Mountain, and Temblor Range (Fig. 1). Exactly the same area has been called the southern Coast Ranges in the past (e.g., Page, 1981; see also Jennings, 1977).

The central and southern Coast Ranges have been studied for more than a century, attention being largely devoted to stratigraphy, petrology, structure, and resources potential. In recent years, geophysical investigations have provided fundamental subsurface information. In this paper we examine the main Cenozoic tectonic events and the culminating rise of the ranges in their present configuration, the timing, manner of uplift, possible causes, and mechanics.

A review of the geology of the central and southern Coast Ranges and references to many studies may be found in Page (1981), and convenient summary papers are available in Wahrhaftig and Sloan (1989). Pioneering groundwork was done by such notables as A. C. Lawson (e.g., 1893, 1914), R. D. Reed (1933), and N. L. Taliaferro (e.g., 1943). A prodigious amount of geologic mapping by T. W. Dibblee, Jr., presented the areal geology in a form that made general interpretations possible. E. H. Bailey, W. P. Irwin, D. L. Jones, M. C. Blake, and R. J. McLaughlin of the U.S. Geological Survey and W. R. Dickinson are among many who have contributed enormously to the present understanding of the Coast Ranges. Representative references by these and many other individuals were cited in Page (1981). Additional significant papers are noted in this article.

Probably the earliest study that specifically focused on the uplift of the present-day ranges was that of Christensen (1965), who recognized that the Coast Ranges as we know them are younger than 3–4 Ma. Subsequent writings on uplift of the ranges include Page (1981, p. 415–416); Zandt and Furlong (1982), Montgomery (1993), Anderson (1994), and Burgmann et al. (1994). Many other authors have contributed important data, as noted herein. Our review is an attempt to integrate an earlier paper (Page, 1981) with new and significant geophysical studies that have enhanced our knowledge of the crust and mantle within the Coast Ranges.

GEOMORPHIC FEATURES OF TECTONIC SIGNIFICANCE

The ranges and major intervening structural valleys are subparallel with the continental margin and the Pacific–North America plate boundary. The San Andreas fault is parallel with some of the ranges, but is slightly oblique to the Coast Ranges province as a whole, and crosses it with a prevailing trend of about N41°W. The orientation of the ranges with respect to the plate boundary pre-

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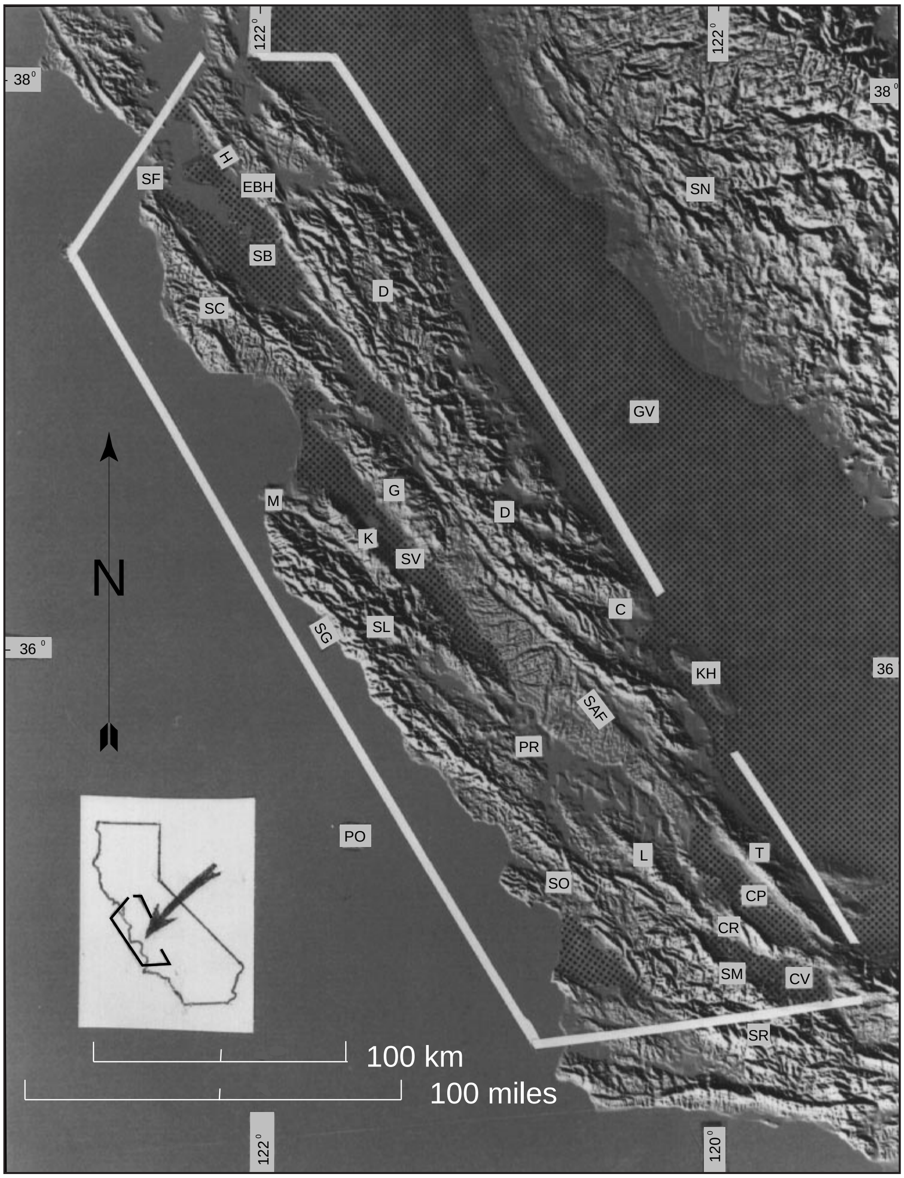


Figure 1. Physiographic map of the central and southern Coast Ranges, a subprovince (within the white polygon) of the California Coast Ranges. Inset shows location within the State of California. The subprovince approximately coincides with the broad transform plate boundary, characterized by active dextral transform faults, three of which are labeled (SAF, SG, and H). Abbreviations: C—Coalinga; CP—Carrizo Plain; CR—Caliente Range; CV—Cuyama Valley; D—Diablo Range; EBH—East Bay Hills; G—Gabilan Range; GV—Great Valley; H—Hayward fault; K—King City fault; KH—Kettleman Hills; L—La Panza Range; M—Monterey; PO—Pacific Ocean; PR—Paso Robles; SAF—San Andreas fault; SB—San Francisco Bay–Santa Clara Valley depression; SC—Santa Cruz Mountains; SF—San Francisco; SG—San Gregorio–Hosgri fault zone; SL—Santa Lucia Range; SM—Sierra Madre Mountains; SN—Sierra Nevada; SR—San Rafael Range; SO—San Luis Obispo; SV—Salinas Valley; T—Temblor Range.

cludes an origin by wrench tectonics (i.e., folding and faulting produced directly by horizontal shear such as in transform interaction), in which case folds and thrusts would be expected to form at an angle to the boundary, in an en echelon pattern.

The individual ranges are mutually similar in their trend and in their quasilevel crests, which have few conspicuous peaks (Figs. 2 and 3). They are 120 km to more than 300 km long, 10 to 50 km wide, and generally 400 to 1200 m high. The modest widths, together with geophysical data, argue against some proposed modes of origin, such as the presence of low-density roots or local causative thermal conditions associated with individual mountain ranges. The scarcity of throughgoing transverse river valleys, coupled with the weak nature of most Coast Ranges rocks, is suggestive of (but not proof of) geologically recent, ongoing uplift.

Except for their mutual parallelism and relatively level crests, the overall morphologic aspects of the several ranges differ widely. Some (e.g., the northern Santa Lucia Range, Fig. 4, and northern Gabilan Range) show eroded fault-block features including steep scarp-like flanks, faceted spurs, and V-shaped canyon mouths at the apices of alluvial fans. Several ranges, including the two just mentioned, have remnants of upland pediments or vestiges of gently rolling uplands, presumably formerly near sea level, which are sharply incised by headward-growing modern canyons. Nearly level upland remnants suggest even uplift. However, the southern part of the Gabilan Range is a tilted fault block that has deeply eroded, gently sloping upland surface remnants that dip gently toward their juncture with the floor of Salinas Valley. Some ranges (or foothill ranges, such as the one visible in Fig. 2 bordering the Santa Clara Valley south of San Jose, and the East Bay Hills) look like linear arched ductile welts that have marginal faults but do not have commanding scarps. In contrast, parts of the Diablo Range and parts of the Santa Lucia Range have somewhat depressed wide central portions bordered on both sides by crestal ridges. The northeast flank of the Diablo Range is locally accompanied by broad, gently sloping, alluviated, plateau-like tracts that have been only slightly uplifted along unspectacular marginal faults. Each of these diverse morphologic aspects must tell us something about the tectonics of the subprovince.

It is important to note that parts of the flanks of several ranges are deeply embayed by erosion and lack indications of marginal faults (e.g., the northeast flank of the Santa Lucia Range between Greenfield and Paso Robles, and much of the southwest flank of the Gabilan Range). Hence, fault slip is not always necessary for the rise of the ranges, although it has locally played a role.

The principal intermontane valleys, such as the Santa Clara, Salinas, and Cuyama Valleys, and the Carrizo Plain, are relevant to the tectonics of the province, as they are clearly of structural origin and are presumably products of the same events that caused uplift of the ranges. Their floors are aggraded rather than eroded, and they are bordered by piedmont fans. The Carrizo Plain lacks a fluvial outlet and contains a saline playa. The longitudinal drainage of the Santa Clara Valley reverses its direction of flow near the midpoint of the valley, indicating that the valley probably existed prior to, and independent of, the present drainage system. Most likely, the crustal tracts marked by these valleys were forced down or remained at a neutral level when the adjacent mountains rose.

A few streams have an antecedent relationship to ranges or parts of ranges. For example, the Pajaro River cuts across the south end of the Santa Cruz Mountains, flowing out onto a coastal alluvial plain near Watsonville, and Alameda Creek crosses the East Bay Hills in a steep-sided canyon, debouching onto an alluvial fan that descends to San Francisco Bay. Because these streams flow upon aggraded surfaces before and after crossing the mountains, they could not have been superposed from an uplifted erosion surface. The rise of the transected ranges must have been gradual enough, and the stream flow vigorous enough, to allow the survival of the transverse drainage. On the other hand, most parts of most ranges have very few through-going transverse streams and appear to have risen too rapidly to allow antecedent drainage to persist. Striking examples of rapid uplift include the East Bay Hills (even though cut by Alameda Creek) and the frontal foothill ridge of the Diablo Range southeast of San Jose (Fig. 2).

Each of the Coast Ranges is flanked by alluvial fans except where these have been prevented from developing, or have been removed by destructive processes. Many (most?) of the fans show abrupt changes in sediments or gradients that can be related to abnormal rainfall or, very often, episodic uplift of the mountainous source areas. In some instances, suggestive features of the fans correlate with paired stream terraces upstream, as observed along much of the east side of the Diablo Range, supporting the likelihood of a tectonic origin (Bull, 1964). The latter is confirmed in cases where paired stream terraces project “into thin air” above an alluviated piedmont plain; this can be a criterion for uplift of a range along a marginal fault.

These geomorphic observations indicate that some (most?) ranges have risen, in some cases as blocks with concomitant slip on range-front faults, but in other cases perhaps by ductile squeezing up or variable mechanisms that are difficult to categorize. The morphology of some linear foothills suggests ductile arching, in some cases accompanied by faulting at the margins.

Figure 2. View looking south several ranges, from southwest flank of Diablo Range, showing their alignment and even crests. Most distant is the Santa Lucia Range (80 km away); just beyond fog are the Santa Cruz Mountains (32 km away); this side of fog are very recently uplifted foothills, an outlier of the Diablo Range (16 km away). The active Calaveras fault is along valley on near side of foothills. Fog occupies Santa Clara (structural) Valley.

Figure 3. Upland surface remnants and concordant ridges collectively representing paleotopography of low relief, probably formed near sea level. View is to northeast across the Diablo Range north of Mount Hamilton and east of San Jose. The Franciscan Complex, including melanges, underlies most of the area. Bare ridge in middle distance is approximately 10 km long.

ROCKS COMPOSING AND BORDERING THE RANGES

Overall Characteristics

Figure 5 is a simplified geologic map of the Coast Ranges. Many, but not all, of the formations and assemblages in the Coast Ranges are mechan-

Figure 4. Steep, linear northeast front of the northern Santa Lucia Range, surmounting large alluvial fans that descend toward the Salinas River. These features suggest fault-block uplift of this part of the range. Relief on mountain front is approximately 900 m.

ically weak and relatively nonresistant to erosion and mass-wasting. There is an abundance of poorly consolidated sandstone and shale, and an abundance of rocks that are rendered weak by pervasive fracturing. These generalizations apply to the abundant Franciscan (Mesozoic) rocks as well as the Neogene and Quaternary sedimentary rocks. The fact that such materials compose topographically high domains implies uplift rates that can successfully compete with erosional surficial processes.

Mesozoic Complexes

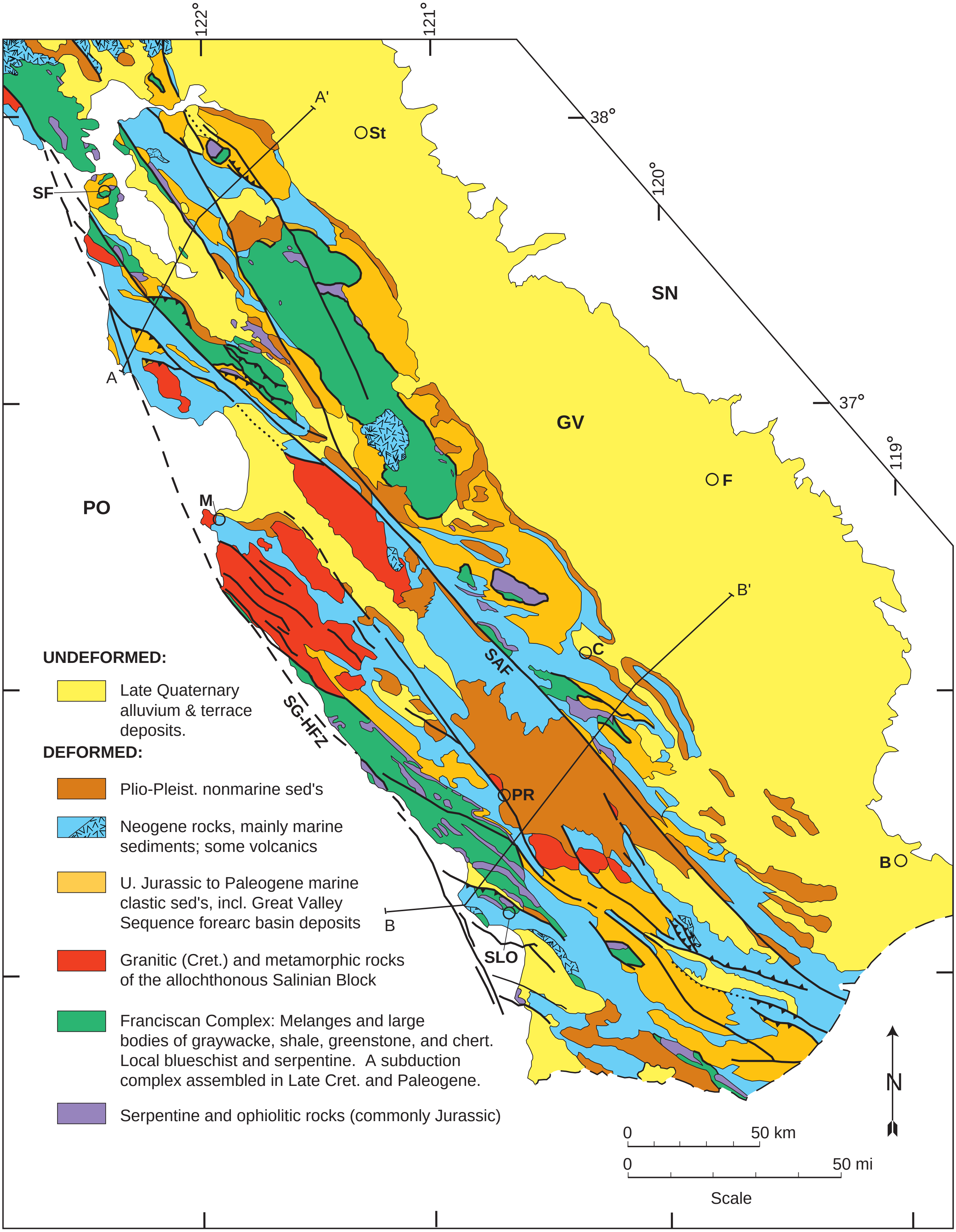
The underlying rocks of the subprovince include four categories of Mesozoic complexes: (1) the Franciscan Complex, (2) the Salinian block, (3) the Great Valley sedimentary sequence, and (4) serpentine and ophiolites, including the Coast Range ophiolite. These complexes partially overlap in age, but they differ fundamentally and were formed in different ways. Moreover, most of them are multiple; for example, the rocks that are generally called Franciscan include diverse subcomplexes, which have little in common except tectonic association.

Franciscan Complex. The complex is generally regarded as an accreted subduction zone complex, some of which is no longer in its original latitudinal location. It contains oceanic mantle remnants (serpentine), oceanic crustal components (gabbro, diabase, basaltic greenstone, radiolarian chert), and terrigenous sedimentary rocks (graywacke, siltstone, and shale), mainly of Late Jurassic and Cretaceous age (Bailey et al., 1964). It also contains blueschist facies equivalents of the foregoing, metamorphosed under high-pressure (*P*) low-temperature (*T*) conditions, presumably as a result of Mesozoic subduction. The complex includes melanges consisting of strongly sheared argillaceous matrix material in which blocks of the aforementioned rock types are more or less chaotically disposed, and large coherent units of graywacke and metagraywacke, some of which are described as "broken formations." The melanges are particularly nonresistant to tectonic and erosional processes. We think that the large volume and relatively ductile behavior of much of the Franciscan were important factors in the tectonics of the central and southern Coast Ranges.

Salinian Block. In contrast to the Franciscan Complex, the Salinian block is an allochthonous composite of granitic and metamorphic rocks from the axial portion of the western Cordilleran plutonic belt, which it resembles petrologically. It doubtless came from a region south (perhaps far south) of the Sierra Nevada. Hall (1991) gave a recent interpretation of the source and emplacement of the allochthon, with comprehensive references. The Salinian rocks include abundant granitic plutons (e.g., Ross, 1978) that are mainly Cretaceous and are coeval, but totally incompatible, with parts of the Franciscan. Also included are wall rocks of the Sur Series—largely sedimentary rocks metamorphosed under high temperatures. The Salinian rocks are inherently much more resistant mechanically than the Franciscan Complex. Locally they are not severely deformed, but elsewhere they have been pervasively sheared or crushed. In some localities, the granitic rocks are topographically prominent, but elsewhere their influence on the morphology of the ranges is only moderate to slight.

Serpentine and Ophiolites. Serpentine bodies derived from peridotite are distributed throughout Franciscan melanges. Serpentine is also locally present at faults (but not the major strike-slip faults) in or near the Franciscan Complex. Spectacularly, serpentine largely composes diapirs such as the New Idria mass (Coleman, 1996). The New Idria body has risen through the Franciscan Complex and entrains fragments of the latter. It and some other lesser masses must have come from within or below the Franciscan and may represent a mantle component of oceanic lithosphere, which was thrust under the Franciscan or incorporated within it during subduction. Serpentinization and mobilization were probably promoted by the transfer and pressurization of water during subduction and during the later transverse compression in the Coast Ranges (Coleman, 1996; see especially his Fig. 4 for a tectonic interpretation). The better understood Coast Range ophiolite (Hopson et al., 1981) is the oceanic floor upon which the Great Valley sequence sediments were deposited. The exposed remnants of the Coast Range ophiolite, including serpentine, have probably been peeled up along the northeast side of the Coast Ranges, bearing superincumbent Great Valley sequence sedimentary rocks, from an in situ ophiolitic basement beneath the Great Valley (sensu stricto; Griscom, 1982; Jachens et al., 1995). Most Coast Range ophiolite occurrences only compose the upper parts of an ophiolite sequence, commonly pillow basalt, but serpentinized ultramafic cumulates and mantle peridotite, as well as other ophiolitic members, are locally preserved. Wherever the Coast Range ophiolite, other ophiolites, or serpentine bodies appear at the surface, large tectonic displacements can be inferred. Figure 5 shows principal occurrences of such materials. It is manifest from the numerous, scattered exposures of these rocks, derived from various deep sources, that the crust has been profoundly deformed to a considerable depth.

Great Valley Sequence. Marine clastic sedimentary rocks (Bailey et al., 1964) are quasicoeval with both the Franciscan Complex and the Salinian plutons, having been deposited in a forearc basin (Dickinson and Rich, 1972) during the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UNDEFORMED:
Late Quaternary alluvium & terrace deposits.
DEFORMED:
Plio-Pleist. nonmarine sed's
Neogene rocks, mainly marine sediments; some volcanics
U. Jurassic to Paleogene marine clastic sed's, incl. Great Valley Sequence forearc basin deposits
Granitic (Cret.) and metamorphic rocks of the allochthonous Salinian Block
Franciscan Complex: Melanges and large bodies of graywacke, shale, greenstone, and chert. Local blueschist and serpentine. A subduction complex assembled in Late Cret. and Paleogene.
Serpentine and ophiolitic rocks (commonly Jurassic)
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same plate convergence cycle that formed the just-mentioned complexes. (However, these three complexes did not necessarily form at the same latitude.) The Great Valley sequence rocks are abundant, occurring as tectonically emplaced sheets, locally with remnants of the Coast Range ophiolite at the base, tectonically overlying or against the Franciscan Complex. The intervening fault is the regional Coast Range fault (Bailey et al., 1970) or one of many younger successor faults. The most striking occurrence of the Great Valley sequence is a thick homocline (Fig. 6) as wide as 15 km that flanks the Diablo Range and northern ranges and flattens downward to the east to blend in the subsurface with little-disturbed strata of the Great Valley. This homocline is the hanging wall of a tectonic wedge (Wentworth et al., 1984) that played an important role in the construction of the eastern ranges, as discussed later.

Tertiary Marine Sedimentary Rocks

Tertiary marine sedimentary rocks are characteristic of large areas in the Coast Ranges (Fig. 5). Some formations are shallow-water clastic shelf deposits; others (especially those of Miocene age) are deep-water shelf or slope-basin sediments rich in organic matter. The tectonic environment is significant in the evolution of the California margin (e.g., Blake et al., 1978; Crouch et al., 1984), and the widespread occurrence of marine middle and upper Miocene deposits in the coastal mountains places a limit on the time of inception of the Coast Ranges. A convenient condensed overview of Coast Range Cenozoic stratigraphy was provided by Lindberg (1984).

Cenozoic Volcanics

Volcanic rocks are locally present in the stratigraphy of the central and southern Coast Ranges and are of considerable tectonic interest (e.g., Dickinson and Snyder, 1979). Many are submarine extrusives of basaltic composition indicative of extension. However, most are Miocene in age and thus have little or no direct bearing on the Pliocene and Quaternary tectonic events that concern us.

Pliocene and Pleistocene Sedimentary Rocks: Generally Nonmarine, Commonly Deformed

As emphasized by Christensen (1965), the most significant rocks (or deposits) for understanding the youthfulness of the present Coast Ranges uplift are Pliocene–Pleistocene gravels, sands, silts, and clays, which are largely nonmarine and are in the proximity of the ranges in many localities (Fig. 5). These deposits are mainly of fluviatile origin, but some are debris flows, some are lacustrine, and some are tongues of shallow-marine or estuarine sediment. It is noteworthy that the basal and near-basal strata of at least three of the mainly nonmarine formations (Paso Robles, Tulare, and Santa Clara Formations) are shallow-marine or estuarine facies that intertongue with, or are overlain by, dominant fluviatile sediments. Therefore, the basal beds represent a paleo-sea level. The predominantly nonmarine character of the bulk of these Pliocene–Pleistocene formations denotes the “final” emergence of coastal California. The coarser facies contain clasts that indicate source areas, rock assemblages, and erosional levels that were providing sedimentary debris. In addition, some coarse facies provide current directions and other paleogeographic clues. In a few instances, datable tephra layers are interbedded with clastic strata.

It is important to note that Quaternary deposits older than ca. 450 ka are commonly, albeit locally, folded and faulted. The Pliocene–Pleistocene sedimentary rocks contain the history of widespread uplifts, provide evidence of subsequent recent deformation, and constrain the timing of the more recent rise of the ranges in their present form.

Among the Pliocene–Pleistocene nonmarine formations that fit the above generalizations are the following: Santa Clara Formation, Irvington Gravels, Livermore Gravels, Packwood Gravels, San Benito Gravels, Quatal Formation, Morales Formation, Tulare Formation, and Paso Robles Formation. Some salient facts regarding these formations are briefly summarized in Table 1. There is probably considerable overlap in terminology and considerable merging and continuity among sedimentary rocks that have locally been given different names. Several of the Coast Ranges of central California are partially flanked by such deposits (Fig. 5); in many cases, discordantly. The deposits are also generally flanked by younger, undeformed (or scarcely deformed) fan deposits. The Santa Clara Formation was interpreted by Vanderhurst et al. (1982), the Paso Robles Formation by Galehouse (1967), the San Benito Gravels by Griffin (1967), and the Tulare Formation by Woodring et al. (1940). Other authors are cited in Table 1.

The ages of the deformed nonmarine Pliocene–Pleistocene deposits are critically important for the dating of range uplifts. The available age determinations, which generally are between 4 Ma and 0.47 Ma, may be grouped in the following categories: (1) ages obtained by correlation of tephra layers (in sedimentary rocks of numerous localities) with isotopically dated ash elsewhere; (2) radiometric dates from lavas enclosed in the sedimentary rocks, obtained in only one area; and (3) paleontologic ages. A summary of some selected age data is given in Table 1. We rely most heavily on ages in category 2, provided by the important work of Sarna-Wojcicki and associates (e.g., Sarna-Wojcicki, 1976; Sarna-Wojcicki et al., 1985, 1991). Correlations of tephra with the Rockland ash are particularly important, this ash having been dated at about 0.40–0.47 Ma (Meyer et al., 1991; Sarna-Wojcicki, 1996, personal commun.). For expedience, we have arbitrarily picked 0.45 Ma for the age of the Rockland ash and the upper part of the Santa Clara Formation, which contains the ash, but we have no scientific reason for preferring 0.45 Ma over any other age between 0.40 and 0.47 Ma.

INTERNAL STRUCTURES—LARGELY CONTRACTIONAL

The contractional tectonics have also been emphasized by others, including Aydin and Page (1984), Namson and Davis (1988), and Jones et al. (1994).

Folds Within the Ranges

All of the mountains except the Gabilan Range have numerous pronounced folds. Many

Figure 5. Simplified geologic map of the central and southern Coast Ranges. Note localization of the ranges along the broad transform plate boundary, characterized by large strike-slip faults. Note also the indications of contraction transverse to the plate boundary. Numerous fragments of serpentine and ophiolite (both in purple) imply widespread deep deformation bringing pieces of mantle and oceanic crust to the surface. The recent uplift of the ranges is implied by the relations of deformed Pliocene–Pleistocene formations (brown). Locations of crustal transects A–A′ and B–B′ are shown. Abbreviations: B—Bakersfield; C—Coalinga; F—Fresno; GV—Great Valley; M—Monterey; PO—Pacific Ocean; PR—Paso Robles; SAF—San Andreas fault; SF—San Francisco; SG–HFZ—San Gregorio–Hosgri fault zone; SLO—San Luis Obispo; SN—Sierra Nevada; St—Stockton.

trend parallel (or nearly so) with the San Andreas fault and the plate boundary, showing a genetic relationship to inferred transverse compression. However, some folds are oblique with respect to the San Andreas fault (Fig. 7), possibly reflecting the prevalent horizontal dextral shear of the transform regime. A few oblique folds such as the Vallecitos syncline in the Diablo Range (Fig. 5) have been evolving since they were first established as elongate warps, perhaps as early as Eocene time, presumably by dextral transpression during oblique plate convergence. A number of folds began to form in late (or even early) Miocene time. Regardless of trend and the time of inception, the Cenozoic folds commonly involve Neogene rocks, and, in many cases, Pliocene marine strata and/or Pliocene–Pleistocene nonmarine formations.

Although the Coast Ranges folds have different histories, apparently nearly all grew (or tightened) in late Quaternary time. Remarkably, most of these young folds are incorporated within the ranges as internal structures rather than defining the ranges. In other words, an anticline in Neogene rocks usually does not coincide with an antiformal mountain range. This means that the ranges are even younger than most of the Pliocene–Pleistocene folds. However, a few outlying anticlines and synclines that might be considered to be satellites of the Diablo and Temblor Ranges are so young that they have direct topographic expression (e.g., Coalinga anticline and Kettleman Hills, both of which are parallel with the San Andreas fault). Thus, there has been a succession of folding events, some of which preceded the uplifts of the present ranges and some of which have been contemporaneous with the uplifts.

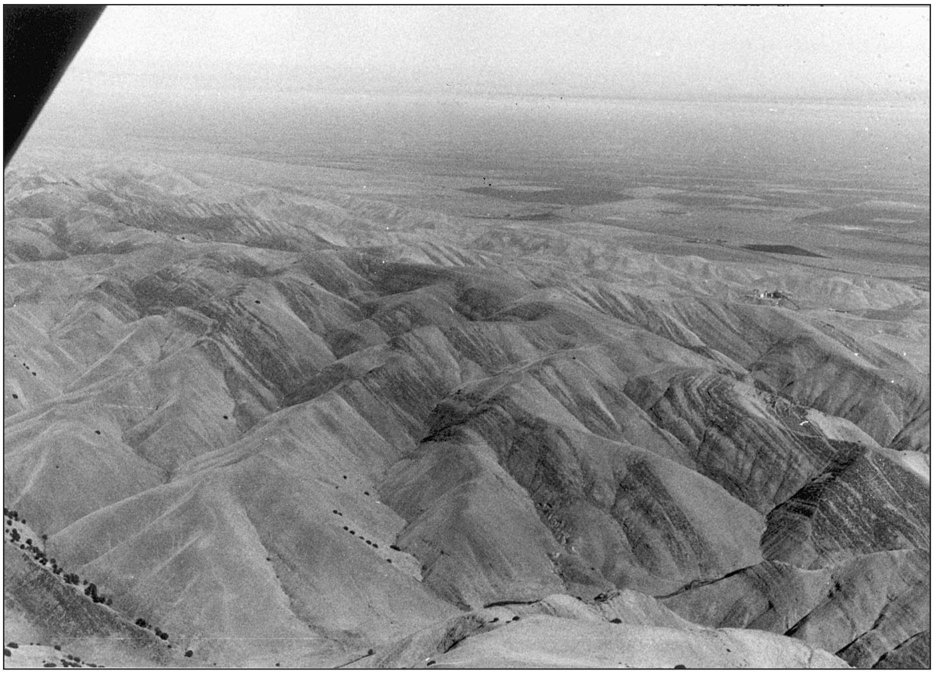

Figure 6. Homocline of Great Valley sequence (clastic Mesozoic forearc basin sedimentary rocks) dipping easterly along northeast flank of Diablo Range. Approximately 3 km across the tilted section.

Internal Faults

Several kinds of faults are within the ranges. Like the folds, most are young, as shown by offsets of upper Miocene or Pliocene–Pleistocene formations, but few have direct topographic expression. There are low-angle detachments of unknown origin, and there are low-angle thrusts (e.g., Fig. 8) and high-angle reverse faults. Some of the thrusts and reverse faults trend west-northwest or east-west and can be explained by wrench tectonics, like the oblique folds. However, many internal thrusts and reverse faults are subparallel with the San Andreas fault and are best explained as products of transverse compression.

Small, young reverse faults that do not seem to play any part in uplift are on the east flank of the

TABLE 1. PLIOCENE–PLEISTOCENE FORMATIONS INVOLVED IN COAST RANGES OROGENY

Name	Character	Age	Associated range
Santa Clara Formation	Mainly fluviatile clastic rocks. Marine tongues near base. A few lacustrine and rare thin ash beds.	Probably ca. 3.6–0.45 Ma. Lower beds are Pliocene, based on mollusks near Stanford (Addicott, 1969) and K-Ar age (3.6 Ma of interbedded basalt near Gilroy (Sarna-Wojcicki, 1976). Upper part has 0.47–0.40 Ma tephra on the basis of correlation with Rockland ash (Sarna-Wojciki, 1976).	Santa Cruz Mountains, Diablo Range
Irvington Gravels (probably = upper Livermore Gravels)	Fluviatile sand, gravel, silt, and clay.	Probably ca. 1.9–0.45 Ma (Lindberg, 1984); derived age of notable vertebrate fauna, basis of Irvington Stage.	Diablo Range
Tulare Formation	Mainly fluviatile clastic rocks, some lake beds, estuarine at base.	Probably ca. 4.0 Ma to late Pleistocene. Pliocene mollusks near base (Addicot and Galehouse,1973). Ash in lower part is ca. 4.0 Ma (Sarna-Wojcicki et al., 1991). Upper part evidently not well dated.	Temblor Range Diablo Range
Paso Robles Formation	Mainly fluviatile clastic rocks; marine tongues at base.	Probably ca. 4.0 Ma to late Pleistocene. Pliocene mollusks near base (Addicot and Galehouse,1973). Ash in lower part is ca. 4.0 Ma (Sarna-Wojcicki et al., 1991). Upper part evidently not well dated.	Gabilan Range, Santa Lucia Range
Livermore Gravels	Mainly fluviatile, sand, gravel, silt, clay, some lake beds.	Pliocene–Pleistocene, based on vertebrates and freshwater mollusks (Griffin, 1967).	Diablo Range
San Benito Gravels	Mainly fluviatile, sand, gravel, silt, clay, some lake beds.	Pliocene–Pleistocene, based on vertebrates and freshwater mollusks (Griffin, 1967).	Diablo Range

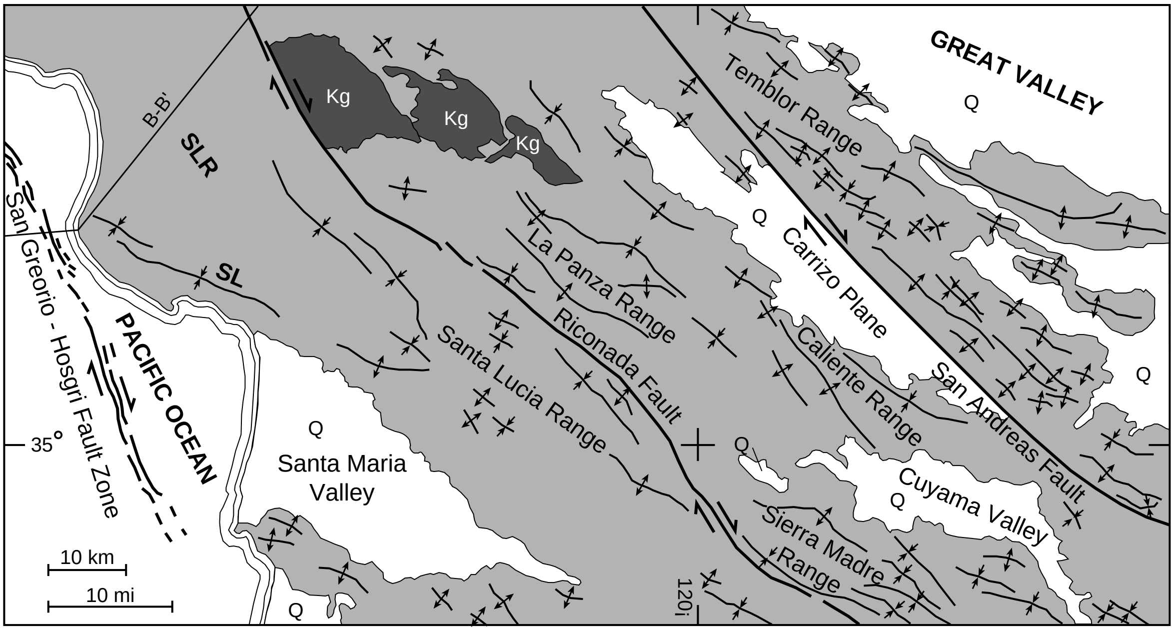

Figure 7. Map of fold axes and major strike-slip faults in part of the southern Coast Ranges at the south end of Figure 5. The strike-slip faults approximate the trend of the plate boundary, and most of the fold axes are nearly parallel with, or at an acute angle to the boundary. Abbreviations: B–B′—line of cross section (Fig. 5); Kg—Cretaceous granitic rocks of the Salinian Block; Q—Quaternary alluvium.

Diablo Range between San Luis Reservoir and Little Panoche Creek (Lettis, 1982). These strike parallel with the range and appear to be bedding-plane faults in the Upper Cretaceous Panoche Formation, which here generally dips northeast, away from the crest of the range. The sense of slip on these faults is reverse, the northeast side moving up relative to the opposite side. Thus, the slip is opposite to that which would contribute to the rise of the axial part of the Diablo Range. Holocene movement has not been demonstrated, but the slip (5 to 100 m) is so recent that Pleistocene pediment surfaces are offset, remnants of scarps survive, and drainage seems to be impeded in some places (Lettis, 1982). Although these rather puzzling faults do not play a critical part in the range uplift, perhaps they are "sympathetic" to a northeast-dipping thrust fault underlying the hanging wall of a Wentworth-type wedge (see section on Crustal Structure—The Middle Crust).

SUMMARY OF CENOZOIC TECTONIC EVENTS

Figure 8 is a structural section across the Santa Cruz Mountains, and shows some ages (denoted by numerals) of certain unconformities, folds, and faults. It is immediately apparent that more than a single tectonic episode has transpired, and that the greater part of the deformation occurred in Cenozoic time, and much of it in Quaternary time.

Table 2 summarizes the main events recorded in the cross section of Figure 8. The late Miocene folding and faulting are so pronounced that one school of thought maintains that the Coast Ranges are largely Miocene, and extremists have suggested that their beginning was in Eocene time. We think that the Pliocene and Quaternary disturbances were genetically related to the rise of the ranges with their present outlines, although at least one Miocene episode (described below) in the area of the Diablo Range may have had some long-term influence. Unconformities are numerous in the Tertiary records of many of the Coast Ranges, as shown in various stratigraphic columns (e.g., Lindberg, 1984), and most of these apparently reflect tectonic warping. They do not correlate spatially with particular ranges, and in most cases their varying chronology from place to place does not support the notion of far-reaching mountain-building episodes. We briefly summarize the tectonic and stratigraphic-sedimentologic record, beginning with the latest Cretaceous.

Franciscan Debris in Upper Cretaceous and Eocene Sedimentary Rocks

The Maastrichtian (uppermost Cretaceous) Moreno Formation is the oldest sedimentary rock unit known to contain Franciscan detritus (S. A. Graham, 1996, personal commun.). It is an upper part of the marine Great Valley sequence and was deposited during subduction, probably receiving Franciscan debris from an elevated outer arc ridge, which was not directly related to the eventual Coast Ranges. The next appearance of Franciscan debris in the stratigraphic record is in Eocene sedimentary rocks. The shallow-marine Domengine Formation (ca. 50 Ma) of the central Diablo Range typically overlies unconformably older Paleogene sedimentary rocks and contains Franciscan detritus in minor quantities. A marked disturbance must have elevated a source area, either far beyond (west of) the accumulated Mesozoic Great Valley forearc basin deposits, or possibly entailing the stripping of these deposits to expose underlying Franciscan rocks. We think that this event took place during subduction, that it represents an unusual uplift of the outer arc ridge, and that it is not relevant to the origin of the Coast Ranges.

Late Oligocene Disturbances, 29–25 Ma

Oligocene rocks in the central Coast Ranges include some nonmarine sediments, and Oligocene deposits are missing in many areas where other Tertiary epochs are well represented. These circumstances are probably related to the change from plate convergence to transform interaction at about 29–25 Ma in areas that were then at the

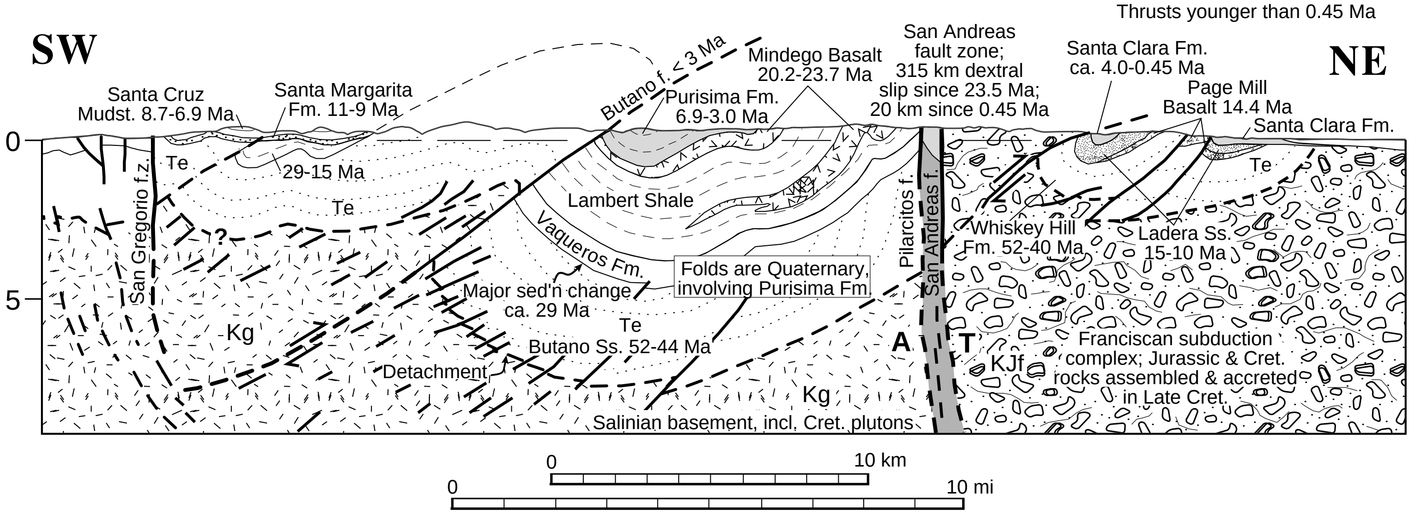

Figure 8. Cross section of Santa Cruz Mountains, and dated formations and structures. This is modified southwest part of line A–A′ (Figs. 5 and 10). The section shows dominance of contractional features and youthfulness of deformation. Folds and faults are young (largely Quaternary), but were deeply truncated by erosion prior to uplift of present ranges, commencing ca. 400 ka. Abbreviations: KJf—Franciscan Complex; Kg—Cretaceous granitic rocks of Salinian block; Te—Eocene sedimentary rocks—mainly turbiditic.

present latitude of southern California. Subsequent prolonged marine sedimentation occurred, so the 29–25 Ma disturbances did not lead directly to the building of the Coast Ranges.

Early and Middle Miocene Events

Oligocene and older rocks are commonly overlapped unconformably by shallow-marine “Vaqueros-type” sandstones and conglomerates. In areas of the present continental shelf, early Miocene sedimentary basins formed during extensional episodes. Most of these basins and their contents were compressed later in the same epoch, but their deformation was not as acute as the later folding and thrusting throughout the site of the Coast Ranges.

Lower and middle Miocene shallow-water marine sandstones that unconformably overlie older rocks in west-central California commonly contain Franciscan detritus as a noticeable minor constituent. This indicates uplift and exposure of the Franciscan marine accretionary complex after the cessation of subduction. Marine deposition was widespread thereafter, and the distribution and character of marine sedimentary rocks precludes the existence of extensive mountains prior to late Miocene time. Nevertheless, local crustal deformation is indicated, especially where marine Miocene rocks rest directly on Mesozoic complexes. This is observed in some areas underlain by Salinian plutonic rocks (e.g., near Carmel and Monterey) as well as in areas of the Franciscan Complex (e.g., in parts of the San Luis Obispo–Atascadero region of the Santa Lucia Range).

A possible Coast Range precursor uplift occurred within the present confines of the Diablo Range south of Pacheco Pass. There, a thin remnant of middle Miocene marine sandstone unconformably overlies the Franciscan core of the range and underlies a patch of Quien Sabe Volcanics at lat 36°50′N, long 121°20′W (Fig. 5). The lowest Quien Sabe flows are dated at 11.6 Ma (as discussed by McLaughlin et al., 1996). Thus, that particular area was elevated above sea level in the Miocene, was resubmerged beneath the sea before 11.6 Ma, and was subsequently uplifted again within a central part of the modern Diablo Range. The initial uplift was restricted to the site of the present range, as shown by the fact that the Franciscan basement rocks were not generally exposed (hence not markedly uplifted) elsewhere. Their cover of Great Valley rocks, which we think had been tectonically emplaced earlier, was not eroded away and is still preserved in extensive remnants around the periphery of the mountains (Fig. 5). The spatial coincidence between the Miocene and Quaternary uplifts suggests that the former was localized by a deep-seated condition that persisted and influenced the later event. However, the trend of the Miocene emergent terrain was somewhat oblique to the axis of the present mountains, as the uplift apparently did not affect the concurrent marine conditions in the adjoining Vallecitos syncline area to the south (Fig. 5). Moreover, there was not a continuous uplift that simply persisted after the early Miocene event, as marine conditions returned in areas around, within, and across the site of the present mountains. Various parts of the Diablo Range incorporate marine sedimentary rocks that were deposited in late Miocene time and were severely folded and faulted prior to the birth of the present range.

Late Miocene Tectonism, 11–7 Ma

Marked folding and some faulting occurred offshore from the present coast late in Miocene time, and more severe folding and faulting occurred in some areas now on land in the present Coast Ranges. An example may be seen near the southwest end of the area shown in Figure 8, where the mildly folded Santa Margarita Formation (uppermost Miocene) unconformably overlies strongly folded Oligocene strata. In areas northeast of the San Andreas fault, late Miocene folding may have been an effect of the passage of the Mendocino Triple Junction, because it, where currently situated near Cape Mendocino, is accompanied by active deformation. The late Miocene disturbance(s) generally did not coincide spatially with individual modern Coast Ranges, and where data are available, the folding was followed by resumed marine deposition.

Major Tectonism Commencing ca. 3.5 Ma

Figure 5 shows the areal distribution of preserved Pliocene–Pleistocene nonmarine sedimentary rocks such as the Santa Clara, Tulare, and Paso Robles Formations. These rocks record the various uplifts that collectively raised western California above sea level, where it remains today, and that exposed Cenozoic and Mesozoic marine

TABLE 2. MAIN CENOZOIC TECTONIC EVENTS

Time (Ma) (approx.)	Event	Comments/questions
0.40	Delineation, and beginning of uplift, of present-day central and southern Coast Ranges	Why was there a delay of about 50 000 yr between severe contractional deformation and uplift of ranges?
0.45–0.40	Widespread, deep erosion, drastically modifying folds	Enormous scale of erosion facilitated by predominance of nonresistant rocks and pluvial, glacial-stage climate
0.45	Approximate beginning of Coast Ranges folding and thrusting, affecting Pliocene–Pleistocene formations	Concurrent erosion
3.5	"Permanent" uplift of California above sea level, initiating Coast Ranges mountain building	Ascribed to change in plate motions (e.g., Harbert and Cox, 1989)
11–7	Folding and thrusting in parts of present Coast Ranges province and in areas now offshore	Not directly related to present Coast Ranges, except by thickening of the crust
29–25	Major change in character of sedimentation; extensive unconformities	Change from subduction to transform regime in domains formerly farther south

formations to erosion. In most Coast Range areas where Pliocene–Pleistocene formations are preserved, these sedimentary rocks are locally folded and faulted. Thus, there was probably a continuum of tectonic activity, beginning with uplift and erosion that fed the Pliocene–Pleistocene nonmarine deposits, and evolving to a stage of pronounced folding and thrusting involving these same deposits. In many cases, the underlying formations are more severely deformed, having undergone Miocene or older deformation as well as this recent episode, but in other cases the Pliocene–Pleistocene formations are folded almost as much as Oligocene and Miocene strata beneath them. Collectively, the Pliocene–Pleistocene tectonic events in this region were probably the most severe of any since the Mesozoic. As discussed later herein, we think that this orogenic sequence was the result of the change in plate motions at about 3.5 Ma.

The deformation cited above did not directly produce the existing ranges, but evidently "prepared the ground." The individual ranges of the present day do not coincide spatially with any particular deformational features of the Pliocene–Pleistocene folding and thrusting, but the entire central Coast Ranges subprovince is restricted to the realm of that deformation. The northeastern boundary of the Coast Ranges tends to coincide with the northeastern limit of Pliocene–Pleistocene folding, thrusting, and wedging. To the southwest, Pliocene–Pleistocene folds are observed beyond the ranges far offshore on the continental shelf, but the intensity of deformation is relatively mild. Landward, the shoreline and the edge of the mountain belt are located roughly where the Pliocene–Pleistocene deformation becomes intense.

Although most of west-central California did not rise above sea level permanently until the Pliocene–Pleistocene orogeny, the east San Francisco Bay area had emerged earlier. The site of the present bay was a topographic high in late Miocene time, shedding sediment eastward (Graham et al., 1984). This points up the youthfulness of the western California landscape.

Rise of the Present Ranges Since ca. 0.4 Ma

The appreciable separation in time and space between the Pliocene–Pleistocene folding and the rise of the present Coast Ranges in late Pleistocene and Holocene time is shown by the fact that the Pliocene–Pleistocene structures were extensively and deeply eroded prior to the uplift of gently rolling topographic surfaces, which truncate the folds. Moreover, the margins of some ranges transect folded Pliocene–Pleistocene formations, some of which contain beds as young as 400 000–470 000 yr (Sarna-Wojcicki, 1976).

GEOPHYSICAL DATA AND CRUSTAL STRUCTURE

Two Crustal Transects

In order to understand the character of the Coast Ranges province at depth as well as at the surface, we have compiled two composite crustal transects, one passing near San Francisco (Fig. 9, inset), and one passing near San Luis Obispo (Fig. 10, inset). The compilations draw heavily from the work of many geologists and geophysicists, as noted (incompletely) on the two figures; we are deeply indebted to all. The two transects show a few significant similarities, despite the differences in rocks and structures. A crustal-scale perspective requires consideration of geophysical data, which are discussed below.

Effects of Transform Movements on Geology of the Transects. In viewing Figures 9 and 10, the reader must be aware that the southwestern parts of the crust have moved (and continue to move) relatively away from the observer, and the northeastern parts have moved relatively toward the observer. Thus, each transect is an instantaneous frame in a moving picture, and the various rocks and structures did not necessarily originate in their present side-by-side positions. The transform motion began about 29 Ma (Atwater, 1970), is now proceeding at a rate of about 4.8 cm/yr (DeMets et al., 1990), and has probably accrued at least 920 km of strike separation. This relative motion between the Pacific and North American plates is distributed in a broad zone, within which the San Andreas fault accommodates about 3.5 cm/yr (Prescott et al., 1981). Thus any rocks or structures of Oligocene age or older, if located near the southwest end of a transect, may have been displaced hundreds of kilometers with respect to rocks of comparable age near the northeast end of the same transect. However, the younger the rocks and structures, the smaller the total relative displacement. Inasmuch as the central and southern Coast Ranges are very young, the transform displacements of relevant age must be relatively moderate, although still substantial. Pliocene rocks on the two sides of the San Andreas fault are probably displaced less than 175 km, and Pleistocene formations are displaced less than 55 km.

An acute problem is posed by the ongoing transform motions vis-a-vis the presence of the seemingly little-disturbed lower crust. This problem is considered in a later section.

Interpretation of Gravity Profiles

Bouguer and isostatic residual anomalies are plotted on the two crustal transects, Figures 9 and 10 (foldouts). Isostatic residual anomalies are obtained by subtracting from Bouguer gravity the calculated effect of idealized isostatic compensation for regional elevation (Roberts et al., 1990; Roberts and Jachens, 1993). Because the calculated effect is regional, the result is a high-pass filter that removes the long-wavelength, crustal scale gravity anomalies (wavelength greater than roughly 50–100 km) while preserving anomalies having a shallower source and smaller lateral extent, such as those caused by sedimentary basins and lithologic variations in the basement.

In section A–A′ (Fig. 9; see Fig. 5 for location), the Bouguer anomaly is strongly positive (about 25 mGal) at the southwestern end, because the continental crust is unusually thin near the coast. On the other hand, the isostatic residual anomaly is compensated for this effect. East of the San Andreas fault, the two anomalies track close to one another (within 10 mGal), and both are strongly negative in the northeastern half of the transect. As explained more fully in the following discussion of section B–B′, the negative anomalies are closely associated with thick, low-density Cenozoic sedimentary rocks. This correlation is clearly demonstrated in the vicinity of the Calaveras fault and in the Great Valley; there is an intervening gravity high over the more dense Cretaceous sedimentary rocks in the northern Diablo Range. On a more detailed scale, Cenozoic deposits preserved in faulted synclines generally show associated negative anomalies. The serpentine bodies, which produce strong magnetic anomalies, have little or no gravity signature at this scale.

In section B–B′ (Fig. 10) the Bouguer and isostatic residual anomalies are virtually coincident and near zero about 15 km landward from the coastline. West of this area of coincidence, the Bouguer anomaly becomes positive (about 10 mGal at the coast) because of the isostatic effect of the thin continental crust, while the isostatic residual anomaly remains near zero. Northeastward from the coastline, the Bouguer anomaly gradually becomes more negative toward the interior of the continent, and at the northeastern end of the transect, it differs from the isostatic residual anomaly by about –25 mGal. Both anomalies clearly show a correlation with structural and/or lithologic features in the transect and on the geologic map (Fig. 5), especially with low-density sedimentary rocks. The isostatic residual anomalies are used here to interpret these features.

Low-density sedimentary rocks, mainly of Cenozoic age but including Cretaceous rocks in the Great Valley, have a strong negative effect on the gravity anomaly. Because these rocks have been intensively drilled for hydrocarbons, their thickness, density, and gravity effect can be roughly estimated. The long-dashed line in section B–B′ (labeled sediment-adjusted gravity) shows the result of removing the effect of these low-density sedimentary rocks; the gravity curve becomes almost featureless. The adjusted gravity curve on this section shows surprisingly little contrast between Franciscan rocks (Mesozoic subduction complex) and Salinian rocks (plutonic arc terrane), even where these are juxtaposed along the San Andreas fault. Thus, the two complexes do not differ substantially in mean density. At a smaller scale, some low-amplitude anomalies can be correlated with local structure and Cenozoic sedimentary rocks preserved in synclines, but the serpentine bodies, which produce the strongest magnetic anomalies, have no gravity signature at the scale of the cross section.

Magnetic Anomalies

As shown on both crustal transects (Figs. 9 and 10), the magnetic field varies by several hundred nanoteslas (nt) and is highest in the Great Valley (well beyond the Coast Ranges), where a well-known magnetic high anomaly trends subparallel with the axis of the valley.

Magnetic Rocks. Generally, rock magnetization depends upon magnetite content, grain size, and other characteristics of the magnetite, and the magnetization history (e.g., strength and polarity of Earth's field at the time of magnetization). On the regional scale of interest here, remanent magnetization (often dominant in quickly cooled lavas) can probably be neglected in favor of induced magnetization (parallel to the Earth's field), which is important in deep-seated igneous and metamorphic rocks. Ophiolitic rocks of the Coast Ranges include (1) peridotite, which is virtually nonmagnetic where unserpentinized (Thompson and Robinson, 1975), because the iron is accommodated in olivine; (2) serpentine, usually intensely magnetic because of abundant secondary magnetite; (3) altered basalt or "greenstone," which is usually weakly magnetic because of alteration; and (4) gabbro, diabase, and mafic amphibolite, which, if unaltered, tend to be strongly magnetic. Partial to complete serpentinization of peridotite is common and progressively increases the magnetization and decreases the density, so that the combined analysis of gravity and magnetic anomalies is meaningful.

Certain Coast Ranges sandstones containing abundant andesite grains (from Sierra Nevada volcanics) are an additional unusual source of magnetic anomalies (R. C. Jachens, 1996, personal commun.). The upper Miocene Neroly Formation is an example (see following). Doell (1956) found that the strong magnetism is remanent but was acquired subsequent to folding and is approximately parallel with the Earth's field. He suggested that the magnetism resides in a bluish iron-rich montmorillonitic coating on the sand grains, but this interpretation awaits further study.

Interpretation of Magnetic Profile in Transect of Figure 9. Beginning at about kilometer 20 from the southwest part of the transect, an anomaly of 50–75 nT west of the San Andreas fault is associated with the Mindego basalt of early Miocene age. These rocks include flows and shallow intrusions of fresh basalt. Northeast of the San Andreas fault a weak multipeaked anomaly of about 25 nT may be attributed to Franciscan greenstones, which are abundant at the surface. Farther northeast, a strong anomaly of amplitude about 150 nT near the edge of San Francisco Bay, and a similar anomaly on the northeast side of the bay, are most likely caused by serpentine in shear zones or melanges in the Franciscan Complex. Serpentine, probably related to the northeastern anomaly, is exposed at one end of Coyote Hills. At the Hayward fault zone, a sharply peaked anomaly of amplitude about 200 nT is ascribed to serpentine bodies like those that are extensively exposed farther northwest along the fault in association with slivers of various lithologic units of the Coast Range ophiolite.

East of the Calaveras fault zone and on strike with the Mount Diablo diapiric structure to the northwest is a broad magnetic high with superimposed peaks. The striking peaks are caused by volcanic sandstone of the Neroly Formation. Note that the gravity low associated with the broad magnetic high is caused by low-density Cenozoic sedimentary rocks more than 5 km thick. An exposure of Coast Range ophiolite in part of the diapiric core of Mount Diablo (13–18 km northwest of our transect) suggests that the high magnetic and gravity anomalies there stem from a substantial subsurface body of ophiolite that includes both magnetic serpentine and dense mafic and/or ultramafic rock, perhaps obducted into the body of the range, or perhaps lifted from the Great Valley ophiolitic basement by an eastward-advancing wedge of Franciscan material. This latter interpretation is adopted (with some misgivings) in Figure 9, in which we extend the Riggs Canyon fault (striking toward the Mount Diablo diapir) downward into the subsurface as the hanging wall of a Franciscan wedge.

From about kilometer 95, the magnetic field declines eastward toward the Great Valley, but begins to rise near the valley edge as part of the major Great Valley magnetic and gravity high associated with an ophiolitic basement. Unpublished maps of magnetic potential and high-pass-filtered magnetic potential, provided by R. C. Jachens, indicate a general westward transgression of the Great Valley magnetic anomaly into the Coast Ranges province, as interpreted by Jachens et al. (1995). Large masses of ophiolite and/or displaced ophiolitic serpentine are inferred to underlie the Diablo Range, on the basis of magnetic and gravity anomalies.

Interpretation of Magnetic Profile in Transect of Figure 10. Beginning in the southwest, most of the prominent anomalies in the transect can be associated with serpentine bodies, either exposed or concealed. Note that corresponding positive gravity anomalies are generally lacking, as expected for serpentine. The large (300 nt) sharp-peaked anomaly in the Santa Lucia Range is associated with the Cuesta Ridge ophiolite (Page, 1972), a fragment of the Coast Range ophiolite. The broader anomaly of similar amplitude near the western edge of the Salinian block has no obvious source. It has no coinciding gravity anomaly and may be caused by an unusually magnetic facies of quartz diorite or by metamorphic wall rocks. In the Diablo Range northeast of the San Andreas fault, a magnetic peak is on strike with the Table Mountain serpentine to the north. A smaller high at 132 km lies over upturned Cenozoic sedimentary strata and is inferred to be caused by "blue" volcanic sandstones (see Magnetic Rocks section). The huge Great Valley anomaly is attributed to ophiolitic Great Valley basement (Jachens et al., 1995), as previously discussed.

Anomalous Heat Flow and Elevated Crustal Temperatures

Heat flow in the central and southern Coast Ranges is anomalously high, averaging about 83 mW/m^2 (Lachenbruch and Sass, 1973), compared to the stable United States continental interior (about 42–63 mW/m^2; Lachenbruch and Sass, 1977). The high heat flow drops markedly from the Coast Ranges province to the adjacent Great Valley. On our crustal transects (Figs. 9 and 10), the values in the Coast Ranges vary from 63 to 100 mW/m^2, whereas, in the Great Valley, the values are 40 to 64 mW/m^2. Thus, there is likely a relationship between Coast Ranges tectonics and the elevated heat flow. However, we are uncertain which came first, the high heat flow or the intense deformation, which may have been one of the causes of the anomaly. In interpreting the heat flow, one needs to bear in mind that tectonic extension tends to increase observed heat flow by necking and convergence of isotherms, whereas tectonic contraction (typical of the Coast Ranges) has the opposite effect. Erosion tends to increase, and deposition to decrease, the observed heat flow. Advection by magma or other fluids, and frictional heat in deforming rocks, can disrupt simple conductive heat flow. Moreover, thermal changes in the lower crust or mantle take millions of years to be felt in the near surface where measurements are made. Nevertheless, thermal data supply important constraints for tectonic history.

As pointed out by A. Lachenbruch and C. Williams (1994, personal commun.), in the central and southern Coast Ranges, heat flow does not decrease southward as might be expected if the cause of the anomalous values were related to the northward passage of the Mendocino Triple Junction, that is, if it were caused by the upwelling of hot asthenosphere into a northward-advancing slab window such as that proposed by Dickinson and Snyder (1979). The high heat flow appears to extend offshore, but the marine measurements are somewhat suspect (A. Lachenbruch, 1995, personal commun.). If an anomalous flux exists offshore, there may be a spatial coincidence between it and the late Miocene folds and reverse faults that are observed on the continental shelf. These folds and faults, although of moderate intensity, are the most pronounced Cenozoic structures recognized offshore. Late Miocene folds and faults are noted in parts of the Coast Ranges, but there they are overshadowed by the effects of Pliocene–Pleistocene deformation.

Significantly, the San Andreas strike-slip fault in west-central California produces no observed local effects, consistent with low friction and little heat generation on this fault (Lachenbruch and Sass, 1973, 1980; J. H. Sass, 1995, personal commun.).

The high heat flow predicts high crustal temperatures. Assuming steady-state flow, Sass et al. (1995, personal commun.) calculated 700 °C or more at the base of the crust and about 400 °C at 15 km, the approximate base of the seismogenic zone (where ductile behavior of rocks becomes more pronounced as a function of depth and temperature) in the Parkfield area in the central Coast Ranges. Such temperatures require metamorphism of particular rocks at particular depths, and this must be taken into account in interpreting variations of seismic velocities at increasing depths. Williams et al. (1994) analyzed the causes of the Coast Ranges high heat flow and concluded that the source of the heat was most likely in the deep crust and/or uppermost mantle and that there is probably no simple cause of the observed anomaly, but rather, a combination of contributing causes such as asthenospheric upwelling, magmatism, and mechanical heating during deformation.

In the larger geologic perspective, Mesozoic and early Cenozoic subduction of an old oceanic plate is thought to have "refrigerated" the Great Valley and Sierra Nevada mantle and to have produced the low reduced heat flow (heat flow corrected for upper crust heat production) still observed there (Blackwell, 1971; Roy et al., 1972). The northwestward advance of the transform system, replacing subduction, would have spared the site of the central and southern Coast Ranges from this type of "refrigeration" progressively; i.e., after about 25 Ma in the south and after about 7 Ma at the site of San Francisco.

At the latitude of southern California, subduction of a platelet (the Monterey plate) of young, initially relatively warm oceanic crust occurred in Oligocene and Miocene time (Lonsdale, 1991). This subduction ceased at about 19 Ma, but apparently the platelet underlying the continental margin, but now attached to the Pacific plate, rode northwestward with the latter as dictated by the Pacific–North American transform plate interaction (Nicholson et al., 1994; Bohannon and Parsons, 1995). Part of the Coast Ranges system evolved in the overlying crust that rode piggyback on this relatively young platelet. The platelet may have served as a (waning) heat source or as a heat sink, depending on how deep it subsided. The remainder of the central Coast Ranges, although beyond the limits of the Monterey plate, is also probably underlain by oceanic crust (e.g., Fig. 9). The origin and age of this presumed oceanic material are unknown. Part (or all) may be a leftover slab of Farallon oceanic crust, in which case it would be too old (Mesozoic and/or Paleogene) to be a likely source of heat. It may be part of an unidentified counterpart of the warm Monterey plate. There is a possibility that the edge of the continent was thrust over the oceanic plate(s) of the Pacific basin during the Neogene-Quaternary transverse compression at the site of the Coast Ranges after true subduction had ceased (Page and Brocher, 1994). If this happened, the oceanic crust beneath western California could be of almost any age, and would not necessarily be a source of heat.

Because the excess heat source seems to have been deep in the crust or in the upper mantle, and because it doesn't seem to have required a slab window and isn't restricted to the area of the Monterey plate, perhaps one likely source is shear along subhorizontal surfaces in the deep crust, for example, shear along the upper surface of the supposed oceanic crustal slab beneath the west part of the Coast Ranges province, and the upper surface of the Great Valley basement beneath the east part. Lachenbruch and Sass (1973) reasoned that because the San Andreas fault lacks the local heat flow anomaly that would be predicted for a high-friction fault, the resistance to plate motion may not come from strike-slip friction, but instead may stem from a broad shear zone below the seismogenic layer. Such a broadly generated thermal anomaly would be consistent with the heat flow of the Coast Ranges. Slip along the subhorizontal upper surface(s) of the seismic basement is one likely source. By seismic basement, we mean the supposed oceanic crustal slab(s) beneath the west part of the Coast Ranges province, and the Great Valley basement under the east part. Another source might be the intense late Miocene folding. This deformation apparently has about the same areal extent, offshore and onshore, as the anomalous heat flow, whereas intense Pliocene–Quaternary deformation is largely confined to the onshore, coinciding with the Coast Ranges province.

A puzzling aspect of the heat flow–tectonic problem is the paradox that the central Coast Ranges, characterized by high heat flow, are also characterized by large volumes of the Franciscan Complex, some of which includes blueschist facies metamorphic rocks. The blueschist type of metamorphism requires high pressure, 3 to 8 kbar (300–800 Mp), equivalent to depths of 10–25 km, and relatively low temperature, 150–300 °C (Ernst, 1965); therefore it is believed to be a product of subduction. In the case of Franciscan blueschists, the metamorphism (hence subduc-

TABLE 3. RATES OF UPLIFT

Type of data	Range	Uplift rate (mm/yr)	Reference	Remarks
Stratigraphic/structural relations	Santa Cruz Mountains	ca. 1.4	This paper	Based on relations and age (ca.3–0.45 Ma) of Santa Clara Formation
Stratigraphic/structural relations	Diablo Range	1.4–2	This paper	Based on relations and age (ca. 2.2–0.5 Ma)
Differential uplift of erosion surface on either side of 1.25 Ma fault	Santa Cruz Mountains	0.8–0.36	This paper	Uplift alongside new, dated segment of San Andreas fault
Uplifted marine terraces	Santa Cruz Mountains near base of range	0.16–0.26	Bradley and Griggs (1976)	Uplifted flight of upper Pleistocene terraces near base of mountains
Uplifted marine terraces	Santa Cruz Mountains near base of range	0.17–0.41	Lajoie et al. (1991)	Approximately same territory; improved dates
Uplifted marine terraces	Santa Cruz Mountains near base of range	0.13–0.35	Valensise and Ward (1991)	Used Santa Cruz territory ca. 1.25 ka
Uplifted marine terraces	Santa Lucia Range, northwest base	0.16	McKittrick (1988)	Mean for past 200 k.y.
Uplifted marine terraces	San Luis–Pismo block	0.11–0.22	Pacific Gas and Electric Company (1988)	Territory (ca. 83 ka) uplifted between reverse faults
Theoretical model	Santa Cruz Mountains near Loma Prieta	ca. 1.4–1.8	Anderson(1994)	Realistic input, sophisticated model
Fission tracks	Santa Cruz Mountains near Loma Prieta	ca. 0.8	Burgmann et al. (1992)	Approximate mean, past 4.6 m.y.
Releveling	East Bay Hills and Diablo Range	ca. 1.4–1.8	Gilmore(1993)	Unrefined but suggestive data

tion) is of various ages, younging up to 90 Ma or possibly 70 Ma. The mineral assemblages, formed at depths of 10–25 km, were somehow protected from heating for 30 m.y. or more, and to receive this protection, must have been brought up to shallow levels quickly. They remained sheltered in the upper crust notwithstanding the active deformation during Cenozoic time, and must have avoided the influences at depth that produced today's high heat flow.

Information from Seismic Reflection and Refraction

Knowledge of the Coast Ranges crust in the subsurface has been provided largely by seismic reflection and refraction imaging. In our crustal transects (Figs. 9 and 10), most of the features shown in the middle and lower crust, and the Moho, are derived from seismic traverses by others. Readers may obtain more details from sources cited in the transects. Particularly useful papers include those by Walter and Mooney (1982), Fuis and Mooney (1990), Howie et al. (1993), and Brocher et al. (1994).

Seismic refraction (e.g., Walter and Mooney, 1982) typically shows that rocks and facies seen at the surface may not be present below depths of 8–15 km, depending on the location. This is usually based on P-wave velocities, which commonly increase markedly at certain subhorizontal boundaries or zones. The question arises, What is the nature of these subhorizontal features? Those that are reflective and which separate units with markedly different velocities are presumed to be petrologic and perhaps tectonic; depending on the circumstances, some reflective zones could be detachments or zones of intense ductile shear such as those envisioned at the upper surfaces of “seismic basements” (higher velocity crystalline crust; see crustal transects, Figs. 9 and 10). Less-definitive velocity transitions may mark differences in facies or different degrees of metamorphism within a single rock assemblage such as the Franciscan Complex, or may denote the contact between two very different assemblages having only slightly different elastic properties. In such cases we have made interpretive judgments based on likely lithologic variability within rock complexes (e.g., Salinian magmatic arc rocks or the Franciscan Complex) and consideration of known regional tectonic relations and other factors. Seismic reflection imaging has been a powerful tool for recognizing the presence and character of subsurface structural features in the Coast Ranges and adjacent provinces (e.g., Brocher et al., 1994), and combined reflection and refraction experiments have been very fruitful.

Earthquake Sources and Focal Mechanisms

Maps showing the distribution of earthquake epicenters (e.g., Goter et al., 1994; U.S. Geological Survey, 1990) indicate some of the sites of ongoing crustal activity. Seismicity is dominated by parts of the San Andreas fault (see articles in Wallace, 1990) and other large dextral strike-slip faults of the transform system. These are prominently marked by linear concentrations of epicenters; in addition, apparently random, sparsely scattered epicenters of small-magnitude earthquakes appear within the individual Coast Ranges. This denotes broadly distributed strain involving occasional slip on variously oriented small- to medium-sized faults, and is clearly a symptom of persistent stress and probably ongoing uplift.

Focal mechanisms in the central Coast Ranges subprovince predominantly indicate dextral strike-slip on northwest-trending near-vertical faults of the transform system. However, thrust mechanisms are occasionally evident, notably at the site of the main shock of the 1983 M 6.7 Coalinga earthquake (Rymer and Ellsworth, 1990). The causal fault, which is near the Coast Ranges–Great Valley boundary, strikes N53°W and dips 23° southwest at the hypocenter, which was at a depth of 10 km. The focal mechanism shows that contractional strain transverse to the plate boundary occurs today as in the recent geologic past. This interpretation is strengthened by the thrust mechanism of the 1985 Kettleman Hills earthquake (Ekstrom et al., 1992), which was centered a few kilometers northeast of the margin of the Coast Ranges (see Fig. 10). The M 7.1 Loma Prieta earthquake of 1989 originated on a San Andreas fault segment, which (atypically) dips 70° southwest and which, at the time of the earthquake, showed a reverse component as well as dextral strike slip (Plafker and Galloway, 1989). Active thrust faults along the northeast flank of the Santa Cruz Mountains west of San Jose strike more or less parallel with the plate boundary and dip beneath the range (McLaughlin et al., 1991). Along the same flank of the Santa Cruz Mountains, on the San Francisco Peninsula, small earthquakes show various focal mechanisms, in some of which the interpreted fault planes strike parallel with the San Andreas and dip southwest toward the latter (Kovach and Beroza, 1993; Olson and Zoback, 1995; Kovach and Page, 1995). These and other seismological observations indicate present-day continuation of transverse contractional strain accompanying the concurrent strike-slip activity.

TABLE 4. TRANSVERSE SHORTENING

Range or locality	Transverse shortening	Reference	Remarks
Santa Cruz Mountains approximate full width; Neogene and Quaternary structures	~21 km = 38% of original width, 56 km	This paper	Collective folding and thrusting since middle Miocene
Santa Cruz Mountains between San Gregorio and San Andreas faults; Neogene and Quaternary structures	~12 km = 32% of original distance, about 37 km	This paper	Collective folding and thrusting since middle Miocene
Santa Cruz Mountains, folds in remnants of Purisma Fm. (Pliocene)	~0.28 km = 4.5% of original length, 6.22 km	This paper	Late Pliocene–Quaternary component of deformation only
Santa Cruz Mountains, vicnity of Butano thrust	~2.3 km = 19% of original distance, 12.1 km	This paper	Late Pliocene–Quaternary component of deformation only
Santa Cruz Mountains, between San Andreas fault and Stanford University	~7.7 km = 43% of original distance, 17.7 km	This paper	Quaternary component only
Santa Cruz Mountains, Sierra Azul area southwest of San Jose; imbricated	~4 km = 100% in belt of thrust faults	McLaughlin et al., (1991)	Local shortening by thrusting since Miocene, mostly since about 3 Ma
Diablo Range, entire width north of Coalinga	~6 km = 14% of pre-Tulare (pre-2.2 Ma) width	Namson and Davis (1988)	Pliocene–Quaternary deformation, based on stylized and balanced cross section
Central Coast Ranges, entire width of subprovince	~33 km = 23%	Namson and Davis (1988)	Quaternary component could be less
Pyramid Hills and Kettleman south dome, vicinity of east side of Diablo Range	3.7 km = 17% measured at five different horizons, 70–14 Ma in age; 2.1 km or 9% at horizon 2.5 Ma in age	Bloch et al. (1993)	Meticulous analysis not wholly within the Coast Ranges sensu stricto

PARTICULAR FEATURES OF THE CRUST

Upper Crust

Cenozoic Folds and Thrusts Viewed as a Whole. Intense deformation at shallow depths is widespread except where strong, rigid Salinian basement rocks have resisted regional stresses, as in the central part of the area shown in Figure 10. (The Salinian rocks have not survived as well in the area shown by the Fig. 9 transect, as discussed in another section). Layered formations, mostly sedimentary, are extensive and are commonly folded and cut by faults. The folding is generally not directly reflected in the morphology of the ranges; for example, anticlines are generally not topographic highs. This is commonplace in mountain belts all over the world, but it is surprising here because the folds are so young. Although folds are not directly expressed in the topography, the Coast Ranges province as a whole is strikingly coincident with the belt of intense Quaternary folding (Figs. 9 and 10). To the west of the Coast Ranges in the offshore region, the numerous anticlines and synclines are comparatively mild; the transition from mild deformation to intense deformation seems to mark (imprecisely) the location of the shoreline along most of the coast. There are, however, outliers of strong deformation both at sea and in the margin of the Great Valley. As emphasized throughout this article, many of the folded and faulted rocks are Neogene.

Amount of Upper Crustal Shortening. We have measured shortening in the upper 10 km of crust implicit in the crustal transects of Figures 9 and 10 and in a section across the Santa Cruz Mountains (Fig. 8), and we show the results in Table 3. The table includes some estimates of other authors for comparison. We conclude that shortening across the entire span of the central and southern Coast Ranges is between 15% and 40% of the original width of a hypothetical horizontal stratum that extended from one side of the subprovince to the other. Our preferred estimate is 20%–40%. We think that the entire lithosphere has undergone comparable amounts of shortening, albeit by different mechanisms at different depths; if this is so, the transverse strain denoted by folds and thrusts in the upper crust is an indicator of the amount of shortening of the entire lithosphere. Note the "pseudo-subduction" of lower crust that is postulated in Figures 9 and 10 in order to accommodate shortening equivalent to that observed in the upper crust. It is of interest to see what rates of deformation would be required in order to achieve various postulated amounts of shortening within permissible time spans. As shown in Table 4, most of the calculated rates are on the order of millimeters, rather than centimeters, per year. This is compatible with calculated rates of uplift of the present-day ranges.

With regard to folding and horizontal shortening in the Coast Ranges, we have concluded that (1) the uplifts that elevated the western fringe of California above sea level, where it has remained to the present, commenced at about 3.5 Ma and were followed by folding and thrusting, but (2) the rise of the individual ranges as currently delineated was delayed until about 0.40 m.y. after most of the deformation had occurred. The transverse shortening that is related to the latter episode is accordingly only a small fraction of the total amount that can be surmised from most cross sections showing folded and thrust-faulted rocks. In areas where Pliocene formations have been folded and thrust faulted, and especially where deformed beds have been dated as 450 000 yr or younger, we can make fairly reliable estimates of the amount of Quaternary shortening (e.g., Fig. 8),but clearly most of this, notwithstanding its recency, occurred before the present ranges were uplifted.

Relation Between Upper Crustal Structure and Areal Extent of Central and Southern Coast Ranges Subprovince. Figures 9 and 10 show an unmistakable spatial relation between the Coast Ranges and the broad belt of intense Neogene and Quaternary folding. Although individual folds within the ranges have virtually no direct effect on the topography, mountain ranges are not present beyond the domain of intense folding. To the east, Cenozoic strata of the main part of the Great Valley are not folded or are folded only slightly, except for recent outliers of the Coast Ranges such as Kettleman Hills (Figs. 5 and 10). These outliers are especially numerous in the southwest part of the Great Valley, where they are so young that some are directly expressed by topographic highs and lows of minor relief. They are not yet integral parts of the Coast Ranges. To the west of the province, offshore, there are many Neogene, and some Quaternary, folds that have been mapped by seismic reflection. However, when shown in cross section without vertical exaggeration, most of these structures are strikingly mild compared with their counterparts in the Coast Ranges (e.g., Fig. 10). Incidentally, we suggest that the location of the shoreline, which imperfectly coincides with the boundary of the Coast Ranges province, was somehow determined by the western limit of intense folding. Another feature that approximately coincides with the southwest boundary of the subprovince is the San Gregorio–Hosgri fault zone (Hall, 1975; Silver and Normark, 1978). This zone has undergone large amounts of dextral strike slip, probably as much as 150 km in its northern reaches (Clark et al., 1984), and most of its principal faults are presumably subvertical. We do not know the reason for its near coincidence with the southwest boundary of the central and southern Coast Ranges. The northeast margin of the subprovince is marked by the northeast-dipping homocline of the Great Valley sequence (Figs. 9 and 10), which owes its genesis to the progressive easterly insertion of a subsurface wedge of mainly Franciscan rocks (see the following section). Clearly the homocline and the wedge are relevant to the origin of the eastern Coast Ranges.

Middle Crust

We consider the middle crust in provincial terms that apply to the central and southern Coast Ranges subprovince. It cannot be defined in terms of composition, which is largely unknown. We regard it as the crustal domain from a depth of about 8–9 km down to the upper boundary of (likely mafic) material characterized by P-wave velocities of 6.6 to 7.1 km/s. On the basis of these criteria, the middle crust is thickest (10–15 km) along the axial region of the subprovince. It generally transmits P-waves at velocities from 5.6 to 6.2 km/s, but this range is locally slightly exceeded. The lower velocities overlap those of the upper crust, where velocities are commonly 3.5 to 5.8 km/s (or even higher in Salinian rocks).

The composition and structure of the middle crust throughout most of the central and southern Coast Ranges are not known with certainty. The generally higher P-wave velocities, as compared with the upper crust, suggest any of the following explanations: the closure of pores and cracks; the presence of somewhat different facies of the same rocks that are seen at the surface; the transition of certain minerals to denser polymorphs; the presence of rock assemblages that are fundamentally different from those at the surface; the presence of low-angle contacts, perhaps gradational; the presence of low-angle tectonic contacts; and so forth.

The middle crust beneath most of the ranges probably consists of Franciscan rocks. This idea is partly based on the near certainty that the Franciscan is an accreted subduction complex of large lateral and vertical extent. Parts of the complex have undergone conditions at great depths and high pressure, as indicated by blueschist facies mineral assemblages. Characteristic variations in petrology and degree of metamorphism can readily account for differences in seismic velocities. Where the upper crust is predominantly Franciscan rocks, the middle crust with higher Vp may also be Franciscan, perhaps with more abundant greenstone or high-*P* metamorphic rocks. (Franciscan rocks may be absent and other assemblages may be present.)

At the surface, the core of the Diablo Range consists of Franciscan rocks. These are extensively exposed, but their downward limit is not known with certainty. Although subhorizontal elements are rarely seen at the surface, puzzling discontinuous low-angle seismic reflectors are imaged locally at depth, and seismic refraction shows one or more subhorizontal velocity boundaries that apparently occur within the Franciscan Complex (Walter and Mooney, 1982). A rather continuous P-wave velocity boundary (about 5.9 vs. 6.7 km/s) occurs in the Diablo Range at a depth of about 16 km, probably marking the contact between Franciscan rocks and underlying mafic basement (Walter and Mooney, 1982). This boundary is adopted in our Figure 10.

Of paramount importance is the presence of wedge(s) of upper or midcrustal Franciscan rocks inserted laterally beneath Great Valley sedimentary rocks along the northeast flank of the Coast Ranges (e.g., Wentworth et al., 1984). Such wedges of middle crustal material are evidence of horizontal compressional strain at least as far down as the top of the mafic lower crust, and their emplacement must have required uplift of the overlying material.

Our ignorance concerning midcrustal material beneath known Salinian magmatic arc rocks (including granitic types) is particularly frustrating. Vp values are high enough to accommodate the hypothesis that the Salinian block is a "flake" that has overridden unrelated, unseen assemblage(s); on the other hand, the observed velocities could be explained by the presence of Salinian anisotropic gneisses and/or schists with high horizontal velocities. The Gabilan Range is one area where seismic evidence favors an interpreted thin (about 8–9 km) Salinian granitic upper crust lacking reflectors, below which a reflective unit extends to a depth of at least 13 km (see Hale and Thompson, 1982). We have adopted an inferred boundary between granite and underlying schist or gneiss in our crustal transect, Figure 10. Just beneath the granite there may be schist comparable to the Pelona Schist of southern California (see Ross, 1976). Between this and the top of the mafic lower crust at about 21 km, the nature of the crust is even less certain. In any case, foliation alone probably does not account for the reflectivity, which Beaudoin (1994) suggested is tectonic banding resulting from ductile deformation during large-scale transport of the Salinian block.

It is important to note in our crustal transects (Figs. 9 and 10) that the combined upper and middle crust are appreciably thicker in the central and southern Coast Ranges than in adjoining domains to the east and west, suggesting a relationship between the thickening and the existence of the subprovince.

Lower Crust

The lower crust is taken to be a deep layer having P-wave velocities generally ranging from 6.6 to 7.1 km/s. Beneath the western half of the central and southern Coast Ranges subprovince, this lower crust apparently consists of oceanic mafic rocks, which can be tracked with a fair degree of confidence by means of seismic reflection and refraction from the crust of the Pacific basin inland. The relevant research has been concentrated in a large offshore-onshore region around San Luis Obispo and a comparable large offshore/onshore region near San Francisco. The results in the San Luis Obispo region were summarized by Howie et al. (1993) and were tested and refined by Lafond and Levander (1995); the results near San Francisco were presented by Brocher et al. (1994) and refined by Holbrook et al. (1996). In both regions a slab-like (presumably mafic) layer descends at a low angle beneath the edge of the continent and is nearly level under the median axis of the central and southern Coast Ranges subprovince (Figs. 9 and 10).

The current relative plate motion, involving dextral horizontal shear, dominates the regional tectonics with respect to total accrued displacements and present-day strain rates. It must involve the entire lithosphere, and the presence of an apparently unbroken subhorizontal lower crust seems incompatible with the large-scale strike-slip faulting observed at the surface. Do the large strike-slip faults, particularly the San Andreas fault, cut and offset the lower crust in some undetected manner? Is the slip on the near-vertical San Andreas fault parlayed into slip on the subhorizontal upper surface of the mafic lower crustal slab (e.g., Brocher et al., 1994; Jones et al., 1994)? This question is still not answered with certainty, but teleseismic shear wave splitting indicates that the upper mantle is highly anisotropic in a broad band beneath the Coast Ranges, the fast direction being aligned parallel with the San Andreas and its companion faults (Ozalaybey and Savage, 1995). According to Ozalaybey and Savage (1995), the anomalous condition prevails in the upper 115–125 km of the mantle in a belt about 100 km wide. This strongly suggests that some modified form of the San Andreas fault system penetrates the entire lithosphere, as would be expected from the basic characteristics of plate tectonics. It is not clear whether individual faults (e.g., San Gregorio, San Andreas, Hayward) that are seen at the surface

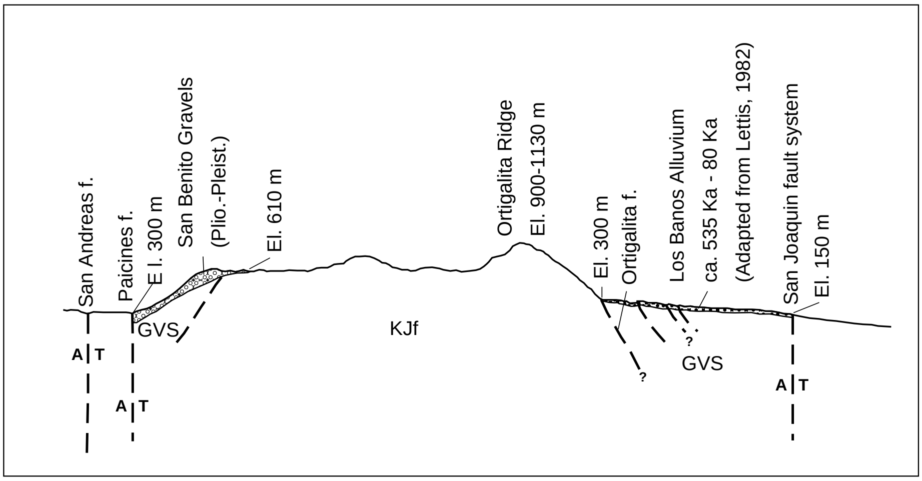

Figure 11. Diagrammatic profile showing partial capping of San Benito Gravels (Pliocene–Pleistocene) and beheaded Los Banos alluvium (Pleistocene) affected by uplift of part of the Diablo Range, showing recency of the uplift. KJF—Franciscan Complex; GVS—Great Valley sequence.

have individual expression in the mantle, but it seems unlikely that the subhorizontal slab-like lower crust could be unbroken between a highly sheared upper crust and a shear zone 100 km wide in the upper mantle. Possibly horizontal offsets of the subhorizontal lower crust, even frequently repeated offsets, have not produced vertical displacements large enough, in most cases, to be detected in the available seismic images. There seems to be a step in the upper surface of the lower crustal slab beneath the surface trace of the San Gregorio–Hosgri fault in a seismic profile by Howie et al. (1993). In Figure 10 we have slightly modified their profile and have extended the fault through the entire crust. No such step is seen beneath the other major faults in Figures 9 and 10, but we have speculatively extended the San Andreas fault and several other faults, through the crust.

With regard to the foregoing problem, the time of emplacement of the mafic slab-like body or bodies composing the lower crust is relevant. In the transect of Figure 10, the lower crust is likely part of the Monterey oceanic microplate, which was subducted before 19 Ma (Lonsdale, 1991; Severinghaus and Atwater, 1990; Nicholson et al., 1994). Inasmuch as the slab has probably been in place for 19 m.y., one might expect it to be offset by large amounts of strike slip, which would probably juxtapose irregularities in the surface, producing conspicuous step-like features; for the most part, however, this expectation is not realized.

The lower crust in the southwestern part of Figure 10 underlies the Santa Cruz Mountains. It is north of the Monterey plate and must have a different ancestry, although it also is considered to be oceanic crust (Brocher et al., 1994). It may be a microplate (as yet undescribed) subducted many millions of years ago; it may be a leftover piece of the Farallon plate; or, it may be Pacific plate lithosphere that was overthrust by the continent during the transform regime (Page and Brocher, 1993). Inasmuch as its upper surface does not share the intense deformation of the upper crust, we concur with Brocher et al. (1994) in regarding the surface as a detachment plane or zone, whether or not it participates in relative plate motion. At depths below the brittle-ductile transition, the shear is most likely distributed and the locus may not be a definitive detachment in the usual sense of the term. The lower crust beneath the San Francisco Bay region apparently passes laterally under the surface trace and earthquake hypocenters of the San Andreas fault without conspicuous interruption.

The northeast part of the central and southern Coast Ranges subprovince is underlain by part of the Great Valley basement, which, although petrologically different, may play a mechanical role similar to that of the mafic slab-like lower crust to the southwest. The pronounced magnetic character of the Great Valley basement is presumably caused by serpentine and perhaps gabbro. In the case of serpentine, the protolith was ultramafic rock, probably mantle material. However, the seismic velocities are too low for unaltered mantle peridotite. Farther north, Godfrey et al. (1997), have carried out seismic reflection-refraction profiling plus density and magnetic modeling of the crust beneath the Great Valley. They interpret the Great Valley ophiolite, including high-velocity ultramafic rocks, to compose an obducted slab of oceanic lithosphere resting tectonically on Sierra Foothills–related crust. In Figures 9 and 10 we are influenced by their interpretation of an obducted slab; we show ultramafic (presumably mantle) rock, partly or largely serpentinized, tectonically emplaced over unrelated crust. We do not attempt to identify the latter.

The Sierra Nevada Foothills rocks include Paleozoic and Mesozoic metasedimentary rocks and accreted Mesozoic island-arc assemblages. All of these rocks are locally intruded by granitic to gabbroic plutons. Under the eastern part of the Great Valley, the foothills rocks are overlain, above a gently sloping planar unconformity, by upper units of the Great Valley sequence. The unconformity continues southwestward in the subsurface, but the underlying Sierran rocks give way to the Great Valley magnetic basement. The exact location and configuration of the contact between the two basements are unknown, so in Figure 10 we have simply shown a vertical contact. The actual contact may represent paleorifting between the not-yet-accreted Mesozoic island arc(s) of the foothills and a seaward backarc basin (Schweickert, 1981). In this interpretation the oceanic crust and mantle underlying the back-arc basin have become the Great Valley basement.

Moho

The Mohorovicic discontinuity has not been located continuously beneath the central and southern Coast Ranges subprovince, but there is considerable agreement between a number of separate data sets. On the oceanward side of the subprovince, the Moho within the base of the oceanic slab accompanies the latter beneath the edge of the continent (e.g., Howie et al., 1993; Brocher et al., 1994). This agrees well with the Moho determined farther north by Walter and Mooney (1982). Beneath the northeast half of the subprovince, the Moho is at the base of a lower crust that is much thicker than that to the southwest, and it descends at a low angle toward the Sierra Nevada (Ruppert et al., 1998). Evidence from xenoliths suggests that the Moho beneath the western Sierra Nevada may be a boundary at the top of a layer of eclogite rather than peridotite, according to Ducea and Saleeby (1996). Currently available data do not show any lateral jump or discontinuity in the depth to Moho under the central and southern Coast Ranges, even across the Rinconada and San Andreas faults (Fig. 10), both of which are loci of large crustal displacements. Thus, the transverse shortening that appears to be a fundamental feature of the strain within the Coast Ranges seems to have been accommodated in a cryptic way at the level of the Moho, judging from the incomplete data available.

BIRTH AND RISE OF THE PRESENT RANGES IN THE QUATERNARY

Structural and Paleogeographic Relations Between Pliocene–Pleistocene Deposits and the Ranges

The areal distribution of nonmarine Pliocene–Pleistocene sedimentary rocks in the central and southern Coast Ranges subprovince (Fig. 5) signifies widespread emergence of western California, but the advent of nonmarine conditions does not coincide spatially with individual mountain ranges of the present day. It was more general and widespread than the more recent uplifts.

There are two main types of relations between the ranges and nonmarine Pliocene–Pleistocene deposits. In the first type, the young nonmarine deposits are draped over the flanks of the ranges and have been tilted and uplifted as a partial blanket. In the second type, folded Pliocene–Pleistocene formations are truncated by faults at or near the margins of the ranges, and are not preserved in the second type. The first type implies uplift with some arching or tilting of superincumbent Pliocene–Pleistocene sedimentary rocks (Fig. 11). One might argue that there was a near continuum from withdrawal of the sea, nonmarine deposition fed by further uplift, and eventual rise of the present ranges. In the second case, a more discrete tectonic discontinuity is implied, between the deposition of nonmarine formations and the truncation of these by faulting involved in the uplift of the modern ranges. Both types of relations show that today's mountain ranges are markedly younger than the Pliocene–Pleistocene formations.

In some areas, changes in drainage directions after deposition of Pliocene–Pleistocene fluviatile sediments prove that the locus of uplift shifted. Christensen (1965) noted that the paleocurrent directions in the Pliocene–Pleistocene San Benito Gravels and the Pliocene–Pleistocene Hans Grieve Formation (both in the Diablo Range) differ strikingly from present-day directions. Such changes, including reversals in direction of flow, presumably occurred as the modern ranges rose. These circumstances reinforce the conclusion that the rise of the ranges was not, in spatial terms, simply a continuation of the uplift that fed the Pliocene–Pleistocene nonmarine deposits.

The paleogeography of the Paso Robles Formation (Galehouse, 1967) is instructive. Galehouse found that the paleodrainage area in the Paso Robles region bypassed the present Salinas Valley and nearby ranges as we now know them, even though the headwaters of the system were in areas now encompassed by the Santa Lucia and Sierra Madre Ranges. The streams flowed eastward, southeastward, and northeastward across the sites of the future Salinas Valley and Gabilan Range (Fig. 12). Some believe that the drainage continued across the San Andreas fault and site of the Temblor Range (Fig. 7) to the San Joaquin Valley; however, there is no direct proof of this idea. The point to be emphasized is that the present Gabilan Range and its neighbors are even younger than the Pliocene–Pleistocene Paso Robles Formation and are not closely coincident with the uplifts that fed sediment to that formation.

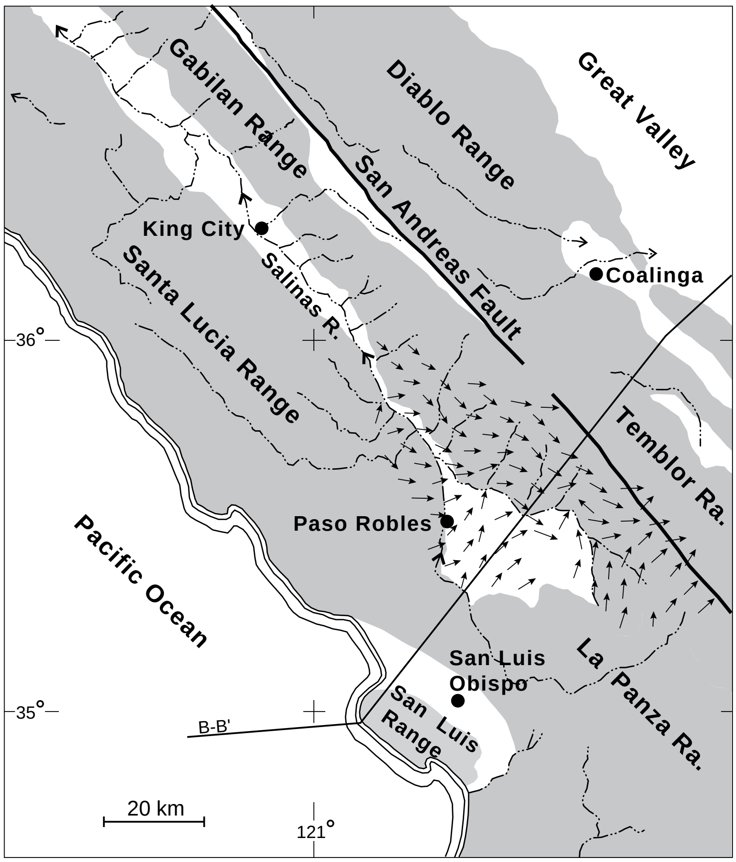

Figure 12. Map contrasting Pliocene–Pleistocene paleocurrent directions with locations and trends of present-day ranges. Arrows show moving averages of paleocurrent measurements by Galehouse (1967).

The rise of the Diablo and Temblor Ranges in the Pleistocene accentuated the blocking of the San Joaquin Valley, which had begun to lose its connection with the ocean in Pliocene time at the inception of Tulare deposition. Pliocene marine sedimentary rocks extend nearly across the Diablo Range south of the New Idrea serpentine mass, so the blockade was apparently incomplete at this latitude until latest Pliocene or early Pleistocene time.

Timing of Uplift

Not only did the original distribution of the Pliocene–Pleistocene formations transgress the future range boundaries, but locally, these young deposits were folded prior to, or synchronously with, the beginning of uplift of the present ranges. Figures 8, 9, 10, and 14 show examples of this relationship in the Santa Cruz Mountains, Diablo Range, Sierra Madre Range, and Caliente Mountain. Probably all of the individual Coast Ranges are younger than Pliocene–Pleistocene formations such as the Paso Robles, Tulare, and Santa Clara.

As summarized in Table 1, upper horizons of the Pliocene–Pleistocene Santa Clara Formation appear to be as young as 0.4–0.47 Ma on the basis of tephra dated by Sarna-Wojcicki (1976) and Sarna-Wojcicki et al. (1985); the Irvington Gravels are probably 1.9–0.45 Ma, on the basis of vertebrate faunas (Lindberg, 1984); the upper Livermore Gravels are likely 0.45–0.6 Ma, on the basis of vertebrate fossils (E. E. Brabb, 1980, personal commun.); and the upper part of the Tulare Formation contains several Pleistocene tephras, including the Bishop ash and Lava Creek B ash (0.758 Ma and 0.62 Ma, respectively; Sarna-Wojcicki et al., 1991). Incredible as it seems, apparently the Coast Ranges as currently delineated began to rise about 400 000 yr ago; this amazing

Figure 13. Relation between the Gabilan Range and the San Andreas fault, looking west across Cienega Road south of Hollister. The wide fault zone passes obliquely across the view from lower left to the white buildings (approximately 0.6 ha) and beyond, forming (approximately) the boundary of the adjacent Gabilan Range. The latter, despite deep erosion, retains block-like characteristics, including nearly level summit remnants of a broad paleoerosion surface presumably formed near sea level. The vertical rise of the range made expedient use of the preexisting San Andreas fault, the main function of which is dextral strike slip.

date is subject to revision as new data become available.

Near-Surface Clues to Modes of Uplift of Present-Day Ranges

We have discussed geomorphic indicators of the several styles of uplift. These indicators and other near-surface features show that the various ranges did not rise in precisely the same way, although they were more or less synchronous and almost all show important transverse contraction. Among the modes of uplift are high-angle block faulting, marginal thrusting, welt-like ductile arching, and mixed modes that defy categorization. Most of the ranges vary in aspect from place to place, and most show combinations of different styles.

High-Angle Block Faulting. The northern part of the Gabilan Range exemplifies high-angle block faulting. This part of the range is bounded by an unseen (reverse?) fault on the southwest side, surmounted by a deeply dissected but steep mountain front that overlooks an apron of large alluvial fans sloping down into Salinas Valley. The northeast side of the range is also steep, and descends to the San Andreas fault. A relatively small part of the summit area exhibits nearly flat pediment surfaces (Fig. 13). Thus, the northern Gabilan Range appears to have risen as a rigid fault-bounded block, the fault on one side being an active strike-slip rupture. The southern part of the same range is low and is known as the Gabilan Mesa. It has no significant fault on the southwest side, and it rose simply by tilting as a crustal block alongside the San Andreas fault. East of Paso Robles the nonmarine Pliocene–Pleistocene Paso Robles Formation forms a gently southwest-dipping homoclinal cap, which extends across much of this part of the range. The cap is locally slightly flexed, but its relatively minor deformation, together with the semiplanar surface of the underlying bedrock (Salinian granite), shows that this part of the range behaved rigidly, like the higher northern part. Tilting of this portion of the range reversed the regional drainage, which had previously flowed eastward, northeastward, or southeastward (Fig. 12; Galehouse, 1967).

The northern part of the Santa Lucia Range (Fig. 1) is a complex, composite fault block. As we have described, the northeast side (Fig. 4) is a deeply eroded fault scarp at the foot of which is the trace of a high-angle oblique-slip dextral reverse fault, a segment of the King City fault. The southwest side of the range at this latitude descends precipitously to the Pacific Ocean along the scenic Big Sur coast. This side is imprecisely delineated by the active San Gregorio strike-slip fault.

The fact that the high-angle fault blocks described here are variously bounded by oblique reverse-slip faults and by strike-slip faults shows that these ranges are not Basin and Range–type extensional features, and suggests that they were boosted upward by forces that took advantage of preexisting high-angle faults. Some of the adopted faults are active, and strike-slip surface features and seismic focal mechanisms show dextral strike-slip. Thus, these faults are still carrying on their original functions, which had little to do with the uplift of mountains. The stress, which elsewhere in the subprovince produces contractional strain normal to the plate boundary, is almost certainly acting upon fault blocks such as the Gabilan Range. The uplift of these blocks is probably ultimately caused by compression, although the exact means is obscure.

Marginal Thrusting. High-angle faults are important in the type of block faulting described here. In contrast, parts of some range margins are defined by thrust faults that dip at a low to moderate angle beneath the mountains. Examples include the East Evergreen fault at the southwestern base of the Diablo Range foothills southeast of San Jose (Fig. 14). On the opposite side of the Santa Clara Valley, the Sargent-Berrocal fault system (McLaughlin et al., 1991) consists of thrusts and oblique-slip faults dipping southwest beneath the edge of the Santa Cruz Mountains. Apparently, a few of these faults are active at the surface, and similar thrusts in the Stanford–Palo Alto area produce small earthquakes at depth (e.g., Kovach and Page, 1995). The focal mechanisms verify the thrust sense of slip. These faults undoubtedly facilitate the rise of the ranges and they prove the ongoing existence of compression transverse to the plate boundary.

Strike-Slip Boundary Faults. The Gabilan Range and Temblor Range (Fig. 1) are bounded (approximately) by the San Andreas fault. The Gabilan Range adjoins the fault on the southwest side and the Temblor on the northeast side (Fig. 12). The Temblor Range rises above the northeast side of the fault, which, however, only is immediately at the foot of the range for a distance of 17 km and elsewhere is 1–3 km from the base of the mountains. It should be noted that although both the Gabilan Range and the Temblor Range show a spatial relationship to the San Andreas fault, they are on opposite sides of it. Thus, it cannot be said that the San Andreas fault throughout its length shows a consistent component of dip slip. We think that the fault happened to be present when uplifts were about to occur, and the uplifts took advantage of a ready-made locus for vertical movement. From about 40 to 130 km southeast of Hollister, both the Gabilan and the Diablo Ranges are bounded by the San Andreas fault. The fault is between the two ranges, its surface trace generally marked by a system of narrow linear valleys and ridges. On the San Francisco Peninsula, the main body of the Santa Cruz Mountains rose alongside the San Andreas fault, leaving an adjacent foothills belt on the other side of the fault at a distinctly lower elevation (Fig. 15). The San Andreas fault is not the only strike-slip fault that has facilitated uplift. A long segment of the Calaveras fault is between differentially uplifted portions of the Diablo Range (Fig. 2), and another segment of the same fault forms the eastern boundary of the East Bay Hills (Fig. 10), which in turn are bounded on the west by the Hayward fault. The main body of the Diablo Range is on the northeast side of the Calaveras fault, whereas the East Bay Hills are on the southwest side, reinforcing the concept that the major strike-slip faults merely played a passive role in the rise of the ranges.

Indefinite and Mixed Modes of Uplift. Some of the Coast Ranges locally lack definitive boundaries. Marginal faults may be absent or inconspicuous. Parts of numerous range fronts are gently sloping and deeply embayed. The interiors of some mountains are topographically recessive, i.e., the central parts are lower than crests on either side. This condition cannot always be explained by erosion. Examples include parts of the Diablo and Santa Lucia Ranges.

Nonuniform uplift behavior is well shown in the Diablo Range southeast of Hollister. Figure 11 shows Pliocene–Pleistocene San Benito Gravels (nonmarine) partially lapping over the flanks of the range and tilted in a manner suggesting an arching type of uplift. One facies of the gravels was partly derived from preliminary uplift of part of the site of the Diablo Range (Griffin, 1967). However, the area of deposition encroached over the position of the present mountains, and the gravels were tilted and partly eroded away during the rise of the latter. Remnants of the gravels are seen, with some fault offsets, from an elevation of about 300 m at the foot of the range to more than 610 m near Panoche Pass (Ernst, 1965, Plate 1). On the other side of the mountains, the topographic surface drops from a high crest down a steep eroded fault scarp. The fault at the foot of the scarp beheads a gently dipping blanket of the Los Banos alluvium, which is dated at 0.06–0.12 Ma (Lettis, 1982). The Los Banos alluvium forms a broad fragmentary cap in the foothills. Surprisingly (to us), the alluvial blanket is offset by two or three minor reverse faults that dip away from, rather than toward, the axis of the range (Lettis, 1982, Plate 20). The faulted alluvial blanket descends from elevations exceeding 300 m to about 150 m at the brow of a low fault scarp delineating the edge of the San Joaquin Valley. These several relations show that this part of the Diablo Range rose differentially in the late Quaternary by some means that left very diverse clues at the surface.

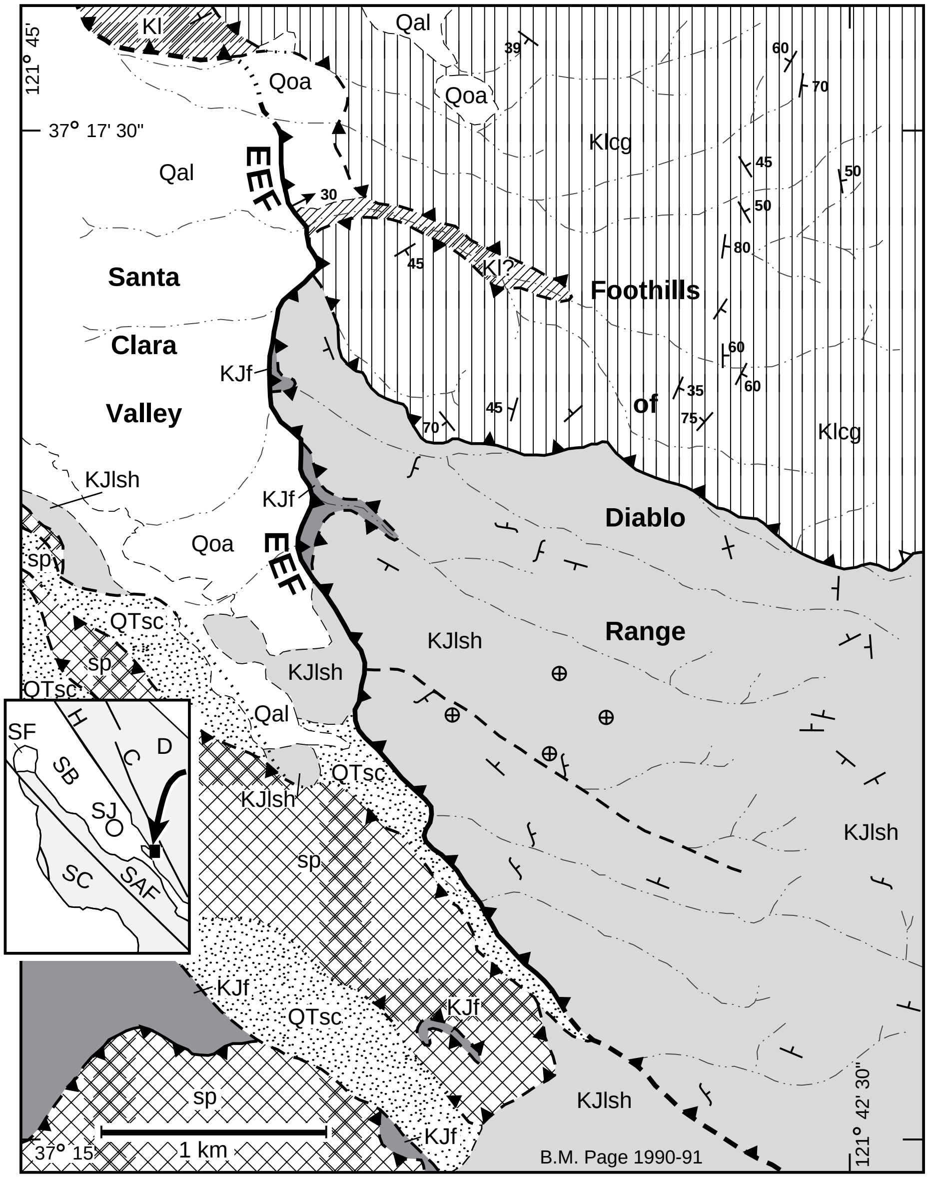

Figure 14. Map of marginal thrust fault at southwest base of Diablo Range near San Jose. This part of the East Evergreen fault, which bounds the range for only a few kilometers, offsets the Pliocene–Pleistocene Santa Clara Formation but lacks evidence of Holocene movement. Inset shows location. Abbreviations: C—Calaveras fault; D—Diablo Range; EEF—East Evergreen fault; H—Hayward fault; KJlsh—Upper Jurassic and Lower Cretaceous (?) shale; KJf—Franciscan Complex; Kl—Lower Cretaceous sandstone and shale; Klcg—Lower Cretaceous conglomerate; Qoa—older alluvium; Qal—younger alluvium; QTsc—Santa Clara Formation; SAF—San Andreas fault; SB—San Francisco Bay–Santa Clara Valley depression; SC—Santa Cruz Mountains; SF—San Francisco; SJ—San Jose; sp—serpentine.

The interior of the Diablo Range has many remnants of a mature upland erosion surface. South of Mount Hamilton, these remnants are offset vertically by unseen high-angle faults, showing that discretely separated parts of the mountains have risen faster and farther than adjacent parts. North of Mount Hamilton, concordant ridges and a few broad uplands preserve vestiges

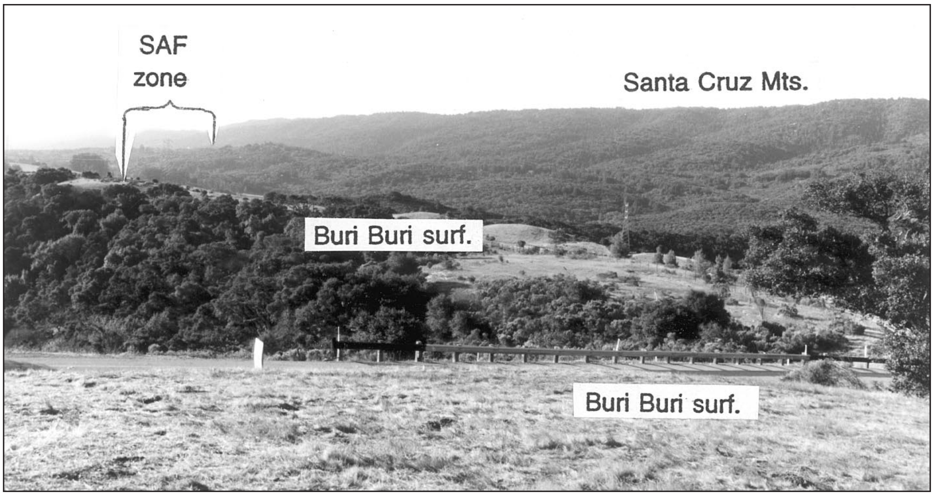

Figure 15. Relation between the San Francisco Peninsula segment of the San Andreas fault (SAF) and the Santa Cruz Mountains. This active segment of the San Andreas is only about 1.25 Ma in age, having replaced an ancestral SAF now hidden in the forested mountain slope in the background. The Santa Cruz Mountains in the distance rose alongside the young segment of the SAF, attaining their present elevation since 1.25 Ma. We believe that the Buri Buri topographic surface (approximately 3 km exposed in photo) in the foreground and middle distance was once coextensive with the summit upland on the distant skyline.

TABLE 5. TRANSVERSE SHORTENING RATES

Present width, central and south Coast Ranges	Estimates of amount of shortening	Time span considered	Apparent average rate (mm/yr)
~120 km	20%–40%; 24–48 km (our preferred estimate)	Middle Miocene to middle Pleistocene, 10.5 m.y.	2.3–4.6
~120 km	15%; 18 km	Middle Miocene to middle Pleistocene, 10.5 m.y.	1.7
~120 km	20%–40%; 24–48 km	3.5–0.45 Ma; 3.05. m.y.	6.9–15.7
~120 km	15%; 18 km	3.5–0.45 Ma; 3.05 m.y.	5.9

of the same rolling paleosurface, here at a generally lower elevation. Thus the mechanisms of uplift acted unevenly and allowed large and small parts of ranges to rise either farther, or not as far, as neighboring parts.

Interplay Between the San Andreas Fault and Coast Ranges Uplifts

As mentioned above, the San Andreas fault and individual ranges are virtually parallel, yet the fault crosses the Coast Ranges province obliquely from one side to the other (Fig. 1). The San Andreas fault originated long before the mountains; however, during the Quaternary, mountain building and strike-slip faulting have been active concurrently, apparently responding to two different influences. The Temblor Range rose along the northeast side of the San Andreas fault, while the Gabilan Range rose along the southwest side. Where the Gabilan and Diablo Ranges are adjacent to one another for a distance of more than 80 km, the San Andreas fault forms the boundary between them.

The Santa Cruz Mountains and the San Andreas fault share a close spatial relationship. From southeast to northwest, the fault is on the northeast side of the range, then follows the crest, makes a small angular bend north of Loma Prieta, follows a course northeast of the crest, and finally (near the ill-defined terminus of the range) crosses to the southwest side and passes out to sea. It is clear that the faulting mechanics did not produce the range (although the bend probably played a role), and that the rise of the mountains did not terminate the fault activity. The bend in the San Andreas fault north of Loma Prieta, in conjunction with dextral slip, evidently increased transverse compressive strain (Valensise and Ward, 1991; Burgmann et al., 1994). The San Andreas fault near Loma Prieta has an anomalous dip (about 70° southwest) and at the time of the 1989 earthquake showed a reverse component of slip, consistent with transverse compression. The bend may have a long-term effect, as the Santa Cruz Range is much wider near it than elsewhere, and thrusting at the base of the northeastern foothills is pronounced (McLaughlin et al., 1991).

Rates of Uplift

Uplift rates have been estimated by a number of authors using various methods, which are briefly discussed. Typical results are compiled in Table 5. From the measurements and calculations summarized in Table 5 and discussed here, we conclude that today's ranges have probably risen at rates between 0.1 and 2.0 mm/yr. If caused by lateral compression, the uplifts would require only a small amount of transverse shortening per year, proportional to the vertical thickness of the compressed slab.

Clues from Geologic Circumstances. The fact that the ranges for the most part consist of weak rocks and yet are topographically high indicates that uplift was geologically rapid and that it is probably ongoing, otherwise, erosion and mass wasting would have reduced the high areas. Large-scale landslides, piedmont debris flows, and alluvial aprons bespeak of destructive processes, which, although vigorous, have been unable to completely counter the persistent uplift. The scarcity of through-going antecedent streams across the ranges and across subsidiary uplifts such as the East Bay Hills and the foothills of the Diablo Range southeast of San Jose can only mean that uplift has been too rapid for successful crosscutting.

As we have emphasized, most of the uplift of the present mountains postdates typical Pliocene–Pleistocene nonmarine formations bordering the ranges. The Santa Cruz Mountains in their present configuration postdate the Santa Clara Formation. Near its Pliocene base, the Santa Clara contains sea-level sedimentary rocks; near its top (presumably) it contains the Rockland ash, which is between 0.40 and 0.47 Ma (Sarna-Wojcicki, 1996, personal commun.). Inasmuch as the mountains in many places are about 650 m above sea level, the average uplift rate was probably at least 1.4 mm/yr. Similarly, the Diablo Range in its present form postdates the Santa Clara Formation and the Tulare Formation. The latter contains the 0.665 Ma Lava Creek B ash (Sarna-Wojcicki, 1996, personal commun.) in the upper part and sea-level beds near the Pliocene base. Many portions of the range are about 900 m above sea level, so these parts may have risen about 2 mm/yr, on the basis of the age of the youngest beds of the Santa Clara Formation, or 1.4 mm/yr on the basis of the age of the Lava Creek B ash in the Tulare Formation. The Temblor Range, which is about 915 m above sea level, rose about 1.4 mm/yr, on the basis of the age of the Tulare Formation. In summary, these crude geologic estimates of uplift rates are around 1.4 mm/yr, but may be as much as 2 mm/yr.

A unique circumstance allows a rough estimate of the minimum uplift rate of the Santa Cruz Mountains on the San Francisco Peninsula. As indicated in Figure 15, a "new" segment of the San Andreas separates two parts of an erosion surface that is now at different elevations on the two sides of the fault. This segment of the fault replaced an old segment ca. 1.25 Ma, and has slipped along strike approximately 20 km since that time. If we undo 20 km of dextral strike slip, we find a difference of elevation on the order of 100–450 m between the undulating summit area of the main range around Montara Mountain and remnants of the upland Buri Buri surface on the foothills (on the opposite side of the San Andreas fault near Redwood City and Woodside) that were formerly in juxtaposition (Fig. 15). This suggests, but does not prove, that the difference in elevation may have been achieved at a rate of 0.08–0.36 mm/yr, and that the major part of the rise of the main range was attained at almost as modest a rate. However, uplift may have occurred during only a part of the time interval, at a higher rate.

Data from Marine Terraces. Marine terraces can provide rates of uplift if they can be dated (which is usually difficult) and if eustatic changes in sea level are taken into account. Unfortunately, along most of the central coast of California, the terraces make an acute angle with the long axes of the ranges; nevertheless, valuable data have been obtained.

Bradley and Griggs (1976) determined that a flight of 6–8 marine terraces along the flank of the Santa Cruz Mountains 12–25 km northwest of Santa Cruz have been progressively elevated and gently tilted or arched. They noted that the older, higher terraces slope seaward at 9 m/km, whereas the youngest wave-cut bench slopes only 1 m/km, suggesting that the mountainside has tilted during uplift of the range. Their data on terrace age vs. elevation indicated that the southwest flank of the Santa Cruz Mountains at a distance of 10 km from the crest has risen 0.16–0.26 mm/yr. These results are similar to revised rates, 0.17–0.41 mm/yr, by Lajoie et al. (1991), who used improved age data. Following the 1989 Loma Prieta earthquake, Valensise and Ward (1991) reexamined the Santa Cruz terraces and carried the observations farther east and southeast along the shore of Monterey Bay, obliquely across the structural grain of part of the mountains. After taking into account changes in sea level, they found average uplift rates of 0.13–0.35 mm/yr. Others (e.g., Anderson, 1990) have inferred rates that are more than twice as high, but most estimates are within the same order of magnitude.

To the south, McKittrick (1988) found that marine terraces at the northwest end of the Santa Lucia Range (Fig. 1) have risen at a mean rate of about 0.16 mm/yr during the past 0.2 m.y. Farther southeast along the Sur Coast, several terrace remnants on the mountainside of the Santa Lucia Range indicate uplift rates of about 0.1–0.3 mm/yr (K. R. Lajoie, 1994, personal commun.). Near San Luis Obispo, the Pacific Gas and Electric Company made a detailed survey of marine terraces across the small San Luis Range, which is a northwest-trending uplifted block about 15 km wide. The ages, elevations, and longitudinal profiles of terraces at several levels show that the San Luis block has risen en masse 0.11–0.22 mm/yr, without folding, during the past 0.5 m.y. (Pacific Gas and Electric Company, 1988, p. 2/37–2/41). A similar investigation by the same company showed that the small Casmalia Range has risen at a rate of 0.14–0.17 mm/yr.

In summary, marine terraces consistently show uplift rates of 0.1–0.4 mm/yr, the mean probably being near 0.2 mm/yr. It should be noted that, except for the measurements across the San Luis Range, the data relate to the flanks of mountains rather than crestal areas, and therefore may show only fractions of the rates at the crests.

Uplift Rates Derived from Releveling. First-order spirit leveling and releveling has been carried out across some of the Coast Ranges and parts of ranges, but to the best of our knowledge, it has not been analyzed and corrected by modern means. Preliminary results of releveling along highways and railroads were studied by Gilmore (1992), who reported apparent progressive upwarping of the East Bay Hills between the Hayward fault and Calaveras fault in the interval 1912–1965. Comparable uplift occurred across Altamont Pass in the Diablo Range during the same time span, and westward tilting occurred across Pacheco Pass in the same range east of the Calaveras fault, 1933–1989. The apparent cumulative changes in height are as much as 70–90 ± 20 mm over 50 yr time spans, suggesting uplift rates of roughly 1.4–1.8 mm/yr. These apparent rates are compatible with geologic estimates summarized here. Some other lines of leveling and re-leveling showed no systematic uplift, and some showed changes in elevation restricted to the vicinity of active faults.

Fission Track Studies. Fission track studies were carried out as part of a comprehensive analysis of uplift of a portion of the Santa Cruz Mountains by Burgmann et al. (1994), in the region around Loma Prieta. In 1989 an Ms 7.1 earthquake in this region originated on the San Andreas fault or a steeply dipping affiliated fault. Fission tracks in apatite grains in sedimentary rocks on the northeast side of the San Andreas fault gave ages averaging 4.6 ± 0.5 Ma, representing the time of cooling below approximately 110 °C. It is estimated that approximately 3 km of unroofing must have occurred subsequently, to expose the rocks that were sampled. Allowing for the present elevation of about 1 km, this suggests an average uplift rate of approximately 0.8 mm/yr during the past 4.6 m.y. (Burgmann et al., 1994). Two samples of Salinian granodiorite (which was emplaced 91–103 Ma) collected southwest of the San Andreas fault in the same mountainous region gave fission track ages of 61.2 and 67.3 Ma, suggesting a different thermal history on the two sides of the fault.

Theoretical Calculations. Anderson (1994) created an elaborate, comprehensive model for the part of the Santa Cruz Mountains that has been influenced by strike slip past the bend in the San Andreas fault near Loma Prieta. Using realistic input values, he derived a crustal thickening rate of about 0.7 mm/yr. This could approximately produce the uplift rate and would be in good agreement with rates based on various geological observations.

Likelihood of Ongoing Uplift. Most of the foregoing data do not prove that the Coast Ranges are still rising, although uplift of marine terraces as young as 105 000 yr is strongly suggestive that this is true. Results of releveling within the last few decades is qualitatively, if not quantitatively, persuasive. Occasional seismic activity on reverse faults in the subprovince shows that uplift continues, at least locally. In the near future, new geodetic methods will probably confirm or disprove the ongoing rise of individual ranges; in the meantime, we tentatively conclude that the ranges are rising at a rate of about 1 mm/yr. If their uplift began about 400 ka, as we think, the average rate since that date has been 0.76 mm/yr for the mountains that are now 305 m high, and 2.29 mm/yr for those that have attained 915 m.

POSTULATED CAUSE AND TECTONIC MECHANISMS OF THE RISE OF THE CENTRAL AND SOUTHERN COAST RANGES

Summary of Provisional Conclusions Regarding Origin of the Ranges

We conclude that late Neogene ancestral central Coast Ranges were created by horizontal contraction approximately normal to the plate boundary, and that this transverse shortening produced folding and thrusting above a zone of decoupling at or near the subhorizontal upper surface of relatively rigid lower crust, resulting in rootless uplift. After extensive erosion throughout the subprovince, the present ranges were created within the last 450 000 yr by a resurgence of the transverse contraction. These ideas are elaborated in the following. The data reviewed in this paper point to certain requirements and probable mechanisms for this mode of origin. However, it is apparent that our conclusions apply most securely to the ancestral phase of the Pliocene–Quaternary deformation in the province rather than the latest phase, which resulted in the present ranges. We make a few modifications and take some liberties in attempting to explain the present ranges, but we do not have final answers.

Basic Postulates

The observations and details discussed in the text, lead us to propose the following.

1. The central and southern Coast Ranges are the result of a component of compressive stress and resultant contraction normal to the plate boundary.

2. Preliminary folding and uplifts occurred in late Miocene and Pliocene time, but these uplifts were obliterated by erosion. The present ranges were delineated and began to rise during, or immediately following, a tectonic pulse at about 0.4 Ma.

3. Regionally, the present-day ranges rose largely in response to broad active thickening in the middle crust, in response to horizontal contraction.

4. Horizontal transverse shortening was accomplished by folding and thrusting in the upper crust and by ductile deformation in the middle crust. Both types of strain resulted in thickening.

5. The thickening was largely confined to the material above the relatively undeformed lower crust (seismic basement), which acted relatively rigidly. The individual ranges appear to be rootless.

6. The lower crustal basement and probably the entire subjacent lithosphere achieved or accommodated in various ways the same amount of horizontal shortening that resulted in thickening of the middle and upper crust. Shortening of the lower crust and lithospheric mantle probably entailed "pseudo-subduction" of the lower crust, which was decoupled from the middle and upper crust.

7. Some present-day ranges (or parts thereof) took advantage of preexisting strike-slip faults of the transform system, adopting these for high-angle slip at the uplift margins. Such faults may have served to delineate certain ranges at the inception of uplift.

8. The rise of some ranges was facilitated or enhanced by marginal thrust faulting.

9. The central and southern Coast Ranges, individually and as a subprovince, are subparallel with the plate boundary for a combination of reasons. (1) A change in plate motions at ca. 3.5 Ma caused a misfit between established transform faults and the azimuth of new relative motion between the Pacific and North American plates, resulting in a component of transverse contraction normal to the plate boundary. (2) The Franciscan accretionary subduction complex is widespread regionally in a broad band parallel with the plate boundary. Its weakness concentrated deformation in a corresponding belt. (3) The subprovince originated in the most active part of the transform zone, which had numerous strike-slip faults. These faults have low frictional strength, causing the regional maximum compressive stress to be deflected toward a direction normal to the faults.

Concept of Bodily Uplift of Present-Day Ranges with or Without Internal Folding

The commonplace folds and faults seen at the surface reveal the predominant tectonic mechanisms at shallow levels. However, only certain faults seem to have played a role in the uplift of the present ranges. There must have been times during the late Cenozoic when growing folds influenced the developing topography. Anticlines were topographically high, although modified by contemporary erosion, and synclinal areas were low. We see this today in the southwest part of the Great Valley province, but we do not see it in typical parts of the Coast Ranges. Instead, each range has recently been uplifted more or less bodily, usually carrying within itself a series of internal folds (Figs. 9 and 10). Reverse-slip focal mechanisms of some (albeit relatively few) earthquakes show that transverse contraction and consequent uplift of the ranges continues. Marginal thrust faulting locally plays an obvious part in uplift. Despite the direction of relative plate movements, the maximum horizontal stress is nearly perpendicular to the San Andreas and other major strike-slip faults because of the remarkably low friction on these faults.

Each of the modern ranges has risen en masse, carrying its passive internal architecture. This is well illustrated by the small San Luis Range between Morro Bay and San Luis Obispo. The Pismo syncline is the dominant internal structure (see left side, Fig. 10.) Although the Pismo syncline incorporates late Pliocene strata and is therefore a very young fold, it shows no direct influence on the mountains that contain it. A survey conducted by the Pacific Gas and Electric Company showed that the marine terraces that cross the range and the fold have neither been downwarped nor arched, but have been uplifted during the continuing rise of the range. The terraces, still essentially level, are offset by late Quaternary reverse faults, which bound the range and dip beneath it (Pacific Gas and Electric Company, 1988).

We infer that the San Luis Range and probably all the Coast Ranges were pushed upward by thickening of the more ductile crustal material below in response to horizontal contraction.

It is instructive to examine the ranges that incorporate Salinian granitic and metamorphic rocks. In some areas, as in the northern Santa Lucia Range and northern Santa Cruz Mountains, these formerly strong rocks have been thoroughly sheared and locally crushed; they have participated in the folding and reverse faulting of overlying strata (e.g., Compton, 1966). Thus, it is not surprising that their enclosing mountains have behaved the same way as ranges with folded strata overlying the soft Franciscan Complex. It is surprising that some (but not all) of the areas where Salinian granite has defied recent deformation have also been uplifted, much like the other Coast Ranges. Examples are the northern Gabilan Range and the Santa Cruz Mountains near the city of Santa Cruz. On the other hand, the Gabilan Mesa, where granite and overlying young strata are little deformed (see Fig. 10), is only slightly elevated. The behavior of the Gabilan Range, like the ranges that have uplifted internal folds, supports the idea that all of the central Coast Ranges have been affected by a range-wide upward push that largely disregards near-surface materials and structures. This action has been supplemented or accompanied locally by near-surface marginal thrusting.

Conversion of Transverse Shortening to Vertical Uplift

How is horizontal contractional strain converted to vertical uplift in the Coast Ranges? The behavior of the material must vary with depth, among other factors. Depth largely dictates temperature and confining pressure; these in turn affect fracturing, the behavior of fluids, and the gross rheology of rocks. We review obvious ways in which horizontal contraction may cause deformation at depth, and consequent uplift of overlying material.

Thrust Faulting. Faulting commonly occurs to depths of 10–18 km in the Coast Ranges, as shown by the distribution of earthquake hypocenters. Most of the focal mechanisms denote strike slip, but many microearthquakes and a few sizeable events, for example, the 1983 Coalinga earthquake (Rymer and Ellsworth, 1990), are ascribed to reverse and/or thrust faults. Beneath part of the San Francisco Peninsula west of Palo Alto, microearthquakes originating on both sides of the San Andreas fault commonly show reverse and/or thrust focal mechanisms 3–10 km below the surface, and their focal planes tend to strike subparallel with the San Andreas fault (e.g., Olson and Zoback, 1995). These clues suggest that thrust and reverse faulting within common seismogenic depths is one of the mechanisms of ongoing mountain building, as implied by others (e.g., Crouch et al., 1984; Namson and Davis, 1988; McLaughlin et al., 1991; Jones et al., 1994).

Folding. Figures 9 and 10 show that folding, as well as thrusting, must have created ancestral mountains. Although these mountains no longer exist, the deformation thickened the upper crust, and this probably had a lasting effect. In the middle crust, folding of a different style may have occurred, but where the Franciscan Complex is predominant, it more likely deformed by pervasive ductile shortening and thickening. Overall, folding played only a small role in thickening of the crust during uplift of the present ranges, which postdate the erosional truncation of anticlines and synclines, but these folds may have been tightened in ways that contributed to thickening and uplift.

Ductile Thickening. Ductile shortening and thickening probably are generally more important than thrusting in the middle parts of the crust and even in some of the upper crust, where pervasive fracturing occurs. The mode of deformation must be dependent upon rock type, temperature, fluid pressure, strain rate, and perhaps other factors. We use the term “ductile” in a nonprecise manner to imply differential mobility of solid material, including penetrative slip on a myriad of small fractures producing grossly homogeneous strain. The rocks that behave thus are hard and brittle when subjected to suddenly applied stress, and they transmit seismic shear waves. Ductile behavior is to be expected in Franciscan melanges, which have argillaceous matrices and lack continuous strong layers. Evidence of ductile behavior is seen in melanges, even at the surface, and must be endemic at depths wherever the Franciscan Complex prevails vertically and laterally beneath the ranges. In layered rocks, ductile strain may include flow folding.

As mentioned above, ductility would be enhanced by the presence of fluids, elevated temperatures, and high fluid pressure. The possible tectonic role of high fluid pressure in the upper crust of western California was discussed by Berry (1973), and that in the Coalinga area was discussed by Eberhart-Phillips (1986). Although Berry had few measurements in the Coast Ranges, as distinguished from the Great Valley and oil-bearing anticlines along the southwest margin of the valley, the available data suggested to him that large volumes of Franciscan rocks, perhaps most of these rocks, have anomalously high, near-lithostatic fluid pressure. He thought that this condition prevailed downward to the base of the Franciscan (the base is shown at about 7 km in Berry, 1973, Fig. 5, northern Coast Ranges), and he emphasized that high fluid pressure promotes thrusting, overall ductile behavior, and diapirism.

It is possible that high fluid pressure and ductile behavior enabled Franciscan rocks to form laterally intrusive wedges of the sort envisioned by Wentworth and Zoback (1990). These thick wedges of Franciscan rocks (Figs. 9 and 10) formed along the east side of the Coast Ranges province. Tapering to the northeast, they were inserted between a hanging wall of the Great Valley sequence and a footwall of Great Valley basement. Perhaps their constituent material moved more or less laterally out from a semiplastic Franciscan mass, the mobile body following a thin leading edge. The emplacement would necessarily raise the overlying shallow crustal material. The weak Franciscan rocks may have been squeezed and thickened overall, thus raising the topographic surface and causing the compressed rocks to partially escape more or less in the manner of putty. Alternatively, the Franciscan may have been thrust as a brittle material over the Great Valley basement in intermittent increments, and the Great Valley sequence of sedimentary strata may have been thrust over the Franciscan in alternating complementary events. In any case, progressive wedging broadened, and probably heightened, the Diablo Range.

Perhaps more important, albeit less obvious, the uplift of the rigid Salinian rocks of the Gabilan Range would seem to require the pervasive thickening of a large mass beneath the block-like granitic and metamorphic rocks exposed at the surface. It is unlikely that Franciscan Complex rocks are present beneath the Salinian granite. What plastically behaving material could underlie the range? Seismic refraction (Walter and Mooney, 1982) and reflection (Lynn et al., 1981) data indicate that the granitic rocks probably extend downward to a depth of only about 9–10 km. Below that depth, the strong subhorizontal reflectivity and higher P-wave velocity (about 6.4 km/s) may be compatible with the schist of the Sierra de Salinas, which Ross (1976) reported from the Santa Lucia Range and from the subsurface near the flanks of the Gabilan Range. Ross suggested that it may correlate with the Pelona Schist of southern California (Ross, 1976). Presumably, the schist is weaker than granitic rocks and might tend to behave more ductilely. However, its lateral and vertical extent are entirely unknown.

Shear Along Low-Angle Zones of Decoupling, Including Detachments. Seismic profiles indicate that the surfaces of various lower crustal basement rock at depths of 14–22 km are little deformed compared with rocks in the upper crust; within the limits of resolution of present-day techniques, some such surfaces do not seem to have been deformed at all. Detachments have been invoked by Case (1963), Lachenbruch and Sass (1980), Zandt and Furlong (1982), Crouch et al. (1984), Namson and Davis (1988), McLaughlin et al. (1991), Brocher et al. (1994), Jones et al. (1994), and others. Probably not all of the detachment surfaces that have been proposed are real, and some may be subhorizontal ductile shear zones of appreciable thickness rather than discrete, flat fault-like features. However, the fact that there are multiple plausible reasons for suspecting their presence, including their proposed role in heat generation, the finding of little-deformed subhorizontal surfaces in seismic profiles, and the geometric impossibility of projecting shallow folding into basement, strengthens the likelihood that some such zones of decoupling actually exist.

The likely presence of low-angle zones of decoupling might seem to have little bearing on uplift of ranges. However, differential movement along these zones could facilitate horizontal shortening of overlying midcrustal material, thickening of the crust, and hence uplift at the surface. Above the brittle-ductile transition zone, differential movement could occur along a detachment in the manner of fault-like slip. Below the transition, it would be ductile shear in a less-definitive zone of decoupling.

Beneath the eastern part of the central and southern Coast Ranges, the upper surface of the Great Valley basement probably serves as a locus of decoupling in the same way as the upper surface of the mafic (probably oceanic) lower crustal basement to the west. This is particularly likely at the base of the tectonic wedges postulated by Wentworth et al. (1984), Sowers et al. (1992), Unruh and Moores (1992), and Ramirez (1994) along the eastern margin of the Coast Ranges province. As already noted herein, at least some of the wedges apparently consist of Franciscan rocks that have been inserted between the basement and overlying strata of the Great Valley sequence. Although ductile behavior is manifest, discrete slippage probably occurred at the base of the wedges and perhaps also at the base of the Franciscan masses that gave birth to the wedges.

How did the lower crust accommodate the transverse shortening that we have emphasized? The shortening that is recorded by folds and thrusts in the upper crust must have affected the entire lithosphere but was expressed in different ways at various depths. The upper surface of the mafic (probably oceanic) lower crust generally seems to show little or no deformation. An exception is shown in the southwest part of the transect of Figure 10, where Putzig (1988, as reproduced in Miller et al., 1992) postulated incipient imbrication and where we have shown an offset caused by the San Gregorio–Hosgri fault. Elsewhere, the lower crust may have shortened and thickened without obvious offset, or it may have underthrust a neighboring sector of lower crust and mantle. The thickness of most of the southwestern mafic lower crust in Figure 10 does not seem abnormal for oceanic crust, so we suppose that this slab-like layer has avoided buckling, imbrication, and/or static thickening by descending into the mantle beneath an adjacent crustal sector, as postulated for the San Francisco Bay region by Brocher et al. (1994). We have indicated this both in Figure 9 and in Figure 10, but the representation is entirely speculative. In Figure 10, the site of “pseudo-subduction” was placed at the juncture between oceanic crust on the southwest and Great Valley basement. The upper surface of the western oceanic crust is at a different depth and the rocks have somewhat different seismic velocities from the Great Valley basement on the opposite side of the San Andreas fault. Hence the two adjacent segments of lower crust are truly separate. In Figure 9, the locus of “pseudo-subduction” was placed at the downward projection of the Hayward fault, because the presence of oceanic crust is well established westward from the fault and the Great Valley magnetic basement is believed to extend westward from its eastern boundary at least to the Hayward fault (Jachens et al., 1995).

Crustal Prism of Deformed Rocks Above Low-Angle Loci of Decoupling. Thus, it appears that at least the marginal parts of the Coast Ranges province are underlain by detachments and loci of plastic shear. Above these subhorizontal features, shortening in the upper part of the crust was achieved by folding and thrusting that did not involve the crust below. The transects of Figures 9 and 10 show a prism-like or lens-like body of folded and thrust-faulted upper crustal rocks underlying the Coast Ranges province. The spatial relations strongly suggest a genetic link between the deformed prism and the Coast Ranges, but the prism probably formed largely during Miocene time and at the time of the Pliocene uplifts that immediately preceded the rise of the present Coast Ranges. Perhaps the tardy rise of the present ranges in the past 400 000 yr involved a resurgent pulse in the shortening and overall thickening of the deformed prism and involved shear in low-angle zones of decoupling.

Erosion and Mass Wasting vis-a-vis Uplift. Role of Buoyancy

Geologic cross sections show deep erosional truncation of folds, and topography and the weakness of prevalent rock types suggest the removal of a huge volume of material at geologically rapid rates. In estimating the amount or rate of uplift of the mass of rock underlying an elevated surface, one should add to the present surface the amount of material (measured vertically) that has been eroded away during a specified time interval (Molnar and England, 1990).

The most prevalent Coast Ranges rocks, except for some of the Salinian granite and massive units of the Great Valley sequence, are typically weak, either because of poor lithification or because of tectonic fracturing and shearing. The extent of erosion since the beginning of Pliocene–Pleistocene folding and thrusting is dramatic (Figs. 8–10). Probably 1–5 km of rock (measured vertically) was removed from most parts of the subprovince since 3.5 Ma, leaving the folds and internal faults drastically truncated. Most of this erosion preceded the delineation and uplift of the present ranges. Its effect on the origin of the latter is problematic, but may somehow have been significant.

Judging from remnants of old erosion surfaces that are preserved in summit areas of the central and southern Coast Ranges, the present ranges started to rise at a time of regional subdued topographic relief (Figs. 3, 14, and 15). During the geologically rapid uplift, the subdued topographic surface was sharply incised by intermittent streams with steep gradients. This process continues today, aided by hillside creep, landsliding, and removal of material by debris flows. Many of the ranges are bordered by large alluvial fans, the depositional products of erosion and mass wasting of the rising mountains.

Montgomery (1993) reviewed published measured and estimated contemporary erosion rates in the central Coast Ranges and determined a weighted average of 0.08 mm/yr. On the basis of likely differences in conditions in Pleistocene time, when most of the present topography was formed, he concluded that the Quaternary rate was probably 0.05–0.10 mm/yr, and he used a rate of 0.05 mm/yr for a period of 3.5 m.y. in his model of uplift.

The Santa Cruz Mountains in particular have been studied analytically with respect to the interrelationship of geomorphic and tectonic processes. Anderson (1994) devised a mathematical model that incorporates fluviatile erosion (plus landsliding) at observed rates, and uplift that he ascribed largely to horizontal compression resulting from a slight left bend in the dextral San Andreas fault, the fault being located well within the range in that particular area (around Loma Prieta). He presented equations for key geomorphic processes and for the mass displaced by crustal movement past the fault bend. Dextral strike-slip along the fault causes crustal thickening and uplift. The model approximates many of the observed features of the Santa Cruz Mountains, but does not explain uplift of other parts of the Santa Cruz Mountains and the other Coast Ranges, which generally lack restraining bends in strike-slip faults. However, the Santa Cruz Mountains are wider and higher in the area of the fault bend than elsewhere, so the bend evidently supplements other processes.

In their examination of uplift of the Santa Cruz Mountains, Burgmann et al. (1994) also considered the amount of erosion. They studied the same part of the range (around Loma Prieta) as Anderson. Their geomorphic study supported the conclusion from fission tracks that the northeast side of the range has risen geologically rapidly, and that as much as 2–3 km of unroofing has occurred in the past 10 m.y. Burgmann et al. used the unroofing inferences to estimate uplift rates, but did not discuss possible tectonic effects of the removal of overburden.

One might suspect that erosion must help to sustain and perpetuate uplift; however, this may not be a prime factor for individual ranges. For example, Figure 7b of Burgmann et al. (1994) does not show any marked difference in the elevation of generalized high areas in the two parts of the range with different uplift and erosion rates. Gravity maps do not indicate a large role for buoyancy in domains coinciding with the individual ranges. We conclude that the impressive volume of rock eroded from the Coast Ranges was probably not a determining factor in the rate or amount of uplift of individual ranges, although the subprovince as a whole appears to be in near-isostatic equilibrium.

Origin of Structural Valleys

The large structural valleys in the central and southern Coast Ranges subprovince must be considered in conjunction with the ranges, as the origin of the two entities must be related. The principal valleys are the San Francisco Bay–Santa Clara Valley depression, Salinas Valley, Cuyama Valley, and the Carrizo Plain.

Apparently, the structural valleys, like the ranges, were produced by transverse contraction. There is no indication that they are grabens, pull-apart basins, or erosional features; the evidence is to the contrary. Wherever bordering faults are found, these are thrusts, high-angle reverse faults, or strike-slip faults of the transform system. Figure 16 shows a cross section of Cuyama Valley (modified from Vedder and Repenning, 1975) where the structure has been well determined with the help of oil-well drilling. Neogene strata are asymmetrically synclinal and are overthrust along both margins. Pliocene rocks are involved, indicating that the time of deformation was penecontemporaneous with the southern Coast Ranges mountain building.

San Francisco Bay, which occupies a northwestward extension of Santa Clara Valley, is not a sedimentary basin, although it contains at least one large pocket of sedimentary rock. For the most part, it is underlain at a shallow level by the Mesozoic Franciscan Complex. This has been reached by drillholes 30–350 m deep near the edge of the bay proper, and it crops out on islands and nearby hills. Much of the bay block appears to be structurally high except for the presence of very young Quaternary deposits (Fig. 9). Likewise, at least part of the Salinas Valley is underlain by Mesozoic rocks at a shallow level. Oil-well drilling at the San Ardo oil field between King City and Paso Robles discovered granite beneath thin sedimentary rocks and defined a thrust fault bounding the southwest side of the valley and dipping beneath the adjoining Santa Lucia Range.

Apparently, some structural valleys, including Cuyama Valley, were forced down with the aid of bordering outward-dipping thrust faults, while the adjacent ranges were forced up. In other cases, the valley blocks may have remained at a neutral elevation during the rise of the ranges. The possibility of an erosional origin is entirely untenable, as most of the valley margins are blanketed by alluvial fans, which indicate aggradation rather than the opposite. Part of the Santa Clara Valley drains to the northwest and part to the southeast, so the valley could not have been excavated by a prehistoric through-going river.

CAUSES AND PRESENT STATUS OF TRANSVERSE CONTRACTION

We will not attempt a comprehensive discussion of the cause(s) of contraction normal to the transform boundary in California, but summarize some published data and propose some ideas. Evidently, oblique convergence in late Miocene time produced folding and thrusting in parts of western California during the subduction regime, but as we have shown, the principal chain of tectonic events leading to the evolution of the Coast Ranges occurred in Pliocene and (more drastically) Quaternary time, during the ongoing transform period.

Change in Plate Motions, ca. 3.5 Ma

Geological evidence (Quaternary folds, thrusts, and uplifts quasiparallel with the plate boundary) for transverse compression had been noted for some years when Minster and Jordan (1984) made a quantitative comparison of the direction and rate of slip on the San Andreas fault with the Pacific–North America plate motion and calculated the rate of shortening normal to the fault. They used the RM2 pole of rotation, which they had established earlier (Minster and Jordan, 1978), and derived a transverse shortening of 4–13 mm/yr west of the San Andreas fault. They estimated a lesser rate of shortening east of the San Andreas fault, based mainly on certain parameters for extension in the Basin and Range province.

Very likely, most of the transverse shortening studied by Minster and Jordan began with the well-established Pliocene change in plate motions (Cox and Engebretson, 1985) that Harbert and Cox (1989) dated at between 3.40 Ma and 3.86 Ma. In this paper, for convenience, we have ignored the uncertainty in the date and have arbitrarily adopted the date 3.5 Ma; there is no supporting scientific evidence that this is more credible than any other number between 3.4 and 3.9. Harbert and Cox (1989) calculated a clockwise change of 11° in the azimuth of relative motion between the Pacific and North American plates (as interpreted in Fig. 17), and cited much geologic evidence for compressional tectonics in California after the change.

De Mets et al. (1990) refined the new (the past 3 m.y.) plate motions by using a global circuit improved by a previously unequaled abundance of refined data, and established a new global model (NUVEL-1) that provided updated relative plate motions. They derived a new current pole of rotation for the Pacific and North America plates and found a relative plate motion of 48 ± 1 mm/yr along an azimuth of N36°W at lat 36°N on the San Andreas fault. The rate of boundary-parallel plate motion was determined from spreading near the mouth of the Gulf of California. The new data gave a rate of 7 mm/yr of shortening normal to the plate boundary at lat 36°N across the San Andreas fault, where the azimuth of plate motion differs by 5° from the trend of the fault.

The Pliocene change in plate motions was reanalyzed by Cande et al. (1992), who concluded that the event probably occurred at about 6 to 5 Ma. However, this does not coincide as well as 3.5 Ma with the general uplift of coastal California recorded by deposition of nonmarine Pliocene–Pleistocene sediments. A recent detailed discussion of late Cenozoic interactions of plates and microplates in the region of the southern Coast Ranges, with geologic effects in the Santa Maria area, was given by McCrory et al. (1995).

A change in the azimuth of relative plate motion need not have produced compression if new transform faults had been quickly formed, having the same azimuth as the new relative motion. However, this did not occur, at least on a major scale. The same prechange San Andreas fault persisted, as shown by the lack of any abandoned ancestral faults of appropriate size. (The San Gabriel fault, a former segment of the San Andreas fault, was superseded, but this event was caused by the rapid creation of the Transverse Ranges across the fault system.) To this day, the San Andreas fault is a misfit, and slippage along it must be accompanied by transverse compression. New faults are developing (the Hayward–Rogers Creek, Calaveras, Maacama, Concord–Green Valley, Greenville, Bartlett Springs, and others), mainly oriented more northerly than the San Andreas, and these will tend to relieve the compressional stress.

We can estimate the maximum amount of transverse shortening explainable by the misfit due to the change in plate motions if we make simplifying assumptions such as the following. Let us accept the Harbert and Cox (1989) estimate of 11° of clockwise change in azimuth of the relative plate motion; assume that this happened instantaneously at 3.5 Ma; presume that all the transform faults doggedly retained their previous strike, and no new faults developed (although in reality some did form); and assume a constant relative plate motion of 48 mm/yr parallel with the plate boundary. Then, during the 3.5 m.y. that has elapsed to the present, a maximum of about 31 km of transverse shortening would have resulted from this particular event. The 31 km, which admittedly depends upon the most favorable assumptions, compares with 33 km of shortening that we deduced by unraveling folds and thrusts and by making interpolations throughout the width of the central Coast Ranges (Table 4). The 31 or 33 km of shortening could have been accommodated in two ways. (1) It may have been absorbed entirely by folding and thrusting within the site of the Coast Ranges. Note that the Pliocene and Quaternary deposits on either side of the province are deformed mildly or not at all, in contrast to their deformation within the ranges. (2) Some of the shortening may have been accommodated by thrusting of the continental margin oceanward over the Pacific plate, as proposed by Page and Brocher (1993). This idea has not been proved or disproved. We do not know how (or if) the strain was shared.

Probable Change in Plate Motions Since 0.5 Ma

Why did the present-day ranges begin to rise only as recently as 0.4 Ma? This event would seem to signal a recent marked resumption or surge in the transverse contraction that commenced about 3.5 Ma. We propose that this surge was caused by an accentuation of the ongoing change in plate motions that commenced in Pliocene time. This possibility cannot be examined in detail in this paper; however, note the marked (perhaps 15°) clockwise change in the trend of the hotspot track of the Hawaiian Islands since the time of active volcanism at Haleakala (ca. 0.75 Ma) on the Island of Maui (e.g., see Fig. 1.14 of Clague and Dalrymple, 1987). Inasmuch as the Pacific plate had been moving northwest and the North American plate (much more slowly) southwest, in the hotspot frame of reference, a clockwise change in the path of the Pacific plate would have increased the slight, but important, component of convergence that began in Pliocene time.

Relation to Basin and Range Extension

It has been suggested that extension in the Basin and Range province may have contributed to deformation in California (e.g., Wright, 1976), and this remains a possibility. However, as shown by Minster and Jordan (1984), the magnitude, direction, and timing of extension are probably not adequate to explain the contraction normal to the plate boundary. It is generally agreed that the Basin and Range province has extended tens or hundreds of kilometers since Oligocene time; for example, Wernicke et al. (1988) estimated 140 km of extension between the southern Sierra Nevada and the Colorado Plateau. However, Zoback et al. (1981) concluded that since 10 Ma, the prevailing direction of extension has been northwest-southeast. Abundant geologic data (e.g., Thompson and Burke, 1973) show that much of the overall extension has been approximately N60°W, as shown in Figure 17; this is about 13° counterclockwise from the N47°W azimuth of the relative plate motion prior to 3.5 Ma and 24° from the present azimuth, at lat 36°N on the San Andreas fault.

It seems that some parts of the Basin and Range province behave somewhat differently from other parts (Dixon et al., 1995). Apparently, this happened in the past when variously oriented ranges came into being in different subdomains; note the trends of range-bounding faults in Figure 17. At present, east-west extension is occurring in the eastern part of the province, but its effect in the west is transposed by north-northwest shear (Dixon et al., 1995). Bellier and Zoback (1995) found strong evidence for Quaternary N85°W extension in the Walker Lane zone of western Nevada and eastern California. If some areas at times have undergone extension that is more nearly east-west than the average prevailing direction, and if at the same time northwest-southeast dextral strike slip has lagged in the province, the consequences may have been temporarily significant. Over the long term, local deviant behavior in parts of the Basin and Range province probably either serve to normalize the overall strain in the province as a whole or is sooner or later countered by a different strain, so that the province abides by the requirements of the plate motion. (See following discussion, Coast Ranges Considered in Broad Tectonic Context. Measurements of present-day Basin and Range movements are reported in the section Present-Day Transverse Shortening.

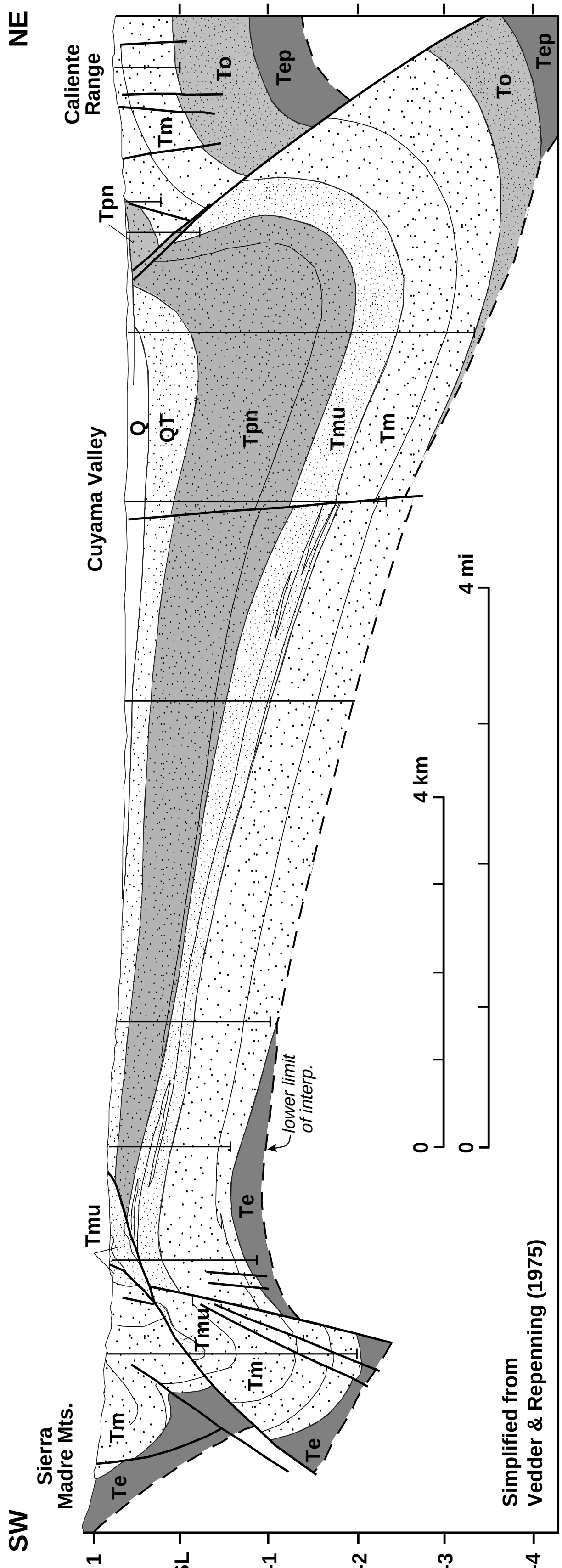

Figure 16. Cross section across Cuyama Valley, modified from Vedder and Repenning (1975). See Figure 1 for location of the valley. Contractional features, largely confirmed by oil wells, clearly indicate progressive compressional origin, culminating during Quaternary time.

Compressional Effects of Bends in Strike-Slip Faults

Some of the large strike-slip faults (i.e., the San Andreas and Calaveras faults) in the transform zone are not straight. A counterclockwise bend (looking northwest) must increase the normal stress on these dextral faults as they attempt to accommodate the relative plate motion. As discussed in a preceding section, this seems to apply to a part of the Santa Cruz Mountains, where a small but marked bend occurs in the surface trace of the San Andreas fault within the range. The mountains surrounding the bend are broader and higher than the northwesterly continuation of the range, and thrust faults that are quasiparallel with the plate boundary are conspicuous. Thus, there is reason to recognize the contribution of such bends. However, most parts of most ranges do not exhibit such circumstances, so the more general existence of transverse compression must be explained in other ways, such as the 3.5 Ma change in plate motions and perhaps the influence of Basin and Range extension.

Present-Day Transverse Shortening

Some geodetic surveys have not detected any present-day shortening normal to the plate boundary. However, analysis by Harris and Segall (1987) of trilateration from the San Andreas fault to the coast near San Luis Obispo indicated transverse shortening of 6.1 ± 1.7 mm/yr, and Feigl et al. (1990), using geodetic data across the Santa Maria region, inferred shortening of 5 ± 2 mm/yr normal to the San Andreas fault. New methods tend to show a small amount of such strain. Argus and Gordon (1991), using data from very long baseline interferometry, concluded that the motion of the Sierra Nevada–Great Valley microplate relative to the interior of North America, in addition to slip on the San Andreas fault, collectively account for most of the Pacific–North America relative plate motion. At lat 36°N on the San Andreas fault, the Argus and Gordon data indicate only about 2 mm/yr of transverse shortening, and 5 mm/yr is the maximum allowed. Recent Global Positioning System measurements spanning the principal part of the transform boundary between the Farallon Islands west of San Francisco and the town of Columbia in the Sierra Nevada Foothills show a transverse shortening of 3 mm/yr (W. H. Prescott, 1996, personal commun.).

Judging from the foregoing, the rate of present-day transverse convergence is small, approximately 1–5 mm/yr, as compared with our estimated 9 mm/yr average for the past 3.5 m.y. This small convergence may be reasonable in view of the fact that drastic folding and thrusting do not seem to have occurred since the present ranges began to rise in the past 0.4 m.y. The recent uplifts appear to have a compressional cause, but most of the boundary faults are high-angle features. As discussed previously herein, the ranges most likely rose because of shortening and thickening of material at depth. The amount of implicit shortening at depth is difficult to assess because it depends on the thickness of the compressed zone, which is unknown. However, Figures 10 and 11 suggest an average of about 10–15 km between the upper crustal folds and faults (which did not play a part in the recent uplifts) and the little-deformed lower crust. This middle crust was most likely the locus of the material that mechanically thickened and thereby raised superincumbent ranges. For simplicity, let us assume that the zone of thickening is everywhere 12 km thick and 114 km wide. If the entire zone shortened at a rate of 3 mm/yr and if the thickening were somehow concentrated beneath ranges occupying two-thirds of the surface area, these ranges would only rise about 0.5 mm/yr. Our best guess is that the actual rate is around 1 mm/yr, but the exact figure is unknown. In the 400 000 yr that the present ranges have been rising, the average rate has been 0.7 to 2.3 mm/yr.

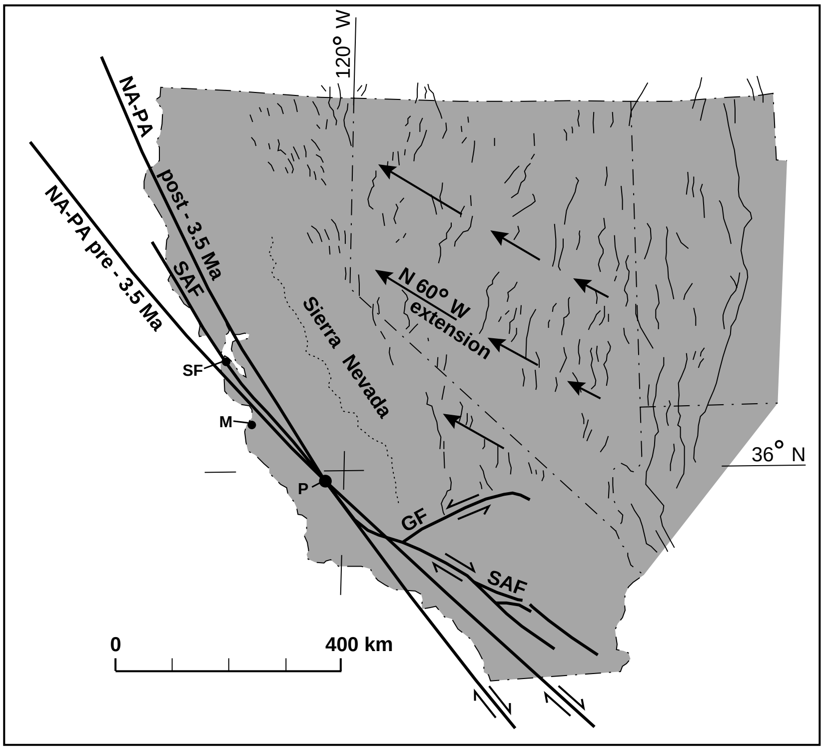

Figure 17. Diagrammatic map showing change in plate motions at ca. 3.5 Ma, vis-a-vis orientation of San Andreas fault (from Page and Brocher, 1993). A convergent component of motion transverse to the plate boundary is implied. Prevailing direction of extension in Basin and Range province is shown to suggest limited possible effect on Coast Ranges.

COAST RANGES CONSIDERED IN BROAD TECTONIC CONTEXT

In regional context, the Coast Ranges are part of the broad, currently deforming boundary between the Pacific plate and the continental interior, a boundary that also includes the Great Valley Sierra Nevada microplate, and the Basin and Range province. Can the bulk Quaternary strain, so different from subregion to subregion, be reconciled with the plate motion, i.e., the Pacific plate moving about 48 mm/yr northwest relative to North America?

Conceptually, one can think of the overall strain in this broad region as a principal west-northwest extension and north-northeast contraction, not necessarily equal. (The principal directions of the strain tensor are directions of no shear.) The San Andreas dextral fault and the Garlock sinistral fault are close to directions of maximum shear, and the north-northeast–trending normal faults of the northern Basin and Range province (Fig. 17) are normal to a direction of pure extension. Structures having trends intermediate between principal strain directions and the maximum shear directions undergo transtension or transpression. For example, the

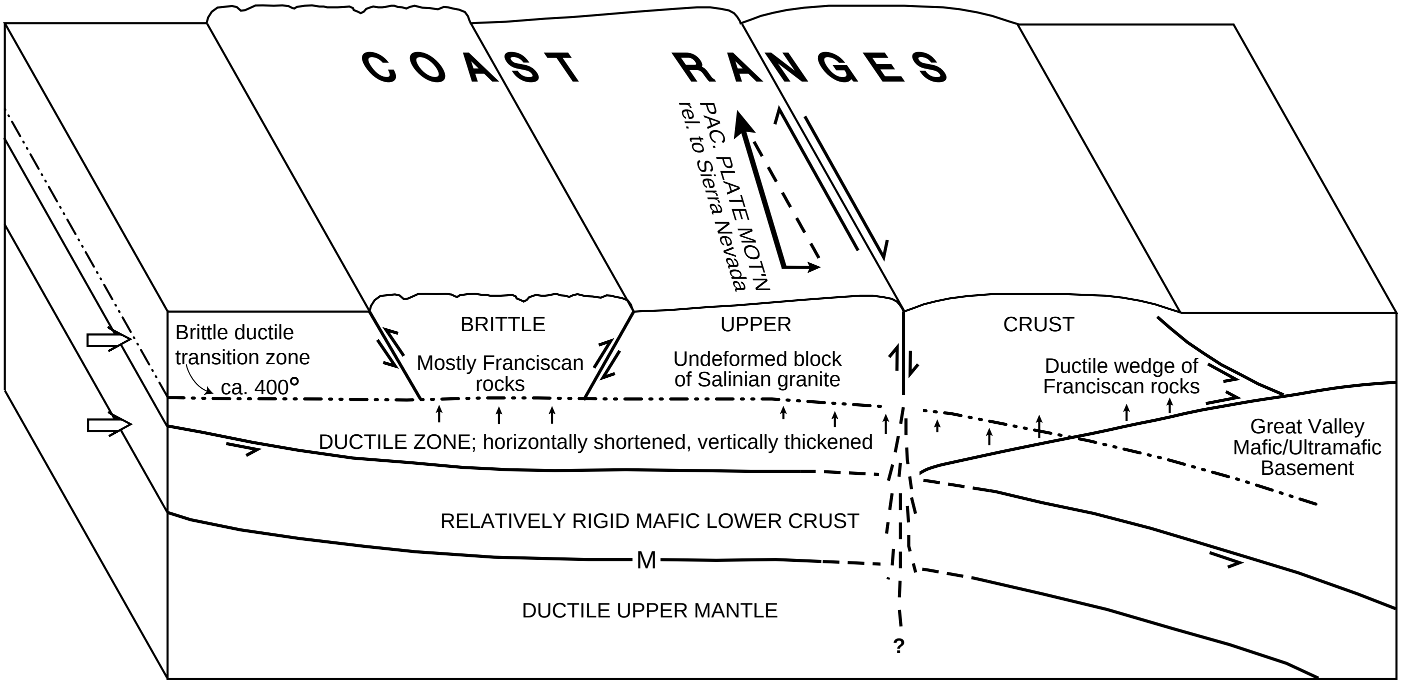

Figure 18. Synoptic diagram showing overall rheology and postulated modes of late Quaternary uplift of present-day central and southern Coast Ranges. Portrayal is influenced by Figure 36 of Eaton (1996).

eastern margin of the Sierra Nevada (Owens Valley and Walker Lane), which has a north-northwest direction, is deforming in dextral transtension, and the central and southern Coast Ranges and San Andreas fault, trending slightly counterclockwise of the Pacific–North America vector (Fig. 17), are deforming in dextral transpression. Thus, the thrusting and folding of the central and southern Coast Ranges are in harmony with the same framework as Basin and Range extension, provided that the directions and response of the yielding structures are predetermined by other factors such as material properties.

The central and southern Coast Ranges are composed largely of the Franciscan subduction complex, generally consisting of materials collectively so weak and voluminous that one would expect later deformation and consequent uplift to be localized in it. In contrast to the Coast Ranges, the Great Valley basement and the Sierra Nevada are strong and rigid, having been thoroughly consolidated by batholithic intrusion and accompanying metamorphism. Unlike both the Great Valley–Sierra Nevada microplate and the Coast Ranges, the northern Basin and Range province was subjected to an enormous late Cenozoic influx of heat and mass supplied by the mantle plume head that initiated the Yellowstone hotspot, the Columbia River flood basalts, and the Northern Nevada rift about 17 Ma (Parsons et al., 1994). The very different tectonic responses of these subregions and their boundaries within the broadly deforming plate boundary are thus explicable in terms of their earlier preparation.

CONCLUSIONS

We have emphasized the youthfulness of the central and southern Coast Ranges and the fact that they developed in response to a component of convergence across the plate boundary, largely in Quaternary time. As summarized in Figure 18, the ranges are entirely within the broad transform zone, and the mountain building was concurrent with dextral strike-slip faulting.

Although compression was marked along the coastal region in late Miocene time owing to oblique convergence during the latter part of the subduction regime, this apparently did not affect the entire subprovince and did not continue into the main Pliocene–Quaternary orogenic development. Rather, the evolutionary development that has persisted almost (but not quite) continuously to the present, began in Pliocene time at the time of change in relative motion between the Pacific and North America plates about 3.5 Ma, when most of western California rose above sea level. Marine deposits having lengthy time spans (more than 150 m.y.) were intensely folded and thrust faulted as the transverse convergence continued. An important component of the deforming deposits was the Franciscan subduction-accretionary assemblage, which was already locally chaotic, markedly sheared, and charged with fluids. This complex was easily susceptible to further deformation, including overall ductile behavior. The progressive Pliocene–Quaternary convergence produced a thickened prism of highly deformed, weakened, middle and upper crustal material above a little deformed lower crust. The lower crust may have accommodated the transverse shortening by a continental subduction-like escape; however, this idea is entirely speculative. Meanwhile, the thickened prism was further weakened by high fluid pressure and anomalously high temperatures, perhaps in part self generated by shearing. This facilitated further deformation.

Although these geologically recent events must have produced mountains and set the stage for the creation of the present-day central and southern Coast Ranges, to the south these ranges show little or no direct effects of the Pliocene–Quaternary folding and internal thrusting. The modern ranges are confined to the geographic realm of the folding and crustal thickening, but they rose after the deformation and after deep, widespread erosion. The present ranges began to appear about 0.4 Ma, and rose partly as coherent, steeply bounded fault blocks, partly as blocks locally bounded by thrust faults, and partly as vaguely bounded uplifts lacking significant marginal faults. Most ranges exhibit all of these modes of uplift, in various localities. We believe that they were lifted by the lateral compression and vertical thickening of midcrustal material, notably the Franciscan Complex. The marked change from intense internal deformation to coherent uplift of entire ranges may have resulted from progressive dewatering and consolidation of the relatively soft materials of the upper and middle crust. Some of the uplifts were guided by preexisting northwest-southeast–trending strike-slip faults, which were utilized as high-angle slip surfaces.

Some of the foregoing hypotheses may not be correct, and even if all of them are tenable, formidable problems remain. Exactly how were the present-day ranges delineated where preexisting faults are not found? Why were parts, instead of the entire subprovince, uplifted? What determined which tracts would become mountain ranges and which would be depressed, or remain neutral, as structural valleys? If, as seems likely, transverse convergence was more or less continuous after about 3.5 Ma, why did Pliocene–Pleistocene folding and thrusting pause for widespread erosion, and why was this deformation soon succeeded by numerous, separate vertical uplifts with minimal additional deformation?

We are impressed by the geologically recent tectonic events in west-central California, and encourage others to test the ideas herein and to find solutions to the outstanding major problems.

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Figure 9. Transect across Santa C
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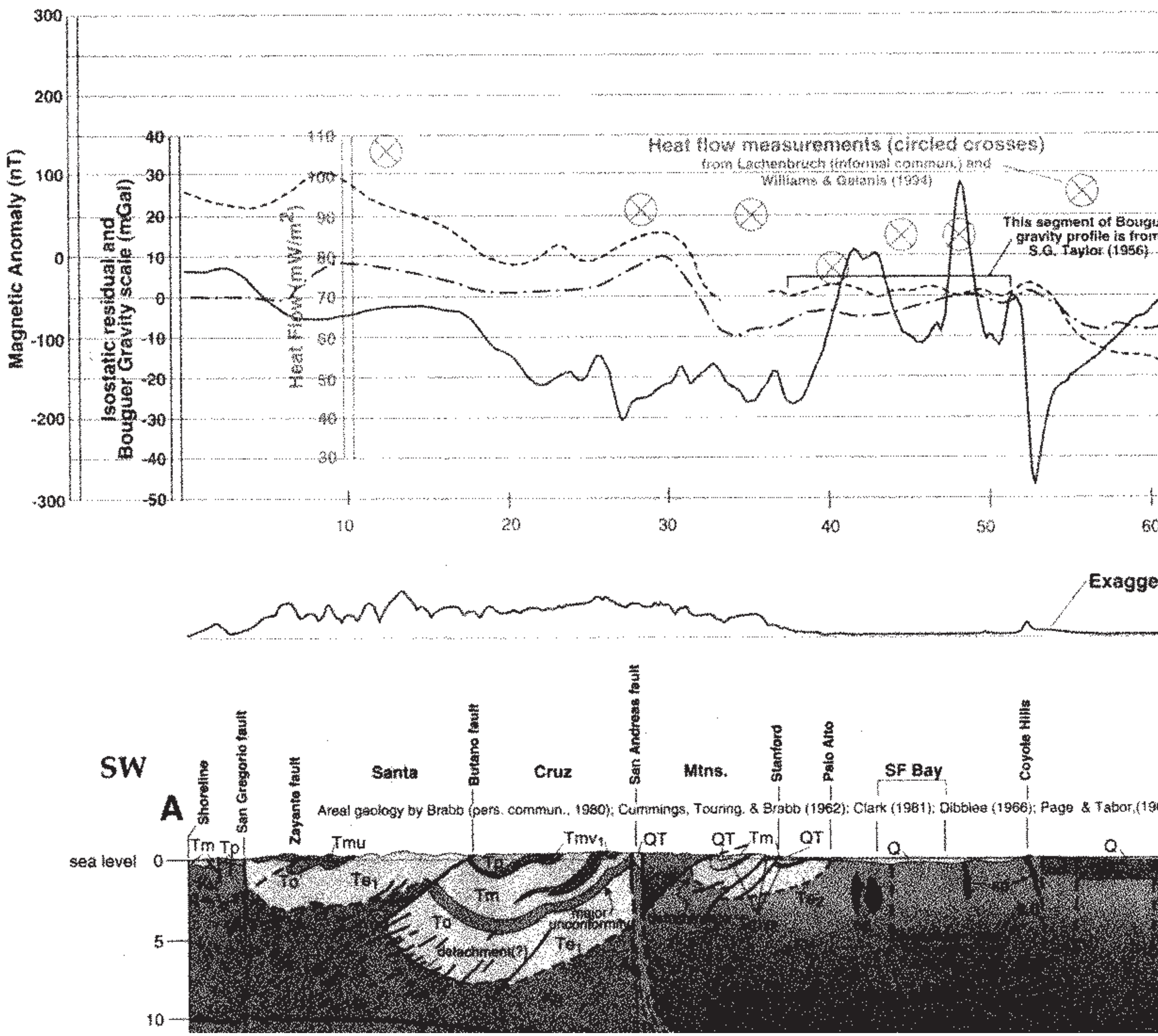
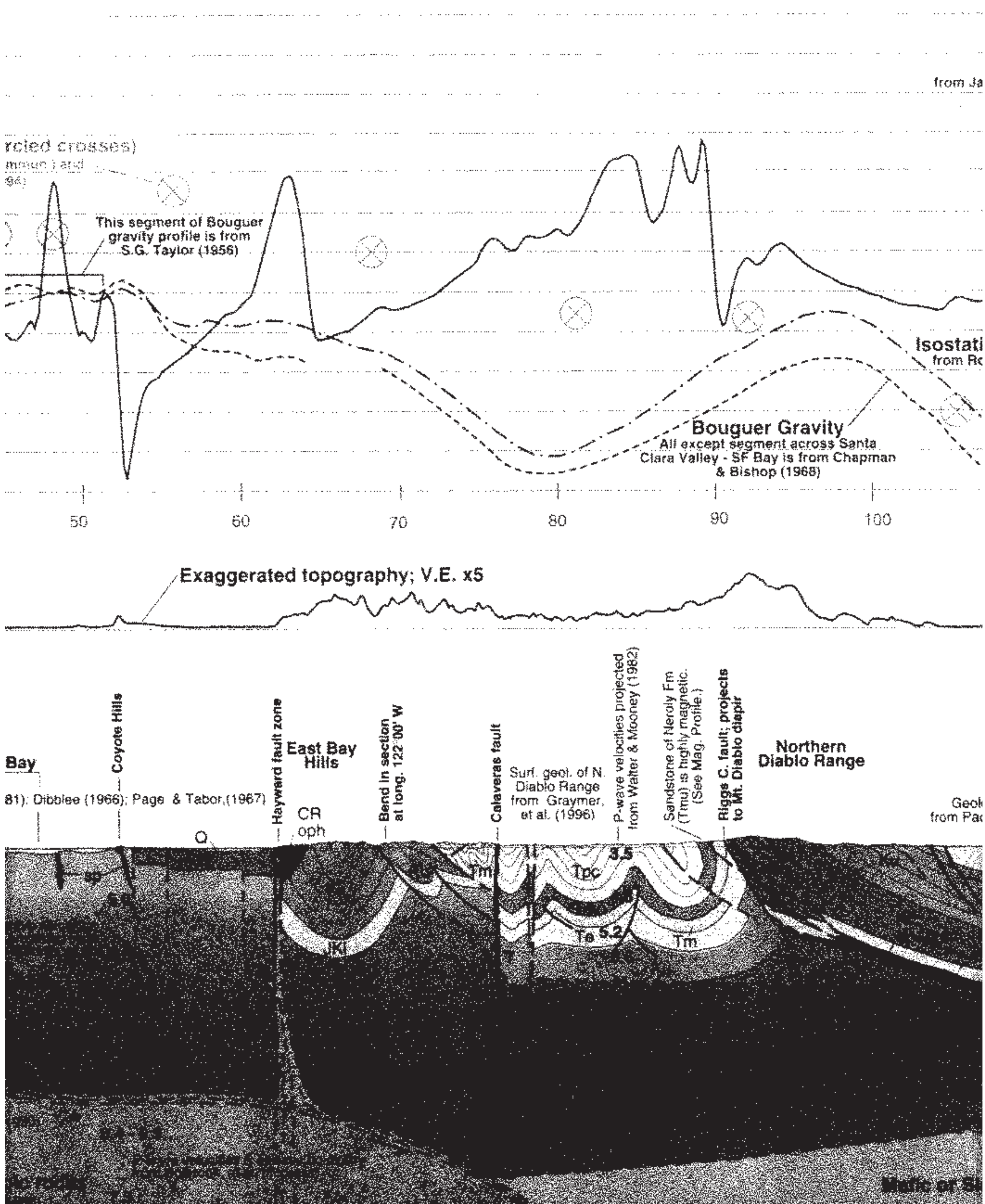


Santa Cruz Mts., San Francisco

Section A-A′ (See Fig.5 for location)



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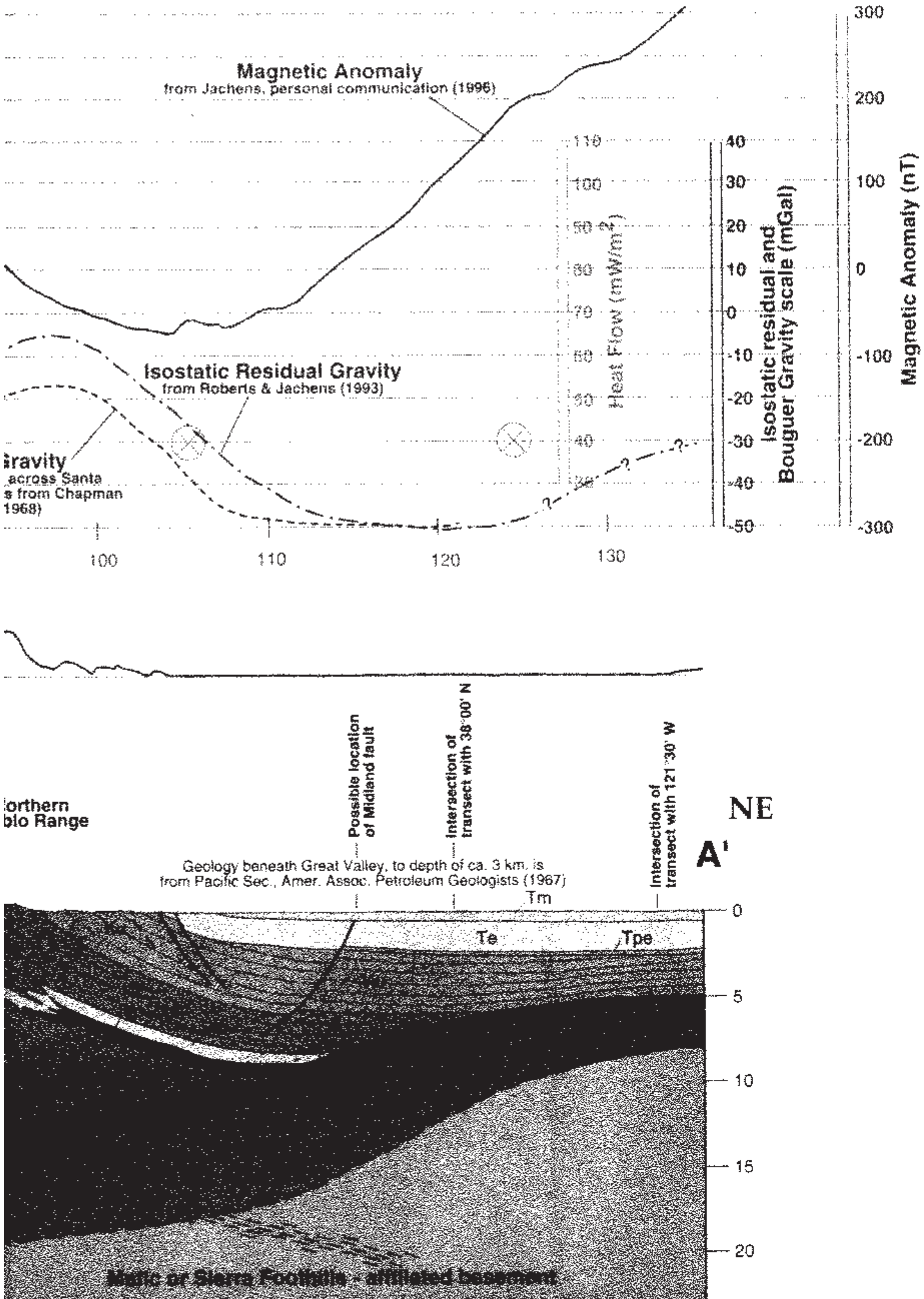
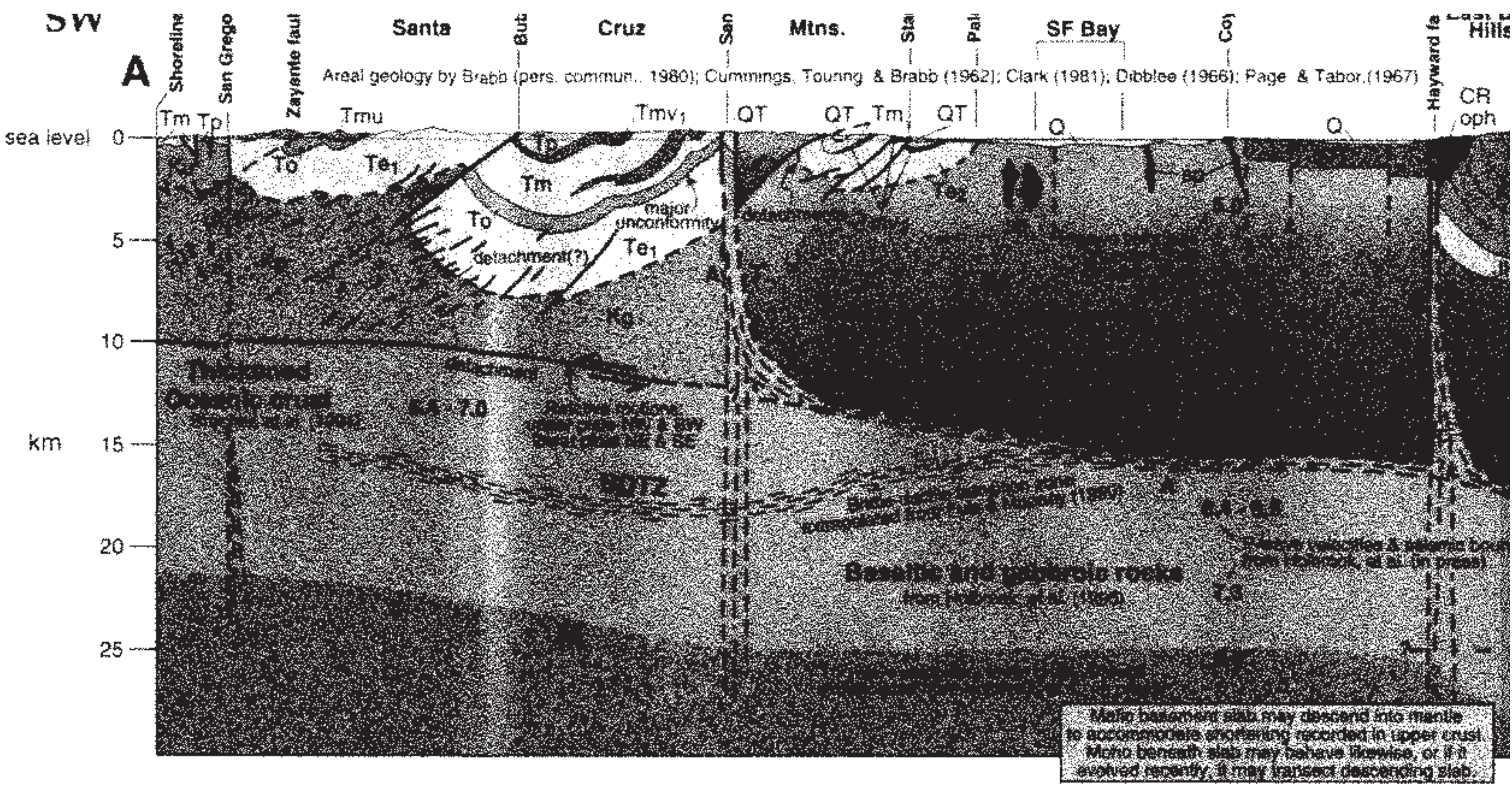




Q Quaternary nonmarine deposits, largely alluvium, generally undeformed.

QT Pliocene-Pleistocene clastic deposits, mainly nonmarine, locally deformed. Includes Santa Clara Fm.

Tpc Pliocene nonmarine clastic sedimentary rocks. Includes Green Valley - Tassajara Gp.

Tp Pliocene marine clastic sedimentary rocks. Includes Purisima Fm.

Tmu Upper Miocene marine sedimentary rocks. Includes Santa Margarita Fm. Santa Cruz Mudstone, Neroly Sandstone.

Miocene volcanic rocks. Tmv_1 is Lower Miocene Mindego Basalt; Tmv_2 is Middle Miocene Page Mill Basalt.

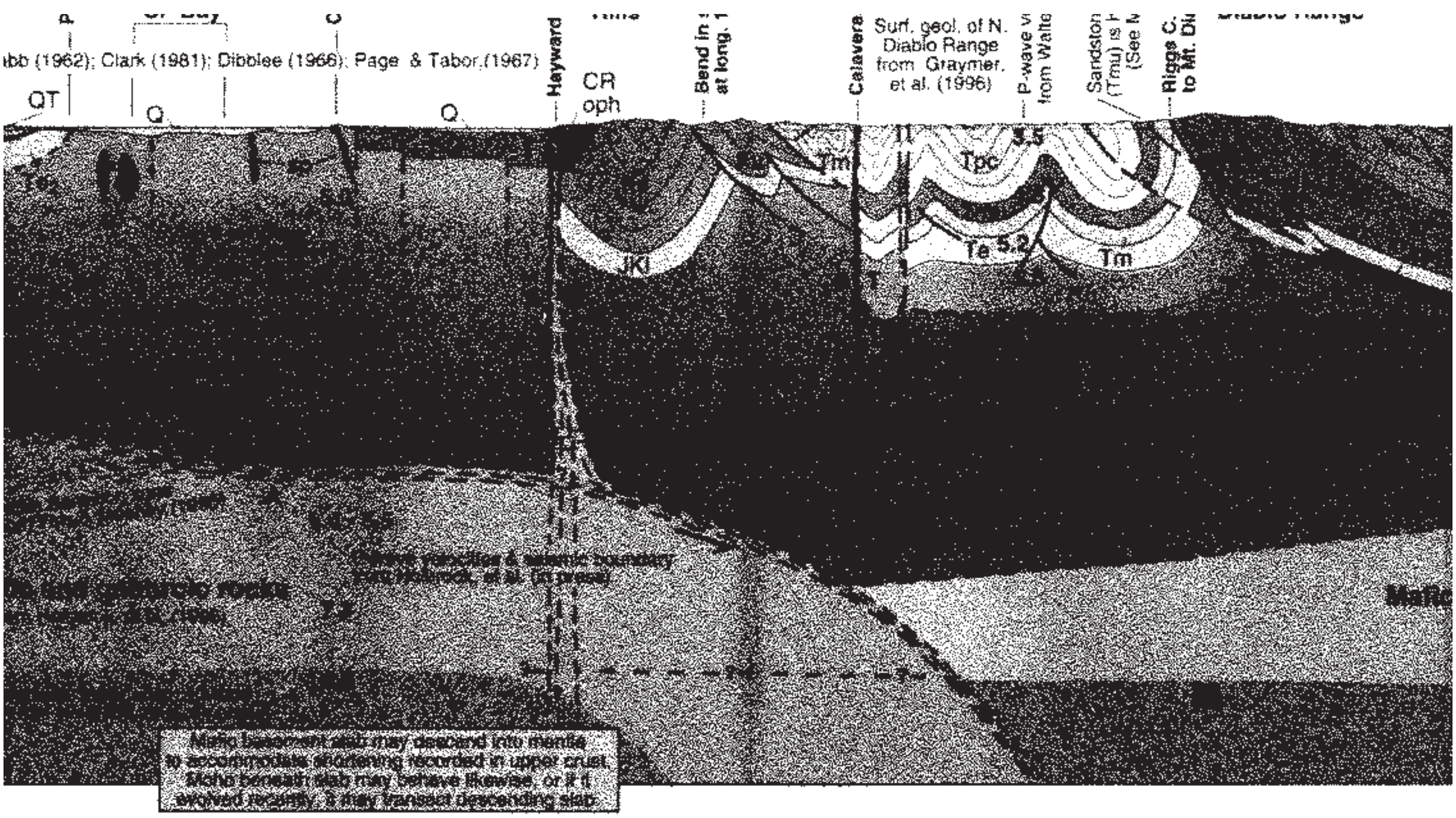
Tm Miocene marine sedimentary rocks. Includes Vaqueros Fm., Lambert Shale.

To Oligocene marine clasti rocks. Includes San Lo

Te Eocene marine sedimer chiefly turbiditic sandsto Te_1 is Butano Sandston Whiskey Hill Fm.

Tpa Paleocene marine clast

Kg Cretaceous granitic rock high-T metamorphic wa of Salinian Block (alloch



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ocks. Tmv_1 is Lower
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edimentary
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To Oligocene marine clastic sedimentary rocks. Includes San Lorenzo Fm.

Te Eocene marine sedimentary rocks, chiefly turbiditic sandstone and shale. Te_1 is Butano Sandstone; Te_2 is Whiskey Hill Fm.

Tpa Paleocene marine clastic rocks.

Kg Cretaceous granitic rocks and high-T metamorphic wall rocks, of Salinian Block (allochthonous).

Great Valley Sequence

Ku Upper Cretaceous marine turbiditic sandstone and shale.

Lower Cretaceous marine turbiditic sandstone and shale.

JKl Upper Jurassic and Lower Cretaceous marine shale (with sandstone).

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Compilation of data and some interpreted de
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Geology of upper crust from many authors,
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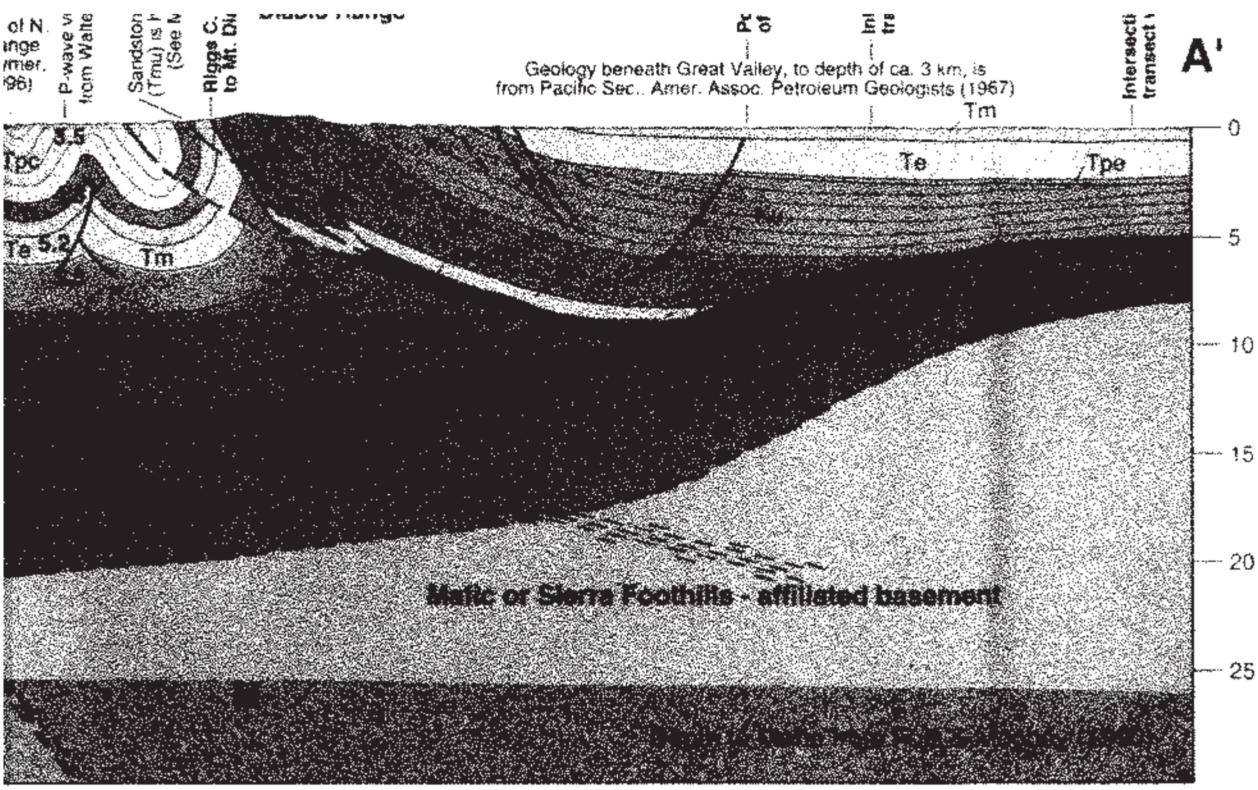

Upper Cretaceous marine turbiditic sandstone and shale.

Lower Cretaceous marine turbiditic sandstone and shale.

Upper Jurassic and Lower Cretaceous marine shale (with sandstone).

Serpentine and / or ophiolite. Where it underlies JKl, Kl or Ku, sp is part of the Coast Range Ophiolite; elsewhere it is affiliated with KJf.

Franciscan Complex. Mesozoic-Paleogene subduction complex; coherent bodies of Jurassic greenstone and Cretaceous sandstone; melanges of sandstone, greenstone, chert, serpentine, and blueschist in sheared shaly matrix.

BDTZ Brittle-ductile transition zone.

M Mohorovicic discontinuity

Numbers arrayed vertically (eg. 4.2, 5.0, 6.4, ...) indicate velocity of seismic P-wave in km/sec.

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Figure 10. Transect from

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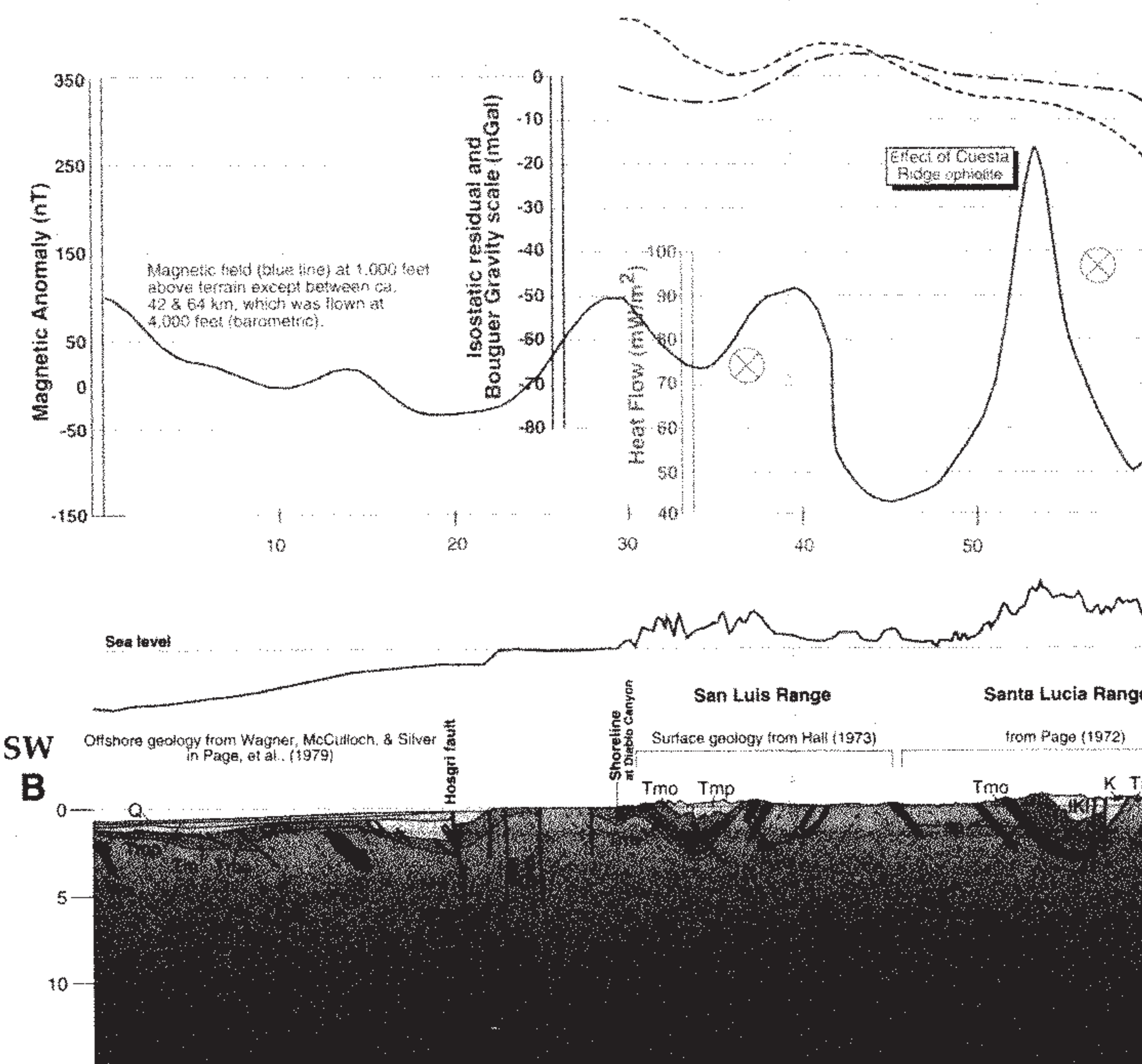

om Vicinity of San Luis Obispo across
Section B-B′ (see Fig.5 for location)

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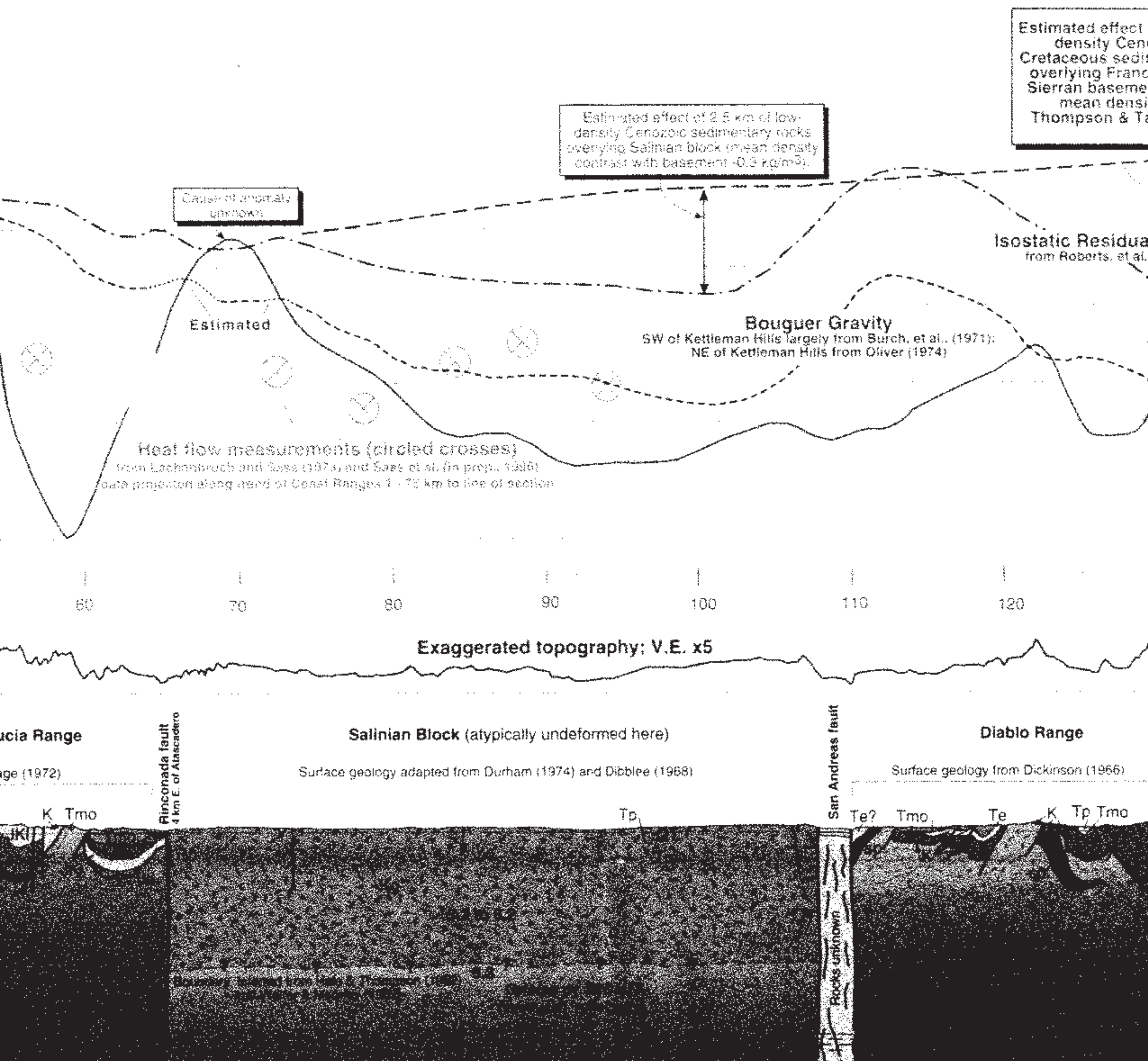

Vicinity of San Luis Obispo across K

Section B-B' (see Fig.5 for location)

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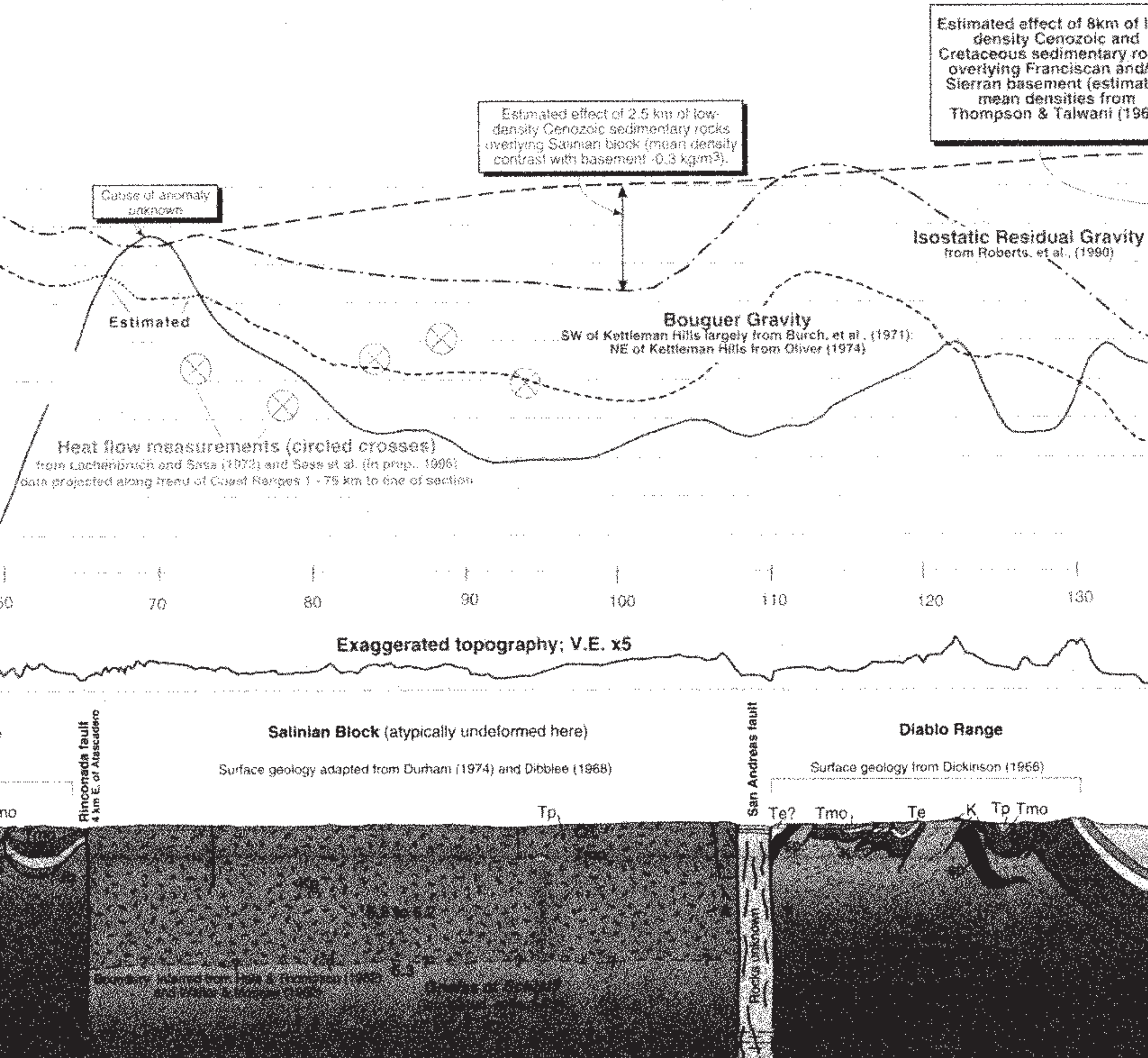

oss Kettleman Hills North Dome

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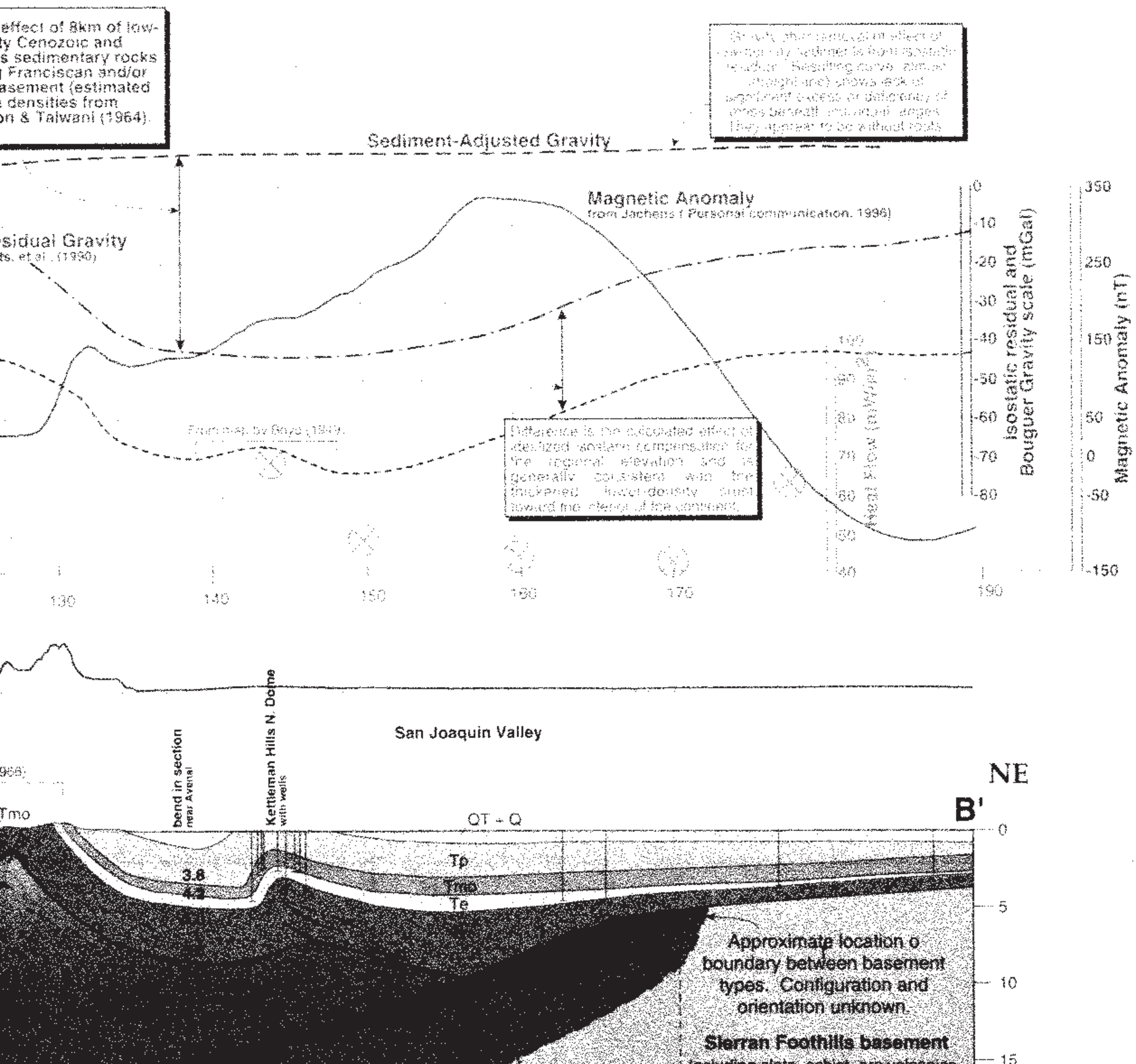
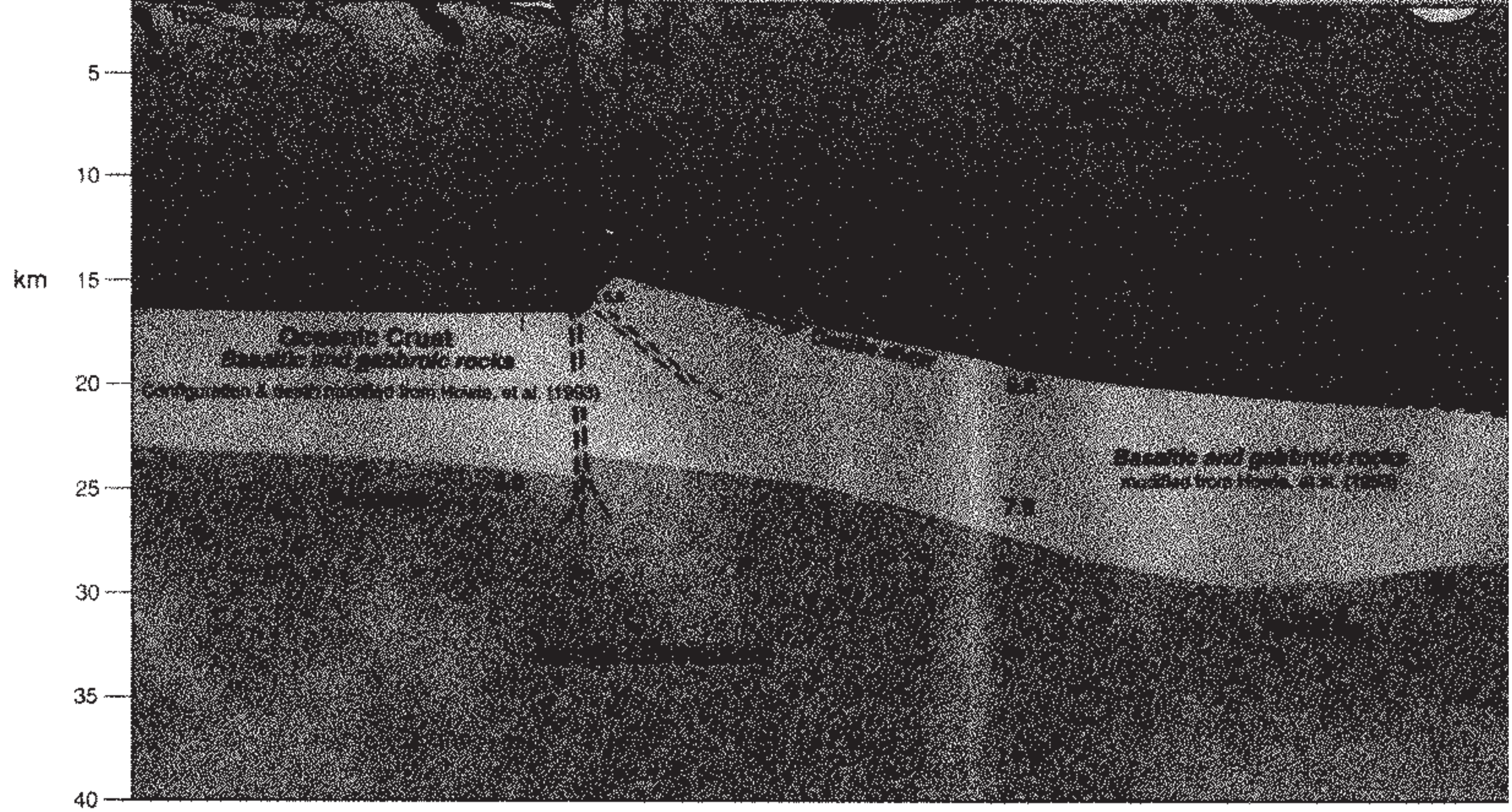
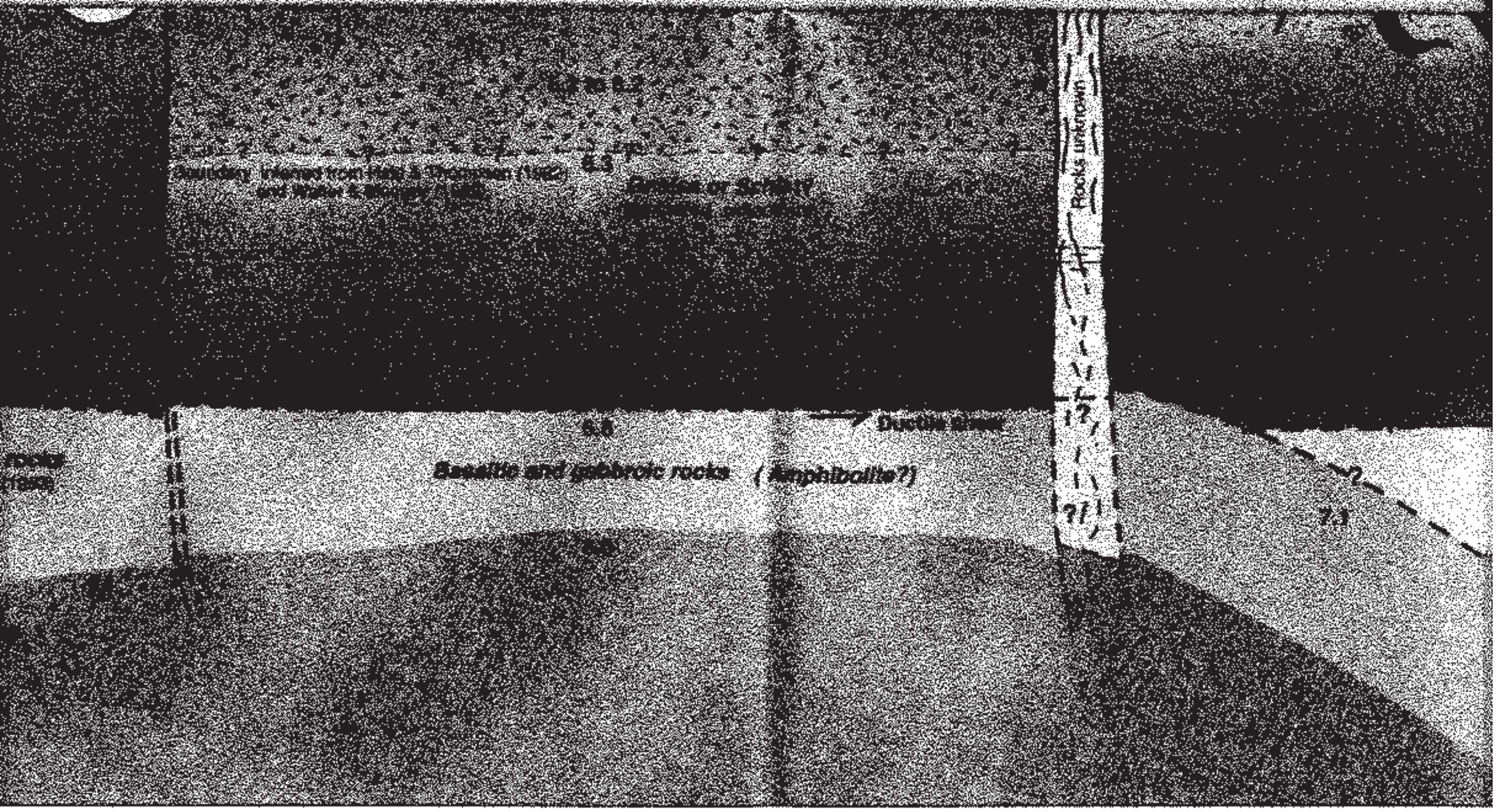




Q	Quaternary nonmarine deposits, largely alluvium, generally undeformed.	Tmp	Miocene-Pliocene marine clastic sedimentary rocks. Includes: Pismo Fm.	Tmo	Oligocene-M clastic sedim
QT	Pliocene-Pleistocene clastic deposits, mainly nonmarine, locally deformed. Includes: Paso Robles Fm., Tulare Fm.		Miocene volcanic rocks. Includes: Obispo Fm.	Te	Eocene mari rocks, chiefly and shale.
Tp	Pliocene clastic sedimentary rocks; marine.	Tm	Miocene marine sedimentary rocks. Includes: Vaqueros Fm., Monterey Fm.	Kg	Cretaceous high-T metar of Salinian B



Oligocene-Miocene marine clastic sedimentary rocks.

Eocene marine sedimentary rocks, chiefly turbiditic sandstone and shale.

Cretaceous granitic rocks and high-T metamorphic wall rocks, of Salinian Block (allochthonous).

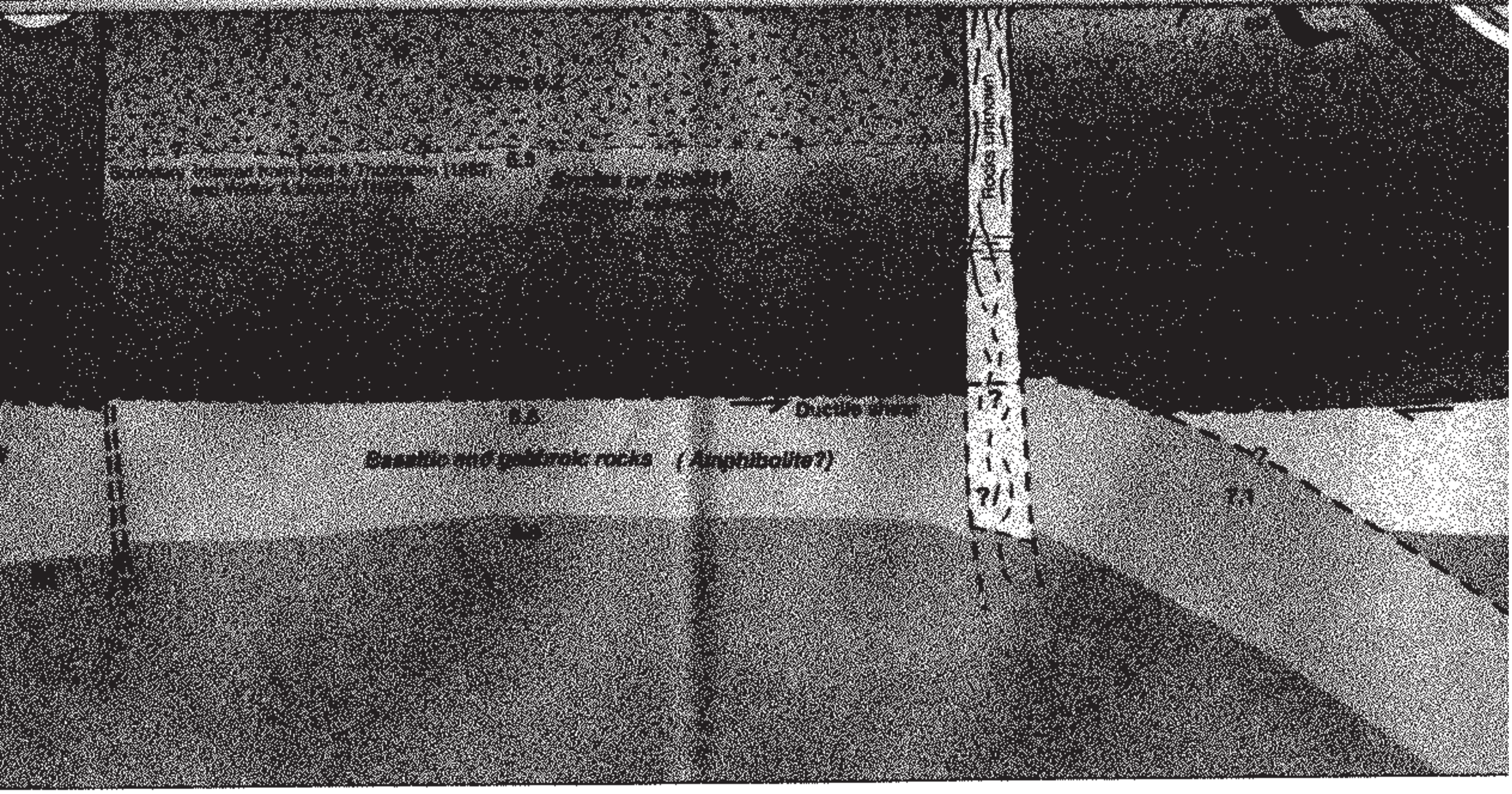
Cretaceous (mostly Upper) marine turbiditic clastic sedimentary rocks.

JKl — Upper Jurassic and Lower Cretaceous marine shale (with sandstone).

Jb — Jurassic basalt, keratophyre, diabase, sheeted dikes, gabbro, plagiogranite. (Part of Coast Range Ophiolite.)

Serpentine. Where it underlies Jb, sp is part of the Coast Range Ophiolite; elsewhere it is affiliated with KJf.

Franciscan Complex. Mesozoic-Paleogene subduction complex; coherent bodies of Jurassic greenstone and Cretaceous sandstone; melanges of sandston greenstone, chert, serpentine, and blueschist in sheared shaly matrix



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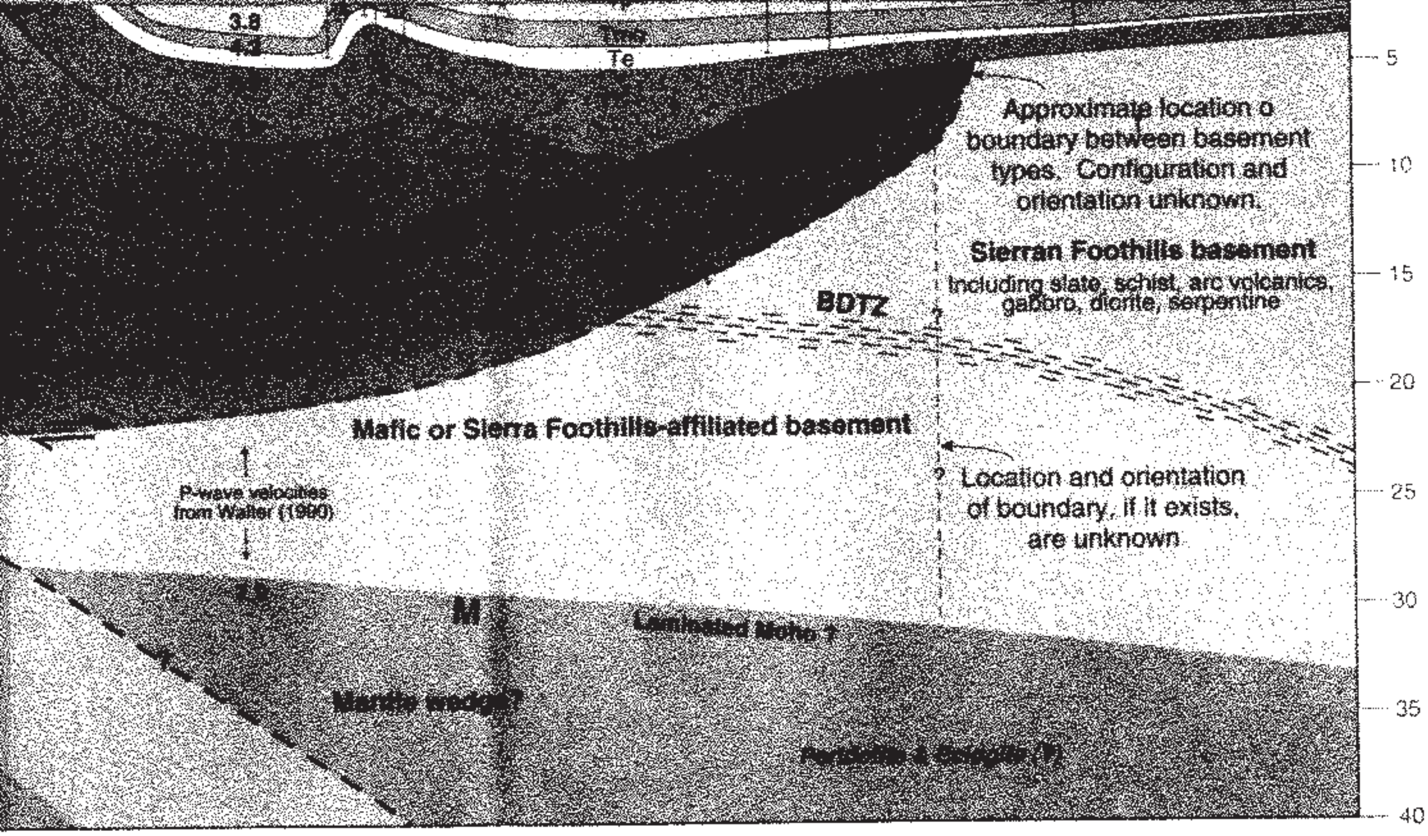

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