DAVID BLOOM

RIDGE BASIN, SOUTHERN CALIFORNIA: INTRODUCTION

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Ridge Basin is a small sedimentary basin within the central Transverse Ranges of southern California that is now uplifted and deeply dissected so that its geological history can be deciphered. During the late Miocene Epoch and lasting into the Pliocene Epoch an extremely thick stratigraphic section, exceeding 13,500 m (44,000 ft) was laid down within the narrow basin by what is interpreted as a conveyor belt mechanism involving a moving depocenter. Sedimentary facies within the basin are unusually well exposed and document a change through time from a marine embayment to a lacustrine and alluvial environment. Beds overlap faults around the margins of the basin and reveal much about the interplay between sedimentation and tectonics in this area, a region sited within the splintered transform boundary between the Pacific and North American lithospheric plates. The papers within this volume, along with the geologic map, cross sections and diagrams, document and illustrate this complex history. In short, we hope that the papers will be of interest to those concerned with sedimentation and the environmental interpretation of facies, with the interplay between sedimentation and tectonics along a transform boundary, and with regional geology. The location of Ridge Basin and the arrangement of stratigraphic units within it are shown in Figures 1 and 2, and an aerial view in Figure 3.

LOCATION AND ACCESSIBILITY

Ridge Basin, crossed by Interstate Highway 5, lies within rugged terrain between Los Angeles on the southeast and Bakersfield on the northwest (Fig. 1). Kilometer after kilometer along the highway of northwestward dipping strata are accordingly viewed by many geologists as they speed along the freeway between the southern California megalopolis surrounding Los Angeles and the San Joaquin Valley. The southeastern part of the basin lies near the town of Castaic, and the northwestern at the San Andreas Fault near Gorman. The basin as now exposed is from 6 to 15 km wide (3.7 to 5.2 mi) and 30 to 40 km long (19 to 25 mi) and has an area of about 200 km^2 (77 mi^2). From Interstate 5 access to the basin is provided by several all-weather roads, including State Highway 138, Templin Highway, the Old Ridge Route, and a few others. Several fire roads and other gravel roads enable any part of the basin to be reached easily during the course of day's field work on foot.

The terrain bordering Ridge Basin, however, is exceptionally rugged and brushy so that large-scale geological investigations are seriously hampered. From near the San Gabriel Fault on Piru Creek, for example, the topographic relief rises from 480 m (1600 ft) to 2050 m (6730 ft) at the summit of Cobblestone Mountain in a linear distance of only 13 km (8 mi). [For comparison, the relief between Grand Canyon Lodge (El Tovar) and the Colorado River, Grand Canyon National Park, is 1333 m (4370 ft) over a straight-line distance of 4.8 km (3 mi).] This "back country" southwest of the San Gabriel Fault is largely without trails, and is underlain by complex igneous and metamorphic rocks which still have not been studied in detail. The area bordering Ridge Basin on the east is also very rugged, especially to the southeast of Liebre Mountain.

PREVIOUS WORK

Although strata within Ridge Basin had no doubt been examined previously by geologists (Blake, 1857; Goodyear, 1888), the first description of them is by Hershey (1902a, 1902b). In a survey of the Tertiary deposits of southern California, he made a reconnaissance map of the northern part of the basin, and published brief descriptions of the rocks. The San Andreas fault zone, along the northern border of the basin, had been recognized by Schuyler (1894). After the San Francisco Earthquake of 1906, geologists traced the fault through the region (Fairbanks, 1907; Lawson et al., 1908). The southern part of the area is mentioned briefly by Kew (1924) with comments on the oil and gas potential.

Clements (1929, 1932, 1937), undertook the first investigations of the structure and stratigraphy of the southern and central parts of Ridge Basin and named the Ridge Route Formation. He assigned it to the Pliocene Series on the basis of its stratigraphic position. Eaton (1939), made the first structural and sedimentological analysis of the basin as whole and concluded that the basin was a structural depression related to displacements on the San Gabriel Fault. He showed that the style of deposition within the basin was largely controlled by tectonic activity along its margins. Stratigraphically, Eaton subdivided the beds into four "divisions" and employed a system of nomenclature differing from modern procedures in stratigraphic nomenclature. Nonetheless, his comments on relations between tectonics and sedimentation, and regional implications, are prescient.

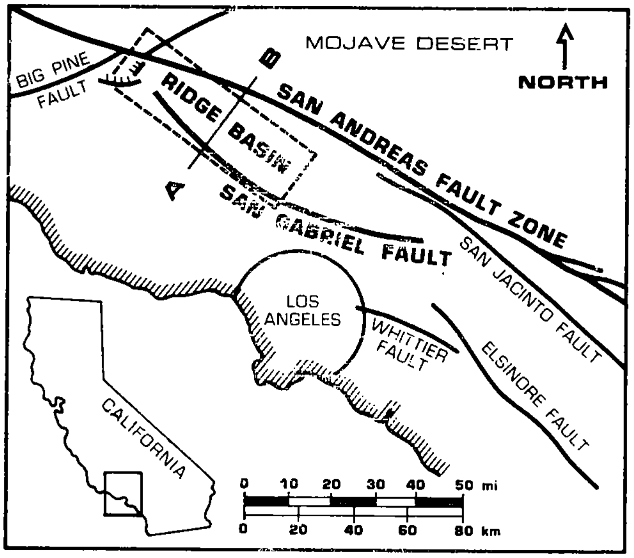

Figure 1. Location of Ridge Basin, southern California Note line of cross section (A-B) shown in Figure 2.

in Crowell, J.C., and Link, M.H., eds., 1982 **Geologic History of Ridge Basin, Southern California**, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 1-4.

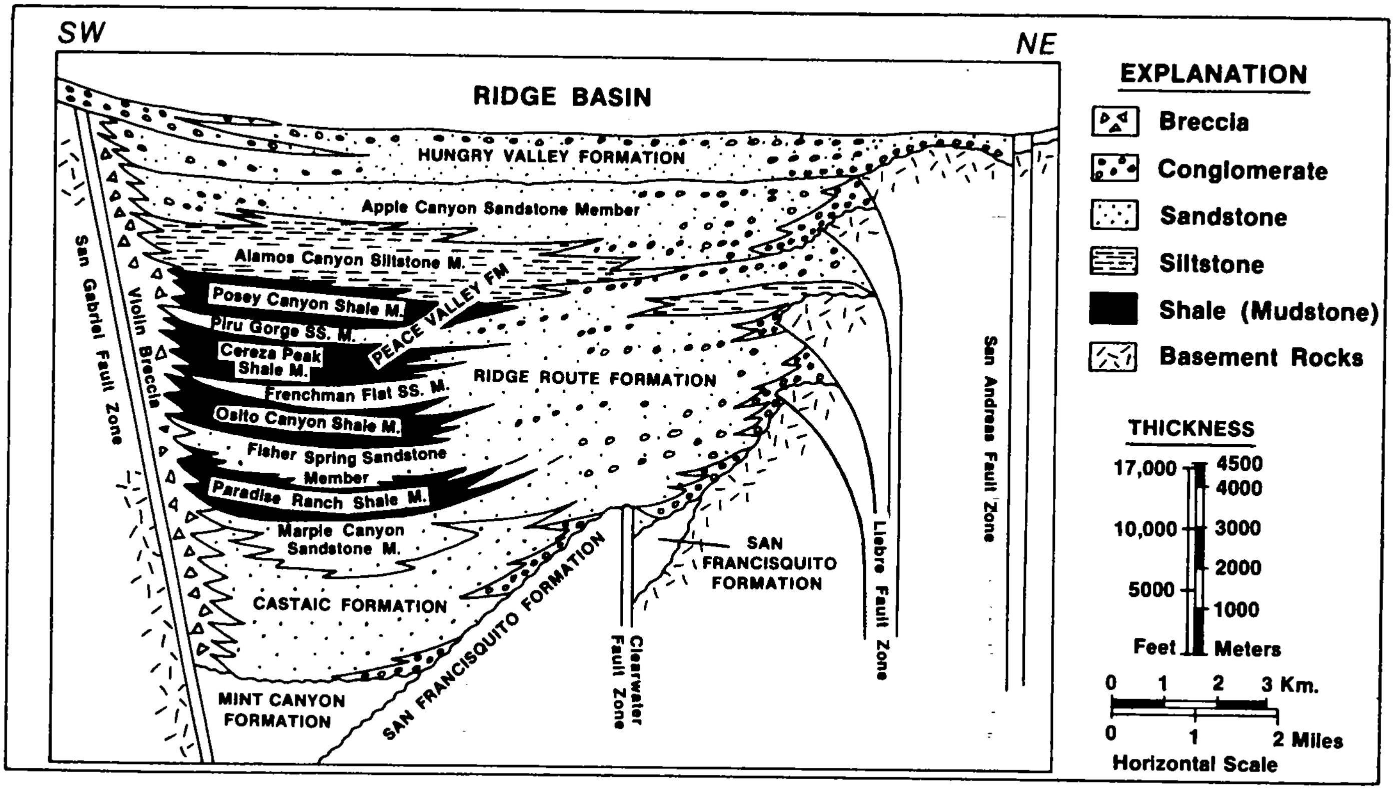

Figure 2. Diagrammatic cross section of Ridge Basin showing major stratigraphic and structural relations (location of cross section depicted in Figure 1).

Beginning in the 1940s, Ridge Basin and surrounding regions, have been the sites of many student investigations, including Masters theses and doctoral dissertations. Those that dealt directly with the basin include Chambers (1947), Crowell (1947), Dehlinger (1950), Harris (1950), Smith (1951), Faggioli (1952), Jennings (1952), Stanton (1960), Shepard (1961), and Stanley (1966). Some of the results of these thesis investigations have been published (Crowell, 1950, 1952; Dehlinger, 1952; Jennings, 1953; and Stanton, 1966). Those concerned with regions adjacent to Ridge Basin, or with rocks and fossils correlative with those in Ridge Basin, include Pfaffman (1941), Kriz (1943), Wright (1943). Martin (1947), Buffington (1947), White (1947), Johnson (1952), Miller (1952), and Shepherd (1960).

Megafossils from the marine Castaic Formation were reported by Eaton (1939), Durham (1948a, 1948b), White and Buffington (1948), Wright (1948), Stanton (1960, 1966, 1967), Smale (1978), Ensley (1980), and Young (1980). A late Miocene (Jacalitos) age and shallow-water, semi-tropical assemblage was suggested by Ensley (1980). Microfossils are rare. Skolnick and Arnal (1959) describe one microfossil assemblage from a well south of Castaic that indicate a restricted deep-water basin. The microfossil age is upper Miocene (Mohnian and Delmontian). McDougall and Stanton present microfossil and megafossil data, respectively, for the Castaic Formation in this volume.

Fossils from the nonmarine Ridge Basin Group include fish (Jordan, 1924; David, 1945; Shepard, 1961; Young, 1980; Smith, 1981); plant fragments (Axelrod, 1950; Shepard, 1961; Irvine, 1977; Smith, 1981); dragonfly nymphs (Shepard, 1961; Squires, 1979); vertebrate remains of horses, sloth, rhinoceros, antelope, turtles, camel, mastodon, rabbit, and a large cat (Stock, in Crowell, 1950; Miller and Downs, 1974; Young, 1980); nonmarine molluscs and ostracodes (Crowell, 1950; Young, 1980); and algae (Link et al., 1978). The vertebrate assemblage (Crowell, 1950; Miller and Downs, 1974), which occurs near the top of the Ridge Basin Group gave an age of latest Hemphillian (Miller and Downs, 1974). The other fossils were not diagnostic of age but instrumental in indicating the paleoclimatic and environmental conditions in Ridge Basin. Permanent lake(s) fed by rivers in a semi-arid climate with grassland and some woody or thicketed areas near sea level are suggested by most of these authors. Forester and Brouwers discuss ostracodes, Young describes freshwater mollusks, Welton and Link deal with the vertebrate assemblages, McDougall with foraminifera, Smith, Harper, and Wood with trace fossils, and Squires treats the fossil insects in this volume.

Sedimentological and paleontological studies highlight the work in Ridge Basin during the 1970's and 1980's. Irvine (1977) studied the Posey Canyon Shale Member of the Ridge Basin Group and concluded these chemical deposits formed in a closed, saline alkaline-lake system. Nilsen et al (1977) described dish structures from the Ridge Basin Group suggesting they formed in turbidite deposits. Ehlert and Ehlig (1977) and Farley and Ehlig (1977) discuss the palinspastic reconstruction and offsets along major faults of Ridge Basin. Cooper (1977) reviews the

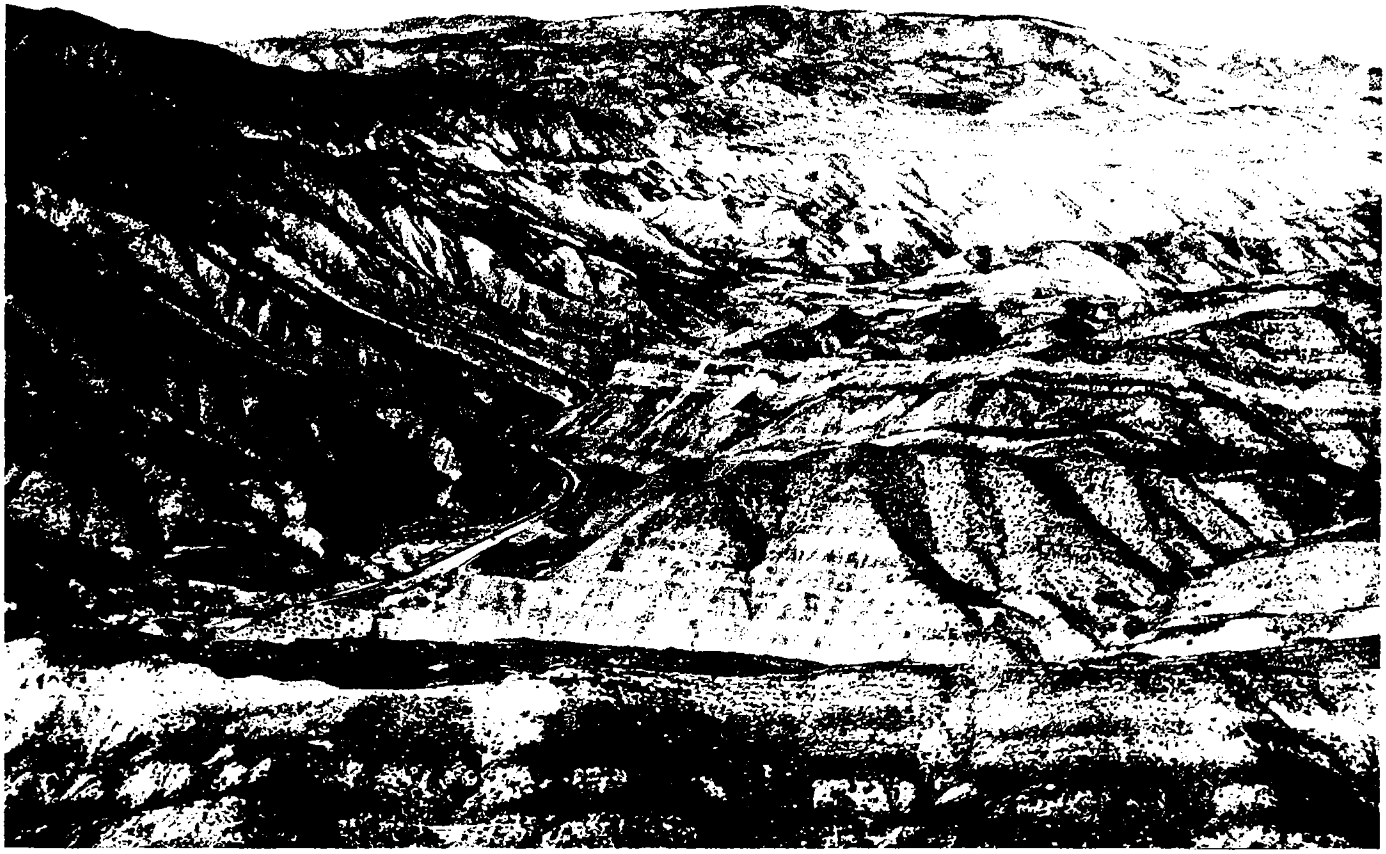

Figure 3. Air photograph of the northern part of Ridge Basin, looking northwest. Frenchman Flat and Highway 99 at lower left, Hungry Valley and Frazier Mountain in background. Photograph taken in 1948 before construction of Pyramid Dam and impounding of Pyramid Lake.

geology and interprets the sedimentary sequences of the Ridge Basin in a guidebook article. Stromatolites and other carbonate sedimentary rocks were discussed and interpreted by Link et al (1977, 1978). Smale (1978) analyzed the Marple Canyon Sandstone Member of the Ridge Route Formation (lowermost unit of the Ridge Basin Group) and identified submarine fan, fan-delta, and supratidal mudflat facies. Link and Osborne (1978) interpreted facies relationships in Ridge Basin and recognized alluvial fan fluvial, marginal lacustrine, and offshore facies. Vertebrate fossils (Miller and Downs, 1974), dragonfly nymphs (Squires, 1979) and nonmarine mollusks, ostracodes, and vertebrate remains (Young, 1980) were collected and identified.

Ensley and Verosub (1979) and Ensley (1980) dealt with the magnetostratigraphy of the Ridge Basin Group based on detrital remanent magnetization (DRM). The results of these studies show that the lowermost beds of the Ridge Basin Group fall into the upper two magnetozones of Epoch 8 and that the middle part of the Group is in the lowest magnetozone of Epoch 5. The age range for the 12 polarity reversal horizons is 8.4 m. y. to 6.07 m. y. In addition, Ensley (1980) correlates the Kinsey Ranch fauna of Miller and Downs (1974) with other known megafossil/magnetostratigraphic zonations and concludes that an age from 5.5 to 5.0 m. y. is appropriate for this faunal assemblage. Investigations recently completed include a study of the Fisher Spring and Frenchman Flat Sandstone Members of the Ridge Basin Group by Harper (1981) and Hollywood (1981), respectively. Smith (1981) examined the fine-grained facies of the Peace Valley Formation and its members. Wood (1981) studied the Apple Canyon Sandstone and Alamos Canyon siltstone Members. Summaries of most of this work are presented in this volume.

Tectonic studies in the Ridge Basin region have gone on hand in hand with geologic mapping and stratigraphic investigations. The displacement history of the major faults, and especially of the San Andreas and San Gabriel, has been a matter of discussion and controversy for many years. Comments by Clements (1937) and Eaton (1939) pointed out that the San Gabriel Fault had been active during deposition of beds in Ridge Basin. Clements named the fault bordering Ridge Basin the Palomas Canyon Fault, but Kew (1924) previously had suggested that this fault was the northwestern extension of the San Gabriel Fault. Eaton (1939), following additional regional work, accepted Kew's suggestion and so have subsequent investigators. Beginning in 1950, much discussion flourished over whether geologic relations along the San Gabriel Fault required major right slip, or whether they could be satisfactorily explained by dip slip primarily in combination with paleogeographic arrangements resulting from activity on the fault zone (Reference in Pacific Petroleum Geologist, v. 4, n. 7, p. 1, July, 1950; Crowell, 1952, 1954, 1960, 1962, etc.; Paschall and Off, 1959, 1961). Additional work along the San Gabriel Fault has included that by Shepard (1961, 1962), Woodburne (1975), Weber (1977), and Stitt (1980). Some of the present views concerning the history of the fault zone are dealt with in the tectonics paper in this volume (Crowell, this volume).

The Clearwater Fault, which enters Ridge Basin on the east and is overlapped by units of the Ridge Route Formation, has been studied by Nickell (1928), Clements (1930, 1932, 1937), Simpson (1934), Eaton (1939), Pfaffman (1941), Smith (1951), and Stanley (1966). Most of these geologists stated that the fault had had components of both right slip and dip slip, but the amounts are not yet adequately established. Ensley (1980) tentatively concluded, based on overlap relations and magnetostratigraphic studies that the Clearwater Fault originated prior to 8.13 m. y. ago and died 7.81 m. y. ago.

The Liebre fault system, which includes at least four strands along the northeastern border of Ridge Basin, has been described by Eaton (1939), Faggioli (1952), Crowell (1952b, 1954b, 1964, 1975c), and Jennings (1952, 1953). As shown on Crowell, et al. (1982) the strands of the Liebre system are successively overlapped toward the northeast and therefore each stopped activity in turn during deposition of strata deposited in Ridge Basin. According to Ensley (1980), displacement on the northeasternmost strand ceased between 5.5 and 5.0 m. y. ago.

The Frazier Mountain thrust system, located to the northwest of Ridge Basin, has been studied by Buwalda et al (1930), Buwalda and Gutenberg (1935), Crowell (1947, 1950, 1954b, 1975c), and farther west by Carman (1954; 1964).

The San Andreas fault zone, which truncates Ridge Basin on the north, has been described in this region by Schuyler (1897), Fairbanks (1907), Lawson et al (1908), Crowell (1947, 1952b, 1954b, 1975c), Fine (1947), Wiese and Fine (1950), Sieh (1978), Crippen (1979), and Duebendorfer (1979). Many papers have been written concerning the displacement history of the San Andreas fault system, and many geologists have made comments pertaining to the Ridge Basin area and the San Gabriel Fault, but they are too numerous to list here.

Guidebooks and guidebook maps to Ridge Basin include Loel (1947), Crowell (1954b, 1964, 1973a, 1975c), Woyski, et al. (1977), and Advocate, et al. (this volume).

Violin Breccia along Piru Creek below Hardluck Campground

STRATIGRAPHIC NOMENCLATURE AND AGE OF MIOCENE STRATA, RIDGE BASIN, SOUTHERN CALIFORNIA

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ABSTRACT

The oldest rocks in the Ridge Basin region are Precambrian gneiss, diorite, gabbro, amphibolite, and anorthosite which are intruded by upper Cretaceous granitic rocks of quartz diorite, granodiorite, quartz monzonite, and granite composition. Unconformably overlying these rocks are Cretaceous and Paleocene marine sedimentary rocks of the San Francisquito Formation. These sequences, in turn, are overlain unconformably by 13,000 to 14,000 m (42,000 to 46,000 ft) thick section of Miocene to Pliocene marine and nonmarine sedimentary rocks. These units are from oldest to youngest: Mint Canyon Formation, Castaic Formation, and Ridge Basin Group. The latter consist of the Violin Breccia, Ridge Route Formation, and Hungry Valley Formation. The Ridge Route Formation is subdivided in central Ridge Basin into five members (listed from oldest to youngest; four are new names): Marple Canyon, Fisher Springs, Frenchman Flat, Piru Gorge (redefined, formerly a formation of the Ridge Basin Group), and Apple Canyon Sandstone Members. These units interfinger and subdivide the Peace Valley Formation (redefined) into five members (new names, listed from oldest to youngest): Paradise Ranch, Osito Canyon, Cereza Peak, and Posey Canyon Shale Members and the Alamos Canyon Siltstone Member.

INTRODUCTION

The basement rocks and the stratigraphic units of Ridge Basin, their age, and stratigraphic relationship are shown in Figure 1 and described here. Individual formations and members as well as facies and facies associations are discussed in later chapters. The oldest rocks in the Ridge Basin are Precambrian gneiss, gabbro, diorite, amphibolite, and anorthosite which are intruded by upper Cretaceous batholithic rocks of quartz diorite and quartz monzonite composition. Nearby on the southeast the Pelona Schist of probable Mesozoic age is exposed. Most of these basement rocks are similar in composition and presumed age to those in the San Gabriel Mountains (Crowell and Walker, 1962; Ehlig and Crowell, this volume). Unconformably overlying these rocks are the Cretaceous and Paleocene marine sedimentary rocks of the San Francisquito Formation. These sequences in turn, are overlain unconformably by Miocene to Pliocene marine and nonmarine sedimentary rocks. These units are from oldest to youngest: Mint Canyon Formation, Castaic Formation, Violin Breccia, Ridge Route Formation, Peace Valley Formation, and Hungry Valley Formation. The latter four constitute the Ridge Basin Group.

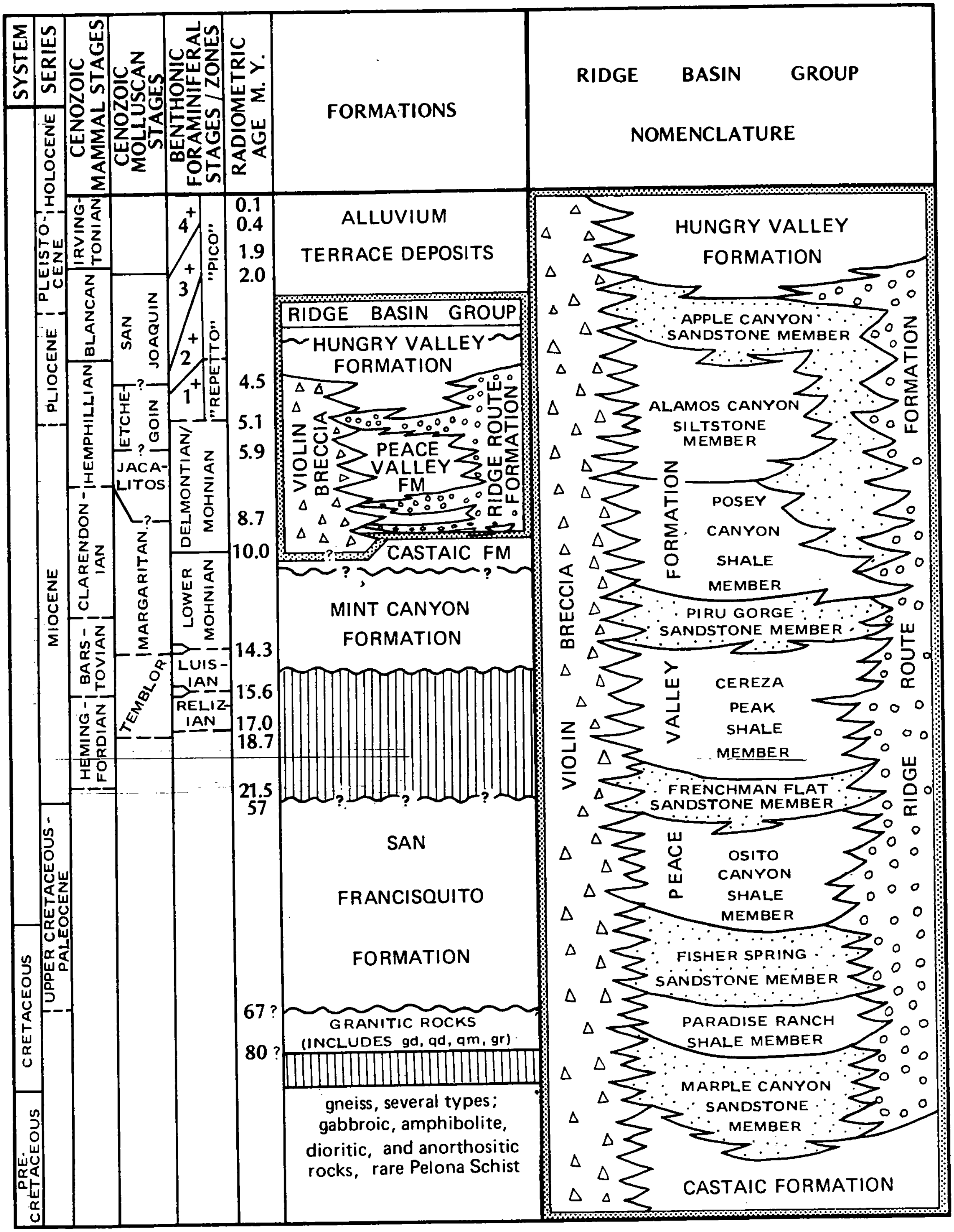
BASEMENT ROCKS

Precambrian metamorphic complexes occur on both sides of Ridge Basin within the younger basement complex of Late Cretaceous age and as fault slivers along the San Gabriel Fault (Fig. 1). They have been discussed in detail by Crowell (1947, 1954b, 1964), Faggioli (1952), Shepard (1961, 1962), and Ehlig and Crowell (this volume), and dated by Silver (1966, 1968, 1971). These rocks are highly fractured and sheared and cut by dikes. Crowell and Shepard separate these metamorphic rocks into two groups: (1) an older gneiss, hornblende diorite, gabbro, amphibolite, anorthosite complex and (2) a younger gneissic complex of gneiss, migmatite, and quartz monzonite. The older metamorphic assemblage consist of 85 percent gneissic hornblende diorite, gabbro and amphibolite, 10 percent quartzo-feldspathic gneiss, and 5 percent anorthosite (Shepard, 1961). Anorthosite occurrences are limited to a small area along the western side of the basin and to areas southeast of Ridge Basin in the western San Gabriel Mountains (Higgs, 1954; Crowell and Walker, 1962). These rocks consist of 95 percent anorthosite (calcic andesine [An_{43}] plagioclase) and 5 percent accessory minerals such as hornblende, chlorite, apatite, epidote, hypersthene, and ilmenite or magnetite.

The younger gneiss complex occurs on the northwest, southeast, and eastern part of the basin and consists of gneiss, migmatite, and quartz monzonite (Crowell, 1964). Because of their similar appearance, structure, composition, and relationship with the granitic plutons, these outcrops are tentatively considered to be correlative with each other. The gneiss is well foliated and contains 20-40 percent microcline and orthoclase, 10-40 percent quartz, and 1-30 percent biotite. Accessory minerals include hornblende, chlorite, magnetite, apatite, and zircon. Where the gneiss has been intruded by the quartz monzonite pluton, it is more highly foliated and occasionally migmatized. The dominant rock types within the complex are quartz diorite and quartz monzonite (Shepard, 1961). Locally amphibolite, marble, and schists occur in the exposures east of Ridge Basin. The age of this gneiss is uncertain. The only valid relationships is the intrusion of the gneiss by the quartz monzonite pluton which is dated as Cretaceous (Evernden and Kistler, 1970).

Cretaceous granitic rocks border the northeastern corner of Ridge Basin and occur in the central part of the western margin (Fig. 1). These exposures are similar in structure, composition, and relationship to the gneiss complexes. Crowell (1952b, 1964) interprets these plutons to be generally of quartz dioritic composition, whereas Clements (1937), Bettinger (1948), Dehlinger (1952), all consider the average composition to be granodioritic with variations into quartz monzonite and quartz diorite or granite. Shepard (1961) indicates that one pluton has a quartz monzonite composition with gradations into granodiorite. Crowell et al. (1982) show the distributions of these rocks grading from quartz diorite to granodiorite to quartz monzonite to granite. Quartz monzonite, granodiorite, granite, and quartz diorite are used here as descriptive field term following the useage of Crowell (1954), and Crowell et al. (1982). In a petrographic sense, these igneous rocks could also be classified as granodiorite, granite, and adamellite (Streckeisen, 1973; J. Lawford Anderson, pers. comm.). They consist of 25-50 percent quartz, 35-50 percent plagioclase with An_{28-37}, and 20-30 percent

in Crowell, J.C., and Link, M.H., eds., 1982, **Geologic History of Ridge Basin, Southern California,** Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 5-12.



+ 1. REPETTIAN, 2. VENTURIAN, 3. WHEELERIAN, 4. HALLIAN.

Figure 1. Stratigraphic nomenclature and ages for Ridge Basin strata and basement rocks.

microcline and orthoclase, 0-8 percent biotite, and up to 2 percent hornblende. Accessory minerals include apatite, sphene, and magnetite (Shepard, 1961). A pluton in this complex has been dated at 66 m. y. by K/Ar methods on biotite by Evernden and Kistler (1970, loc. no. 150) and is informally known as the Gold Hill Alaskite (Crowell, pers. comm.). Some of these basement rocks are described more fully by Ehlig and Crowell (this volume).

STRATIGRAPHIC UNITS

San Francisquito Formation

The San Francisquito Formation of Cretaceous and Paleocene age is exposed only along the southeastern border of Ridge Basin and extends southeastward over 40 km to Bouquet Reservoir (Dibblee, 1967; Kooser, 1980). This unit lies nonconformably upon basement rock and is overlain unconformably by the Ridge Basin Group and Castaic Formation (Fig. 1) and consists of a basal conglomerate member, a middle sandstone member, and an upper shale member. The San Francisquito Formation has been studied by Nickell (1928); Clements (1932, 1937); Pfaffman (1941); Smith (1951); Johnson (1952); Crowell (1954b); Stanton (1960); Stanley (1966); Sage (1973; 1975); and Kooser (1980). Originally, strata assigned to this formation were known as the Martinez Formation (Merriam, 1897), however, it was renamed the Francisquito Formation by Dibblee (1967). The largest outcrop area, 10 km north of the town of Castaic, was designated the type area. The San Francisquito Formation is about 1800 m (5900 ft) thick, predominantly marine, clastic sequence of Cretaceous to Paleocene age (Kooser, 1980; this volume). Its age is based on Cretaceous and Paleocene marine molluscs, in the lower and upper parts of the formation respectively.

Mint Canyon Formation

The Mint Canyon Formation crops out as a thin 50 m thick tongue of reddish sandstone and conglomerate in the extreme southeastern part of Ridge Basin where it unconformably overlies the San Francisquito Formation and is overlain unconformably by the Castaic Formation (Fig. 1). In Soledad Basin to the southeast it lies on Miocene Tick Canyon or Olegocene and lower Miocene Vasquez Formations and reaches a thickness of at least 1220 m (4000 ft) (Jahns, 1940).

Hershey (1902b) first described the unit and Kew (1924) later named it the Mint Canyon Formation for outcrops in the Tick Canyon area of Soledad Basin. In Ridge Basin the formation was first mapped by Clements (1929) and later by Crowell (1954b). Fossils include brackish-water foraminifera, freshwater ostracodes, and vertebrate remains (Stitt, 1980). The age of the Mint Canyon Formation is middle to late Miocene (Barstovian to Clarendonian land mammal ages). The age assignment has been somewhat controversial (see Winterer and Durham, 1962) because the Mint Canyon Formation contains some early Pliocene vertebrates. The invertebrate fauna indicates a late Miocene age and the overlying marine Castaic Formation has been dated as late Miocene on the basis of an abundant marine mollusc assemblage (Stanton, 1966). The Mint Canyon Formation is coeval with much of the marine Modelo Formation of the Ventura Basin (Relizian, Luisian and Mohnian foraminiferal stages of Kleinpell, 1938) and with the Miocene part of the Caliente Formation of the Lockwood Valley area (Crowell, 1960, 1962; Carman, 1954, 1964; Ehlig et al., 1975; Stitt, 1980). Originally, the Mint Canyon and Caliente Formations were probably adjacent and later separated by right slip along the San Gabriel Fault.

Castaic Formation

The Castaic Formation was named by Crowell (1954b) for a 520 to 2800 m (1700 to 9200 ft) thick sequence of predominantly mudstone interbedded with siltstone, sandstone, and conglomerate which contains marine mollusks and foraminifers of late Miocene age exposed in Castaic Canyon north of the town of Castaic (Fig. 2). Castaic Lake now covers part of the type area. The lateral extent of the formation in the basin is about 10 km (6 mi). Earlier workers referred to these strata as "Modelo" because they occur at the same stratigraphic position and are about the same age (late Miocene) as the Modelo Formation of Ventura Basin. The Castaic Formation is, in large part, contemporaneous with the Modelo Formation of the eastern Ventura Basin, but its lithologic character is sufficiently distinctive to justify a separate name. The Castaic Formation unconformably overlies the Mint Canyon Formation and is conformably overlain by and interfingers with the Ridge Basin Group. South of Ridge Basin, the Castaic Formation is unconformably overlain by the Pico Formation in the subsurface, and, north of where the Pliocene and Pleistocene Pico Formation wedges out, by the Pliocene and Pleistocene Saugus Formation (Stitt, 1980). The Castaic Formation appears to be restricted to southern Ridge Basin and the northwestern part of Soledad Basin (Stanton, 1967).

The age of the Castaic Formation is late Miocene (Jacalitos megafossil stages after Ensley, 1980; and Mohnian and Delmontian microfossil stages after Skolnick and Arnal, 1959). Marine molluscs, brachiopods, arthropods, crustaceans, and vertebrates have been reported by Eaton (1939), Durham (1948a, 1948b), White and Buffington (1948), Wright (1948), Stanton (1960, 1966, 1967), Smale (1978), and Young (1980), and most are indicative of shallow-marine environments. Stanton (1960, 1966) indicates the basal Castaic Formation on the east was deposited during a transgressive marine event, whereas deposits in the middle part of the basin were deposited at water depths between 45 to 90 m (150 to 300 ft). Benthic foraminifer assemblages suggest moderately deep water and restricted circulation (Skolnick and Arnal, 1959).

Ridge Basin was born in Castaic time and formed a restricted marine basin behind the San Gabriel fault scarp. The deep end of the basin was to the south where it drained into Ventura Basin near Newhall-Saugus (Crowell, 1954b; Stitt, this volume).

Ridge Basin Group

The Ridge Basin Group is 7000 to 11,000 m (23,000 to 36,000 ft) thick and consists of four formations, three of which interfinger. From east to west these are the Ridge Route and Peace Valley Formations and the Violin Breccia (Fig. 1). These units are conformably overlain by and interfinger with the Hungry Valley Formation, which is also designated as part of the Ridge Basin Group, and conformably overlie and interfinger with the Castaic Formation below. The Ridge Route and Peace Valley Formations are subdivided each into five members, which are described in more detail and interpreted in later chapters. Based on faunal and stratigraphic evidence the lower 300 to 600 m (1000 to 2000 ft) of

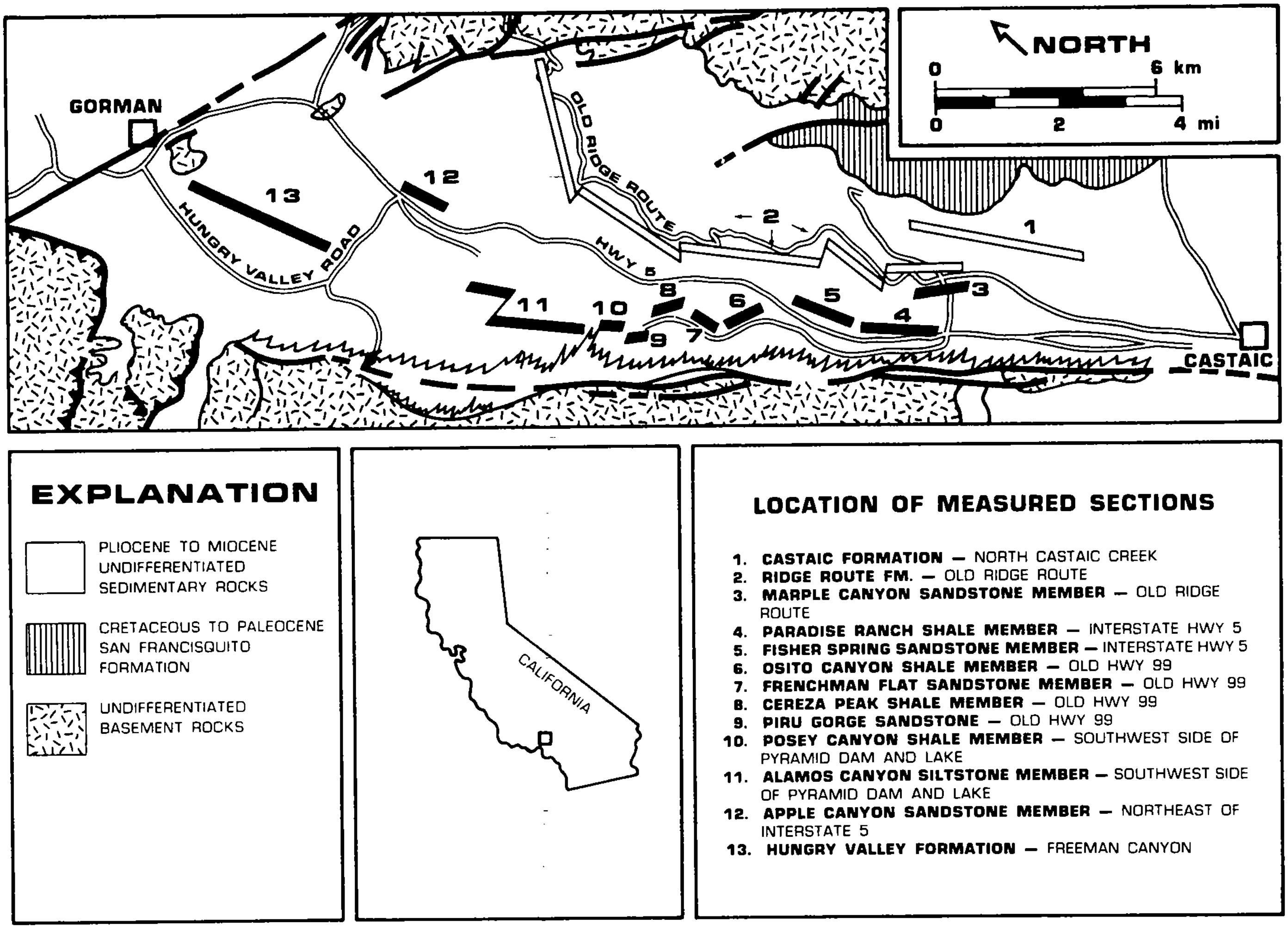


Figure 2. Index map showing location of type sections for various units in Ridge Basin.

the Ridge Basin Group has been interpreted as marine, whereas the remainder (8400-8700 m) (27500-28500 ft) has been interpreted as dominantly lacustrine and fluvial-alluvial fan. Originally, Clements (1932, 1937) named the Ridge Route Formation; Crowell named the Violin Breccia (1954b) and informally introduced the term Peace Valley "beds" for strata underlying the Hungry Valley Formation (1950). Axelrod (1950) named the Ridge Basin Group. Most of the member names for the Peace Valley and Ridge Route Formations were originally proposed by Crowell (written comm., Cohee, 1952) and subsequently used informally by Shepard (1961) and Irvine (1977). Formal type sections for each unit are presented here.

Violin Breccia

The Violin Breccia crops out along the San Gabriel Fault on the western margin of Ridge Basin as a narrow continuous strip for approximately 29 km (18 mi) (Fig. 1). It is between 400-1500 m (1300-4920 ft) wide and has a composite thickness at the surface of over 11,000 m (36000 ft). It can be traced another 1340 m (4400 ft) in the subsurface south of Castaic (Stitt, 1980). This remarkable unit is laterally and vertically transitional into the Castaic, Ridge Route, Peace Valley, and lower Hungry Valley Formations (Fig. 1). Clements (1929) first described the unit and subsequent workers (Clements, 1932, 1937; Eaton, 1939; Chambers, 1947; Bettinger, 1948; and Harris, 1950) continued to map the unit, giving it a variety of names. Finally, Crowell (1954b) named the Violin Breccia for exposures on a divide between Violin and Palomas Canyons. Its lower contact is in the subsurface so that its position is not known precisely; its upper contact is conformable with the middle part of the Hungry Valley Formation. It is flanked on the southwest by the San Gabriel Fault and everywhere grades laterally into units in the axial part of Ridge Basin.

Fossils from the Violin Breccia, and especially from interdigitations with laterally equivalent units, include rare vertebrate remains, wood fragments, ostracodes, and mollusks. The age of the Violin Breccia is inferred to be late Miocene (early Mohnian) to Pliocene (Blancan?). The early Mohnian age is based on interfingering molluscs assemblages of the Castaic Formation in the extreme southwestern part of the basin (Eaton, 1939; Crowell, 1952a, 1954b; Stanton, 1966; Stitt, 1980). The Blancan? age is inferred from vertebrate assemblages in the lower part of the Hungry Valley Formation and the upper part of the Peace Valley Formation which is laterally transitional into the Violin Breccia (Crowell, 1950; Miller and Downs, 1974).

Ridge Route Formation

The Ridge Route Formation was named by Clements (1929) for outcrops along the upper 26 km (16 mi) of the Old Ridge Route in the eastern part of the basin. The type section extends along the Old Ridge Route from south of Templin Highway to Liebre Mountain. It conformably overlies the Castaic Formation and is conformably overlain by the Hungry Valley Formation (Fig. 1). It is over 9000 m (29,500 ft) thick and consists predominantly of sandstone, conglomeratic sandstone, and mudstone. The Ridge Route Formation, undifferentiated along the Old Ridge Route, extends laterally into the basin for 150 to 5180 m (500 to 17000 ft). Toward the northeast it coarsens in average grain size, the maximum clast size increases, and the finer-grained strata decrease in abundance. Toward the southwest the Ridge Route Formation interfingers with mudstone and shale of the Peace Valley Formation, and five major clastic tongues cross the basin and interfinger with the Violin Breccia. Many smaller tongues and beds pinch out in the Peace Valley Formation. Each of the five clastic tongues thickens slightly toward the central part of the basin and then thins toward the southwest into the Violin Breccia (Fig. 1). Several of these tongues also thin toward the northeast where they merge with the Ridge Route Formation. These clastic tongues include from oldest to youngest: Marple Canyon, Fisher Spring, Frenchman Flat, Piru Gorge, and Apple Canyon Sandstone Members (Fig. 1). The age of Ridge Route Formation ranges from late Miocene to Pliocene (Crowell, 1950; Stanton, 1960; Miller and Downs, 1974; Ensley, 1980). Each of these members is discussed and its depositional environment interpreted in later chapters.

Marple Canyon Sandstone Member (new name). The Marple Canyon Sandstone Member is 420 m (1400 ft) thick at its designated type section along the Old Ridge Route. It is named for exposures on the north side of Marple Canyon where the Old Ridge Route crosses these outcrops at the head of Marple Canyon, between 1.0 km south and 0.5 km north of the intersection between the Old Ridge Route and Templin Highway. It consists of brown and white arkosic sandstone interbedded with mudstone and conglomerate. It conformably overlies and interfingers with the Castaic Formation and is overlain conformably by the Paradise Ranch Shale Member (Fig. 1). It ranges in thickness from less than 80 m (265 ft) thick along the eastern and western margins of the basin to over 450 m (1475 ft) thick in the axial part of the basin and reaches a maximum of 1100 m (3500 ft). It interfingers with the Violin Breccia to the southwest and its lateral extend into the basin is about 8000 m (26500 ft).

The age of the Marple Canyon Sandstone Member is late Miocene (late Mohnian) based on its stratigraphic relationship with the marine Castaic Formation. Freshwater mollusks, ostracodes, charophytes, and vertebrate remains occur in the upper part and were not diagnostic of age (Young, 1980). The marine/nonmarine boundary in this unit occurs near the junction of Templin Highway and the Old Ridge Route based on faunal and stratigraphic evidence (Smale, 1978; Young, 1980). Harris (1950), Dehlinger (1950, 1952), Crowell (1954b), Stanton (1960), Smale (1978), and Ensley (1980) have all mapped a contact between the Castaic Formation and the Marple Canyon Sandstone Member. However, since the Marple Canyon interfingers with the Castaic Formation and the sandstone lithologies are nearly identical, selection of the contact is somewhat arbitrary. The base of the Marple Canyon Sandstone Member at its type section is here selected at the north side of the head of Marple Canyon along the Old Ridge Route.

Fisher Spring Sandstone Member (new name). The designated type section of the Fisher Spring Sandstone Member is along the east side of Interstate Highway 5 between Paradise Ranch and Whitaker Peak road (Fig. 2). It is 740 m (2500 ft) thick here and is named for Fisher Spring which is adjacent to these outcrops (Harper, 1981). It consists of light brown, medium-grained arkosic sandstone interbedded with mudstone and conglomerate. It ranges in thickness from less than 100 m (325 ft) thick along the western margin of the basin to over 1300 m (4300 ft) thick on the eastern side. Its lateral extent into the basin is 4310 m (14200 ft) and it conformably overlies the Paradise Ranch Shale Member and is conformably overlain by the Osito Canyon Shale Member. This tongue of clastic rock thins to the southwest where it interfingers with the Violin Breccia and thickens to the northeast where it merges with the Ridge Route Formation. The age of the Fisher Spring, based upon its stratigraphic position in the basin, indicates that it is late Miocene in age. Freshwater mollusks, ostracodes, and vertebrate remains were not diagnostic of age, but suggestive of nonmarine lacustrine-fluvial conditions of deposition (Young, 1980; Harper, 1981).

Frenchman Flat Sandstone Member (new name). At its designated type section, along Highway 99 near Frenchman Flat the Frenchman Flat Sandstone Member is 172 m (565 ft) thick (Hollywood, 1981) (Fig. 2). It consists of white medium-grained arkosic sandstone interbedded with mudstone and rare conglomerate. Its lateral extent into the basin is 4620 m (15200 ft) and it conformably overlies the Osito Canyon Shale Member and is overlain conformably by the Cereza Peak Shale Member. Laterally, it interfingers with the Ridge Route Formation to the northeast and with the Violin Breccia to the southwest. It ranges in thickness from less than 50 m (165 ft) thick along the western margin and 100 m (330 ft) thick on eastern margin of the basin to over 400 m (1300 ft) thick in the axial part of the basin. The age of the Frenchman Flat Sandstone Member is late Miocene based on its stratigraphic position just below the Cereza Peak Shale Member. Freshwater mollusks, plant remains, ostracodes, and stromatolites are the only fossils found and are not diagnostic of age (Young, 1980).

Piru Gorge Sandstone Member (re-defined name). The Piru Gorge Sandstone Member is 185 m (600 ft) thick in Piru Gorge along Highway 99 near Pyramid Dam (Fig. 2). Originally, Axelrod (1950) named the Piru Gorge Sandstone and designated it as a formation of the Ridge Basin Group. It is re-defined here as a member because it is one of five clastic tongues of the Ridge Route Formation. Its lateral extent into the basin is 5540 m (18200 ft) and its thickness ranges from 50 to 275 m (165 to 900 ft) thick. It conformably overlies the Cereza Peak Shale Member and is conformably overlain by the Posey Canyon Shale Member. Laterally, the Piru Gorge Sandstone Member interfingers with the Violin Breccia to the southwest and the Ridge Route Formation to the northeast (Fig. 1). Plants, ostracodes, and fish remains are found in the middle shale interval of the Piru Gorge Sandstone Member. Originally, Axelrod (1950) considered the plants to be middle Pliocene, however, Axelrod (this volume) now considers the assemblage to be late Miocene in age. Ensley (1980) dates the Piru Gorge Sandstone Member at 6.07 m. y. using magnetostratigraphy.

Apple Canyon Sandstone Member (new name). The Apple Canyon Sandstone Member is the uppermost clastic tongue of the Ridge Route Formation and its designated type section is on the east side of Highway 5, beginning at the intersection between Hungry Valley Road and Interstate Highway 5. The area underlain by the member mainly lies on the east side of Highway 5. It consists of light brown, medium-grained arkosic sandstone interbedded with mudstone, conglomerate, and minor limestone. It is 1000 m (3280 ft) thick here and reaches a maximum of 1130 m (3700 ft) and its lateral extent into the basin is 8000 m (26250 ft) (Wood, 1981). It conformably overlies the Alamos Canyon Siltstone Member and is conformably overlain by the Hungry Valley Formation. Laterally, the Apple Canyon Sandstone Member interfingers with the Ridge Route Formation to the northeast and the Violin Breccia to the southwest. Fossils include freshwater mollusks, ostracodes, stromatolites, and plant and vertebrate remains. The Kinsey Ranch fauna of Miller and Downs (1974) and vertebrate locality of Crowell (1950) occur just below this unit. A Hemphillian age (late Miocene) is suggested by Miller and Downs (1974) for these vertebrate remains. Ensley (1980) suggests an age of 5.5 to 5.0 m. y. for this fauna based on a comparison with similar vertebrate assemblages and magnetostratigraphic data. Other fossils were not diagnostic of age.

Peace Valley Formation (re-defined). The Peace Valley Formation is named for outcrops along Highway 99 in Peace Valley, Piru Creek, Osito Canyon, and Big Oak Flat. Originally, the term Peace Valley "beds" (Crowell, 1950) was used to denote finer-grained rocks in the northern half of the Ridge Basin, but since the base of the sequence had not yet been studied, the beds were not given formational status. Later Crowell (1964) informally raised the unit to formational status and Dibblee (1967) formalized the term Peace Valley Formation. Here, the Peace Valley "beds" are raised formally to formation status and its type section is designated to be along Highway 99 from Templin Highway to Hungry Valley Road. Since Pyramid Lake now covers part of Highway 99, outcrops along the west shore of Pyramid Lake constitute the continuation of the section. The formation is about 5000 m (16200 ft) thick and consist of shale and mudstone with minor siltstone and sandstone interbeds. It conformably overlies the Marple Canyon Sandstone Member and is conformably overlain by the Apple Canyon Sandstone Member of the Ridge Route Formation (Fig. 1). The Peace Valley Formation interfingers with the Violin Breccia to the southwest and the Ridge Route Formation to the northeast. It is separated into five members by the clastic tongues of the Ridge Route Formation described above, which locally cross the basin and interfinger with the Violin Breccia to the southwest. Each member of the Peace Valley Formation thickens toward the axis of the basin and pinches out toward the northeast and southwest. These members include from oldest to youngest: Paradise Ranch, Osito Canyon, Cereza Peak, and Posey Canyon Shale Member, and Alamos Canyon Siltstone Member. The age of the Peace Valley Formation ranges from late Miocene to early Pliocene (Crowell, 1950; Stanton, 1960; Miller and Downs, 1974; and Ensley, 1980). None extends outside of the basin.

Paradise Ranch Shale Member (new name). The Paradise Ranch Shale Member is the lowest shale body of the Peace Valley Formation (Fig. 1). At its designated type section along Interstate 5 from Templin Highway north to Three Mile Grade, it is 549 m (1800 ft) thick (Smith, 1981). Paradise Ranch is on the west side of Interstate 5 at the middle of the section. The type section consists of poorly exposed dark gray shale with thin interbeds of sandstone. It conformably overlies the Marple Canyon Sandstone Member and is overlain conformably by the Fisher Spring Sandstone Member. Laterally, it interfingers with the Ridge Route Formation to the northeast and the Violin Breccia to the southwest, with a lateral extent of 5625 m (18450 ft). No fossils have been found in the Paradise Ranch Shale Member and its age is inferred to be late Miocene based on its stratigraphic position above the marine Castaic Formation and the Marple Canyon Sandstone Member of the Ridge Basin Group.

Osito Canyon Shale Member (new name). The Osito Canyon Shale Member is 976 m (3200 ft) thick and consists of gray shale, mudstone, and interbeds of siltstone and sandstone. Its designated type section is along Osito Canyon adjacent to Highway 99 (Smith, 1981). It conformably overlies the Fisher Spring Sandstone Member and is conformably overlain by the Frenchman Flat Sandstone Member. Laterally, it extends for 6154 m (20200 ft) and interfingers with the Ridge Route Formation to the northeast and with the Violin Breccia to the southwest. No fossils have been found in the Osito Canyon Shale Member. Its age is inferred to be late Miocene based on its stratigraphic position above the late Miocene Castaic Formation (Stanton, 1960) and below the late Miocene Cereza Peak Shale Member (Axelrod, this volume).

Cereza Peak Shale Member (new name). The Cereza Peak Shale Member is 914 m (3000 ft) thick with its type locality at the base of Cherry Peak east of Highway 99 (Shephard, 1961; Smith, 1981). "Cereza" is Spanish for "cherry" and is used here because the name Cherry Peak is preoccupied as a stratigraphic name. It consists of dark gray shale and blue gray mudstone with minor sandstone interbeds. The member is up to 940 m (3100 ft) thick and its lateral extent is 5540 m (18200 ft). It conformably overlies the Frenchman Flat Sandstone Member and is conformably overlain by the Piru Gorge Sandstone Member. Laterally, the Cereza Peak interfingers with the Ridge Route Formation to the northeast and with the Violin Breccia to the southwest. Plants, ostracodes, and fish remains are found in it. Axelrod (1950; this volume) collected sixty-nine floral specimens representing 13 species along Highway 99 in the upper part of the Cereza Peak Shale Member. He considers this floral assemblage to consist of woodland, chaparral, and borderland-forest vegetation, representative of a semi-arid to subhumid climate. A middle Pliocene age was originally assigned the Cereza Peak Shale Member by Axelrod (1950), but he (this volume) now considers it to be late Miocene in age. The ostracodes and fish were not diagnostic of age.

Posey Canyon Shale Member (new name). The Posey Canyon Shale Member is 1000 m (3280 ft) thick at Pyramid Dam (Irvine, 1977) and consists of indurated gray shale, mudstone, dolomicrite, pyritic shale and siltstone, and gypsiferous siltstone. This member was originally named for outcrops along Posey Canyon but this area is now under Pyramid Lake which formed behind the dam built in 1970. Its designated type section is along the west side of Pyramid Dam and adjacent terrain where it conformably overlies the Piru Gorge Sandstone Member. It is conformably overlain by the Alamos Canyon Siltstone Member and separated from it by a distinctive tuff horizon. Laterally, the Posey Canyon interfingers with the Violin Breccia to the southwest and the Apple Canyon Siltstone Member and Ridge Route Formation to the northeast. Its lateral extent is 2700 to 3600 m (8860 to 11800 ft) and it is approximately 730 to 1000 m (2400 to 3280 ft) thick. It forms a triangle-shaped outcrop adjacent to the San Gabriel Fault. Indurated shale and mudstone comprise 460 m (1500 ft) of the member, but this induration rapidly pinches out eastward along strike, making the total lateral extent of the indurated strata only 1500 to 1800 m (4900 to 5900 ft) (Irvine, 1977). Plant, fish, ostracode, and mollusk remains are found in the Posey Canyon (Shepard, 1961; Irvine, 1977; Smith, 1981). The plant fragments are similar to collections of late Miocene age obtained by Axelrod, (1950) from the Piru Gorge and Cereza Peak Shale Members. The fish, ostracode, and mollusk remains were not diagnostic of age.

Alamos Canyon Siltstone Member (new name). The Alamos Canyon Siltstone Member is 1500 m (4920 ft) thick along the shoreline of Pyramid Lake, but exceeds this amount to the northeast. It consists of gray-green shale, mudstone, and siltstone, with minor brown sandstone layers. It was named originally for outcrops exposed along Canada de Los Alamos which is now partially covered by Pyramid Lake. Its designated type section is along the western side of Pyramid Lake from north of the tuff horizon to the Piru Creek arm of Pyramid Lake and then across the lake and along the west side of Canada de Los Alamos. It conformably overlies the Posey Canyon Shale Member and is conformably overlain by and interfingers northeastward into the Apple Canyon Sandstone Member. To the southwest, the Alamos Canyon interfingers with the Violin Breccia. Its maximum lateral extent is 6100 m (20000 ft). The age of the Alamos Canyon is late Miocene based on vertebrate remains (Crowell, 1950; Miller and Downs, 1974) at the top of the member and plant remains from the underlying units (Axelrod, 1950, this volume; Crowell, 1950). Ensley (1980) suggests a 5.0 to 5.5 m. y. age for the Kinsey Ranch vertebrate fauna of Miller and Downs (1974), which comes from the member.

Hungry Valley Formation

The Hungry Valley Formation is the youngest formation in the Ridge Basin Group (Fig. 1). Its type section is just east of Hungry Valley, along Freeman Canyon (Crowell, 1950) where it is 1220 m (4000 ft) thick, and consists of five informal members (Fig. 2). The major lithologies are white conglomeratic sandstone, brown sandstone, and minor interbedded gray and brown siltstone. Its maximum thickness is over 1900 m (6250 ft) with a lateral extent of 10,000 m (32800 ft). This formation conformably overlies and interfingers with the Apple Canyon Sandstone Member and the Violin Breccia. In the northwest corner of the basin, the Hungry Valley Formation overlaps the Violin Breccia and the San Gabriel fault zone, indicating that the fault and breccia deposition ceased to be active during the latter part of Hungry Valley deposition. The youngest exposed beds of this formation crop out along the San Andreas Fault and the Frasier and Dry Creek Thrusts in the northwestern corner of the map. The Hungry Valley formation is overlain by terrace and alluvial deposits of Quaternary age. The age of the Hungry Valley formation is late Miocene to early Pliocene (late Hemphillian or early Blancan? based on vertebrate remains in the lower part of the formation and units just below (Crowell, 1950; Miller and Downs, 1974). Ensley (1981) suggests a 5.0-5.5 m. y. age based on magnetostratigraphic correlations for the beginning of Hungry Valley deposition.

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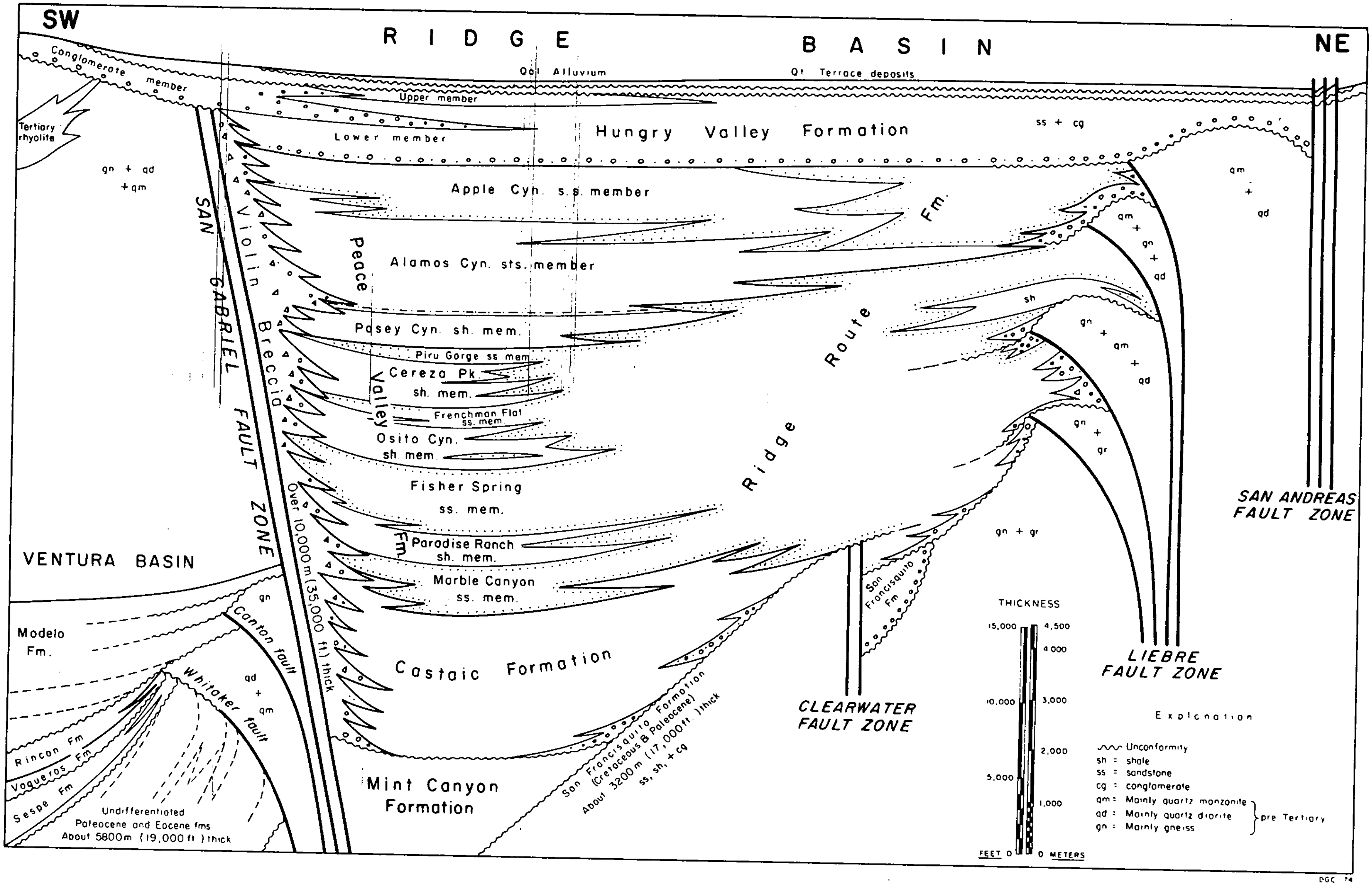

Stratigraphic arrangement, Ridge Basin, southern California.

THE TECTONICS OF RIDGE BASIN, SOUTHERN CALIFORNIA

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ABSTRACT

Ridge Basin originated during the Miocene Epoch as a stretched and sagged crustal wedge within the complex splintered boundary between the Pacific and North American lithospheric plates. Strata filling the basin and overlap relations around its margins document a history of displacement on bordering faults at different times during the late Cenozoic Era. The San Gabriel Fault, the early strand of the San Andreas transform system in this region, began activity about 12 m. y. ago and formed the straight southwestern border of Ridge Basin until about 5 m. y. ago. On the east, first the Clearwater Fault was active, and died about 8 m. y. ago. Successive strands of the Liebre fault zone then took over, also on the east, but ceased displacing between 6 and 5 m. y. ago. These faults died before the San Gabriel. Transform displacements were then wholly taken over by the San Andreas Fault, beginning 5 or 6 m. y. ago. Between about 4 and 2 m. y. ago, the Frazier Mountain thrust system resulted from shortening in the terrane south of the Big Bend in the San Andreas Fault. Uplift and erosion in the Pleistocene, and still continuing, have exposed geological relations that allow an unusual documentation of the interplay between tectonics and sedimentation, including the timing of these fault movements.

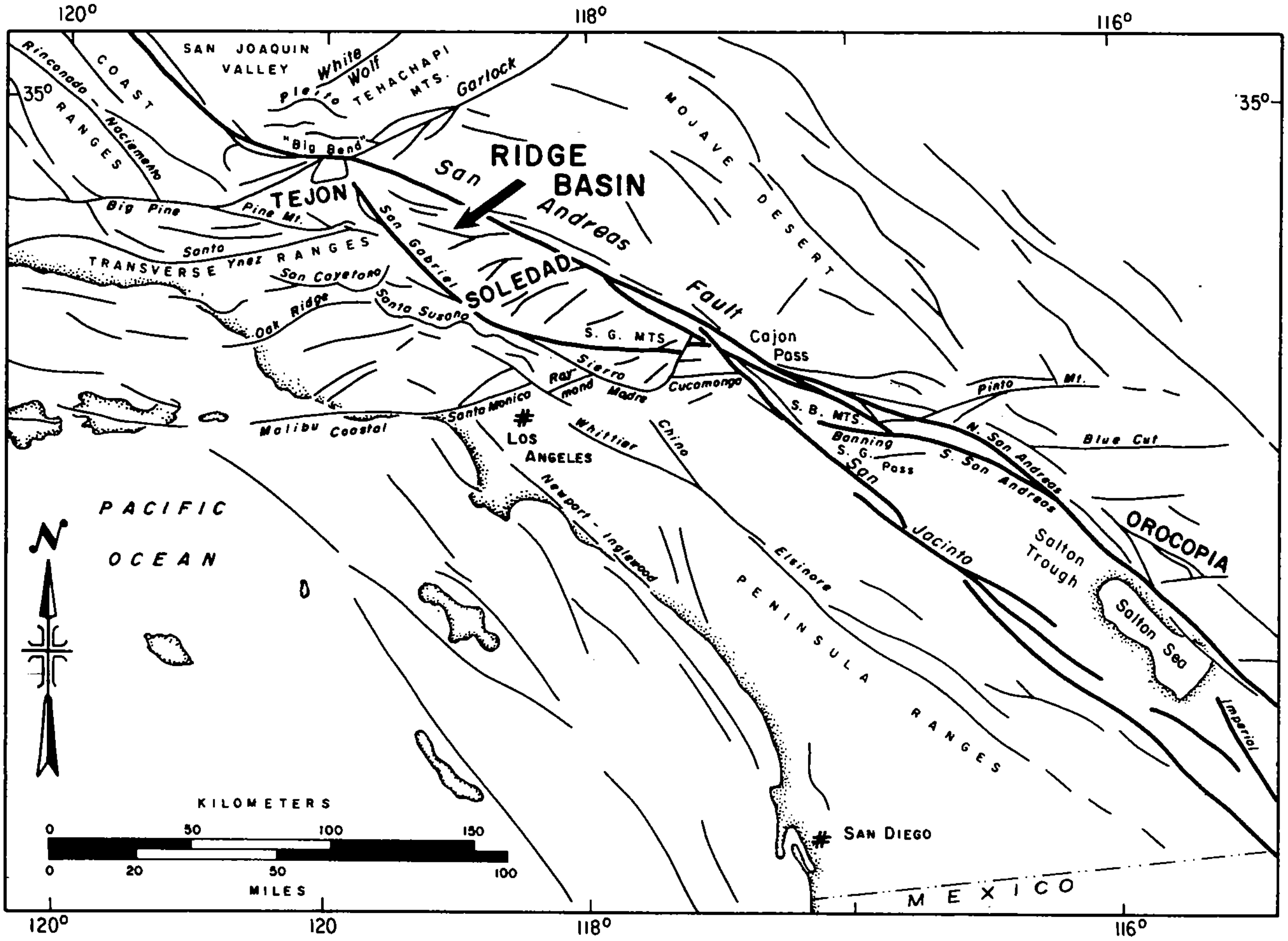

Figure 1. Location and structural setting of Ridge Basin, southern California.

in Crowell, J.C., and Link, M.H., eds., 1982, **Geologic History of Ridge Basin, Southern California**, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 25-42.

Figure 2. Looking northwestward along the San Gabriel Fault from near Castaic, with Ridge Basin on the right (east). Hungry Valley and Frazier Mountain in the distance. Photograph taken in 1933 by Spence Air Photos. Used with permission.

INTRODUCTION

Ridge Basin developed during the Miocene Epoch within the San Andreas fault system, the transform boundary between the Pacific and North American lithospheric plates. Its situation with respect to the many faults in southern California and near the Big Bend of the San Andreas Fault in the central Transverse Ranges is shown in Figure 1. The basin itself formed adjacent to the San Gabriel Fault which bounded the depositional site on the southwest and was a principal strand of the San Andreas system between about 12 and 5 m. y. ago. To the east, several faults separated the basin from uplands and have been successively overlapped as sediments infilled the tectonic depression of Ridge Basin. At present, Ridge Basin is truncated on the north by the modern San Andreas Fault and on the northwest by the Frazier Mountain thrust system. Late Cenozoic deformation and uplift of the region, still continuing, has resulted in deep erosion and disection so that the relations between tectonic deformation and sedimentation through time are unusually well displayed.

About 13,400 m (44,000 ft) of beds were deposited within Ridge Basin, a narrow trough lying between the San Gabriel fault scarp on the southwest and highlands uplifted on the east in part along the Clearwater and Liebre fault zones (Fig. 2). As documented below, this huge thickness of clastic strata, both marine and nonmarine, was laid down as the depocenter moved toward the northwest so that the beds overlap the basin floor in that direction. Deformation during the continuous deposition of strata within the central trough was accompanied by both facies and overlap relations around the margins of the basin. By early Pliocene time the basin became completely filled, and was then uplifted, deformed, and truncated on the north by the modern San Andreas Fault. Previous papers dealing with the tectonics of Ridge Basin include Clements (1937), Eaton (1939), Crowell (1950, 1952a, 1954a,b, 1962, 1974, 1975a,b, 1981).

The main structural features of Ridge Basin, described below, are diagramed in Figure 3, very much simplified from Crowell, et al., (1982), the geologic map.

SAN GABRIEL FAULT

On the surface the San Gabriel Fault can be traced for 130 km (80 mi) from the northwestern part of Ridge Basin to within the San Gabriel Mountains and on eastward into the complex Cajon Pass - San Bernardino region. Between Ridge Basin and the San Gabriel Mountains the fault zone crosses the Castaic lowlands where its course is marked by a belt of braided small faults and steep dips in young beds. Since most of the displacement on the zone took place before these Pliocene and Quaternary beds were deposited, however, its trend here is not nearly as conspicuous as within basement rocks, or along the southwestern margin of Ridge Basin. Within the western San Gabriel Mountains the San Gabriel Fault branches into two major faults. The southern branch extends on eastward and becomes enmeshed with the Sierra Madre and Cucamonga faults that front the range. The northern branch is offset by younger faults before meeting the San Jacinto and perhaps the San Andreas faults. The straight course of the fault zones across the rugged mountainous terrane shows that its dip is nearly vertical. To the southeast of the Castaic lowlands, the fault zone is described by Oakeshott (1958), Ehlig (1966, 1973, 1981), Jahns (1973), Morton (1975), and by others.

At the margin of Ridge Basin near Castaic the San Gabriel fault zone dips nearly vertically and separates Violin Breccia from Modelo Formation. In the central part of Ridge Basin, however, (for example, down Piru Creek below Frenchman Flat), it dips 70° to the northeast. Complexities within the San Gabriel fault zone along the border of Ridge Basin, especially within the crushed belt of basement rocks, are oversimplified on the map (Crowell, et al., 1982). The gneisses, migmatities, and granitic rocks of several types have been retrograded so that chlorite and laumontite (identified by T. H. McCulloh) are conspicuous, and especially along joints and seams. The zone consists mainly of long attentuated slices of these rocks where most observable contacts between the rocks types are faults with fault breccia and gouge zones along them. Foliations within the banded rocks have been the sites of remobilization as shown by slickensides. Deformation within the basement rocks suggests that movement on the fault zone began before that on the fault now truncating the Violin Breccia. Most exposures of the latter fault, however, show that the Violin Breccia itself has been faulted and brecciated, so that many of these movements are later than those involved in the deposition of the Violin Breccia beginning about 12 m. y. ago.

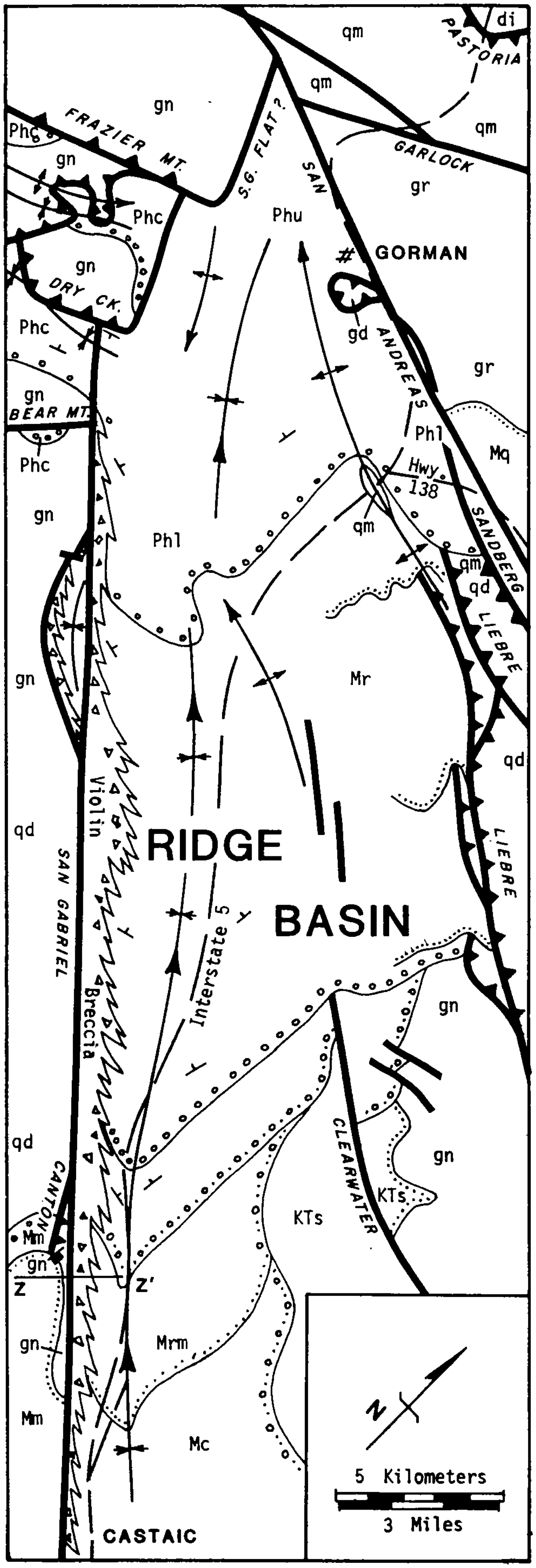

Figure 3. Simplified structural map of Ridge Basin.

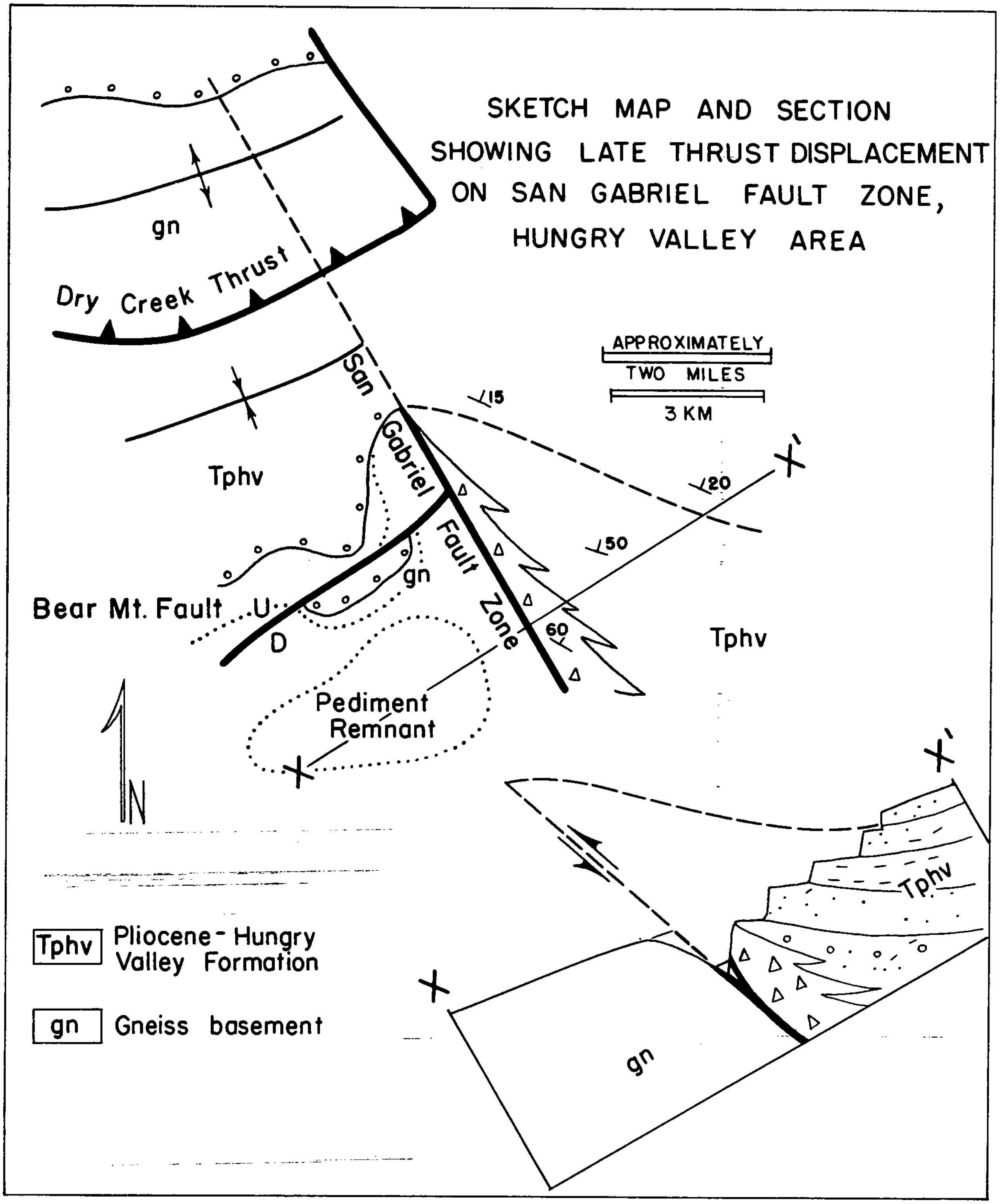

Figure 4. Diagrammatic map and cross section showing late thrust displacement on San Gabriel fault zone, Hungry Valley area.

Farther northwest (Crowell, et al., 1982, cross sections 9 and 10) later movement along the fault zone has repeated the principal strand of the San Gabriel Fault, the belt of Violin Breccia, and units of the Peace Valley Formation. The older and major fault, exposed to the southwest, separates basement terrane from Ridge Basin sedimentary rocks and is bordered by a strip of Violin Breccia. This older strand of the fault, which dips on the average about 70° to the northeast, is cross-cut by later minor faults and has been bent so that in map view the trace of the fault is arcuate. In this region, the northeastern and younger strand of the San Gabriel Fault is a straight-striking fault that dips about 65° to the southwest (Crowell, et al., 1982, Cross section 10). By means of matching isopachs of stratigraphic units across this fault, between the main body of Ridge Basin sedimentary rocks on the northeast and those within the wedge, an approximate right slip of 2500 m (8000 ft) and dip slip of 1800 m (6000 ft), down on the southwest, is suggested. Much larger displacements, however, took place on the older and deformed fault to the southwest.

The San Gabriel fault zone quite certainly trends on northwesterly at depth beyong the point of overlap by the Hungry Valley Formation to join the San Andreas fault zone on to the northwest in about 16 km (10 mi) (Crowell, 1950, p. 1641). Its course is interpreted as lying beneath the Frazier Mountain thrust system. In fact, this system has probably displaced shallower parts of the fault upwards and southeastward, with 15° and 25° of clockwise rotation respectively, to account for the straight boundary faults on the eastern sides of the Frazier Mountain and Dry Creek thrust systems (Fig. 3). They are considered as exhumed faultline scarps where first tectonic uplift and then erosion are involved in the exhumation.

Relationships near the northwestern end of the exposed San Gabriel fault zone require at least two distinct episodes of displacement (Crowell, 1950, p. 1643). The principal fault is overlapped by units of the Hungry Valley Formation and in this vicinity as well the Violin Breccia is also overlapped and disappears from view at the surface. The overlapping conglomerate beds of the Hungry Valley Formation lie unconformably upon a pediment surface cut on basement rocks to the southwest of the fault zone, but are part of the conformable Ridge Basin stratigraphic section to the northeast within Ridge Basin (Fig. 4). These same overlapping beds are cut and downdropped on the southeastern flank of Bear Mountain by the Bear Mountain Fault (Crowell, 1950, Figs. 4, 11). The latter fault trends to the northeast and ends against the San Gabriel Fault and at first glance appears to be truncated by it and therefore older. We therefore have field relations where the basal conglomerate is younger than displacement on the San Gabriel Fault but older than that on the Bear Mountain Fault, yet the latter fault is apparently truncated by the San Gabriel Fault! These perplexing relationships, however, are explained by appealing to renewed movement along the San Gabriel fault zone to the south of where it is intersected by the Bear Mountain Fault. This renewed movement is interpreted as associated with the younger displacement along the fault strand bordering the slice on the northeast, discussed above. The renewal has downdropped the basement terrane south of the corner where Bear Mountain and San Gabriel faults intersect and has raised the band of Violin Breccia in this region with respect to the basement by reverse dip slip. This late displacement is also interpreted as accompanying folding within Ridge Basin, including the steepening of dips in the Violin Breccia and Hungry Valley beds near the San Gabriel Fault.

At one place along the San Gabriel fault zone a mass of diorite about 300 m by 120 m (1000 by 400 ft) in outcrop area is interbedded within the Violin Breccia and abuts against the fault zone on the southwest (Shepard, 1962, p. 1941). Sedimentary breccia and conglomerate can be traced beneath the mass as well as across its top. It is therefore interpreted as constituting a landslide mass that was squeezed from within the San Gabriel fault zone during the Miocene, or one that slid across the zone from its source on the southwestern side. It is made up of diorite and gabbro with stringers of anorthosite and is very similar to terrane in place near the mouth of Bear Gulch on the flank of Bear Mountain (Ehlig and Crowell, this volume).

Both the mass of diorite and the terrane at Bear Gulch are similar in turn to gabbros, diorites, and anorthosite of the western San Gabriel Mountains (Higgs, 1954; Crowell and Walker, 1962; Crowell, 1960, 1962, 1975a,b, 1981; Ehlig and Crowell, this volume). About 60 km (38 mi) of right slip is apparently required to bring the Bear Gulch and San Gabriel terranes back together, the amount of total displacement on the fault. As yet, however, large-scale mapping in the rugged terranes involved, accompanied by petrographic, geochemical, and isotopic investigations, has not been done to prove this displacement unequivocally.

The amount and type of displacement on the San Gabriel fault zone in Ridge Basin and Castaic lowlands have been matters of controversy for many years (Paschall and Off, 1961; Weber, 1979). In my view the controversy should no longer flourish since considerable regional information has come in during the last quarter-century, including a realization of the mobility of crustal blocks along the splintered San Andreas transform belt. Concepts that have evolved from plate tectonics have aided in understanding how such lateral faulting operates kinematically, and how major strike-slip faults terminate. Nonetheless, the proof of major post-mid-Miocene right slip of 60 km on the San Gabriel Fault depends on properly interpreting details in both the local and regional geology, and hinges on relations at the southern end of the Ridge Basin region.

In this critical area near Castaic the interpretation of great right slip on the San Gabriel Fault depends on the recognition of a mismatch in the inferred sedimentation history of two conglomerate masses, the Violin Breccia on the northeast and Modelo conglomerates on the southwest. This mismatch is shown in Figure 5 (Fig. 3 and Crowell, et al., 1982) (Cross section Z-Z'). The Upper Miocene Modelo conglomerates were derived from source areas to the northeast as shown by facies changes, diminution down-current of clast size, and sedimentary structures (Crowell, 1952b, 1954a, 1962, 1975b, 1981). The conglomerates consist of distinctive stone types, including gabbro, norite, anorthosite and gneiss, and some boulders are as large as 1 1/2 m (5 ft) in diameter. No possible source exists upcurrent today in the region just across from the San Gabriel Fault. At present this region is underlain by Miocene and older sedimentary rocks and the basement is not only deeply covered, but consists of different petrographic types. An appropriate source area is present to the southeast in the western San Gabriel Mountains, however, and satisfies the requirement of appropriate lithology. In addition, beds of Soledad Basin overlap in such a way as to indicate that the basement terrane stood high and was exposed to erosion during the late Miocene. The minimum slip needed to bring the conglomerates next to this source area requires right slip of about 35 km (22 mi) but it could be as much as 56 km (35 mi).

Immediately across the San Gabriel Fault from the Modelo conglomerate lies the Violin Breccia, also of upper Miocene age and also marine, and derived from a largely gneissic source to the southwest as shown by sharp facies changes from the fault

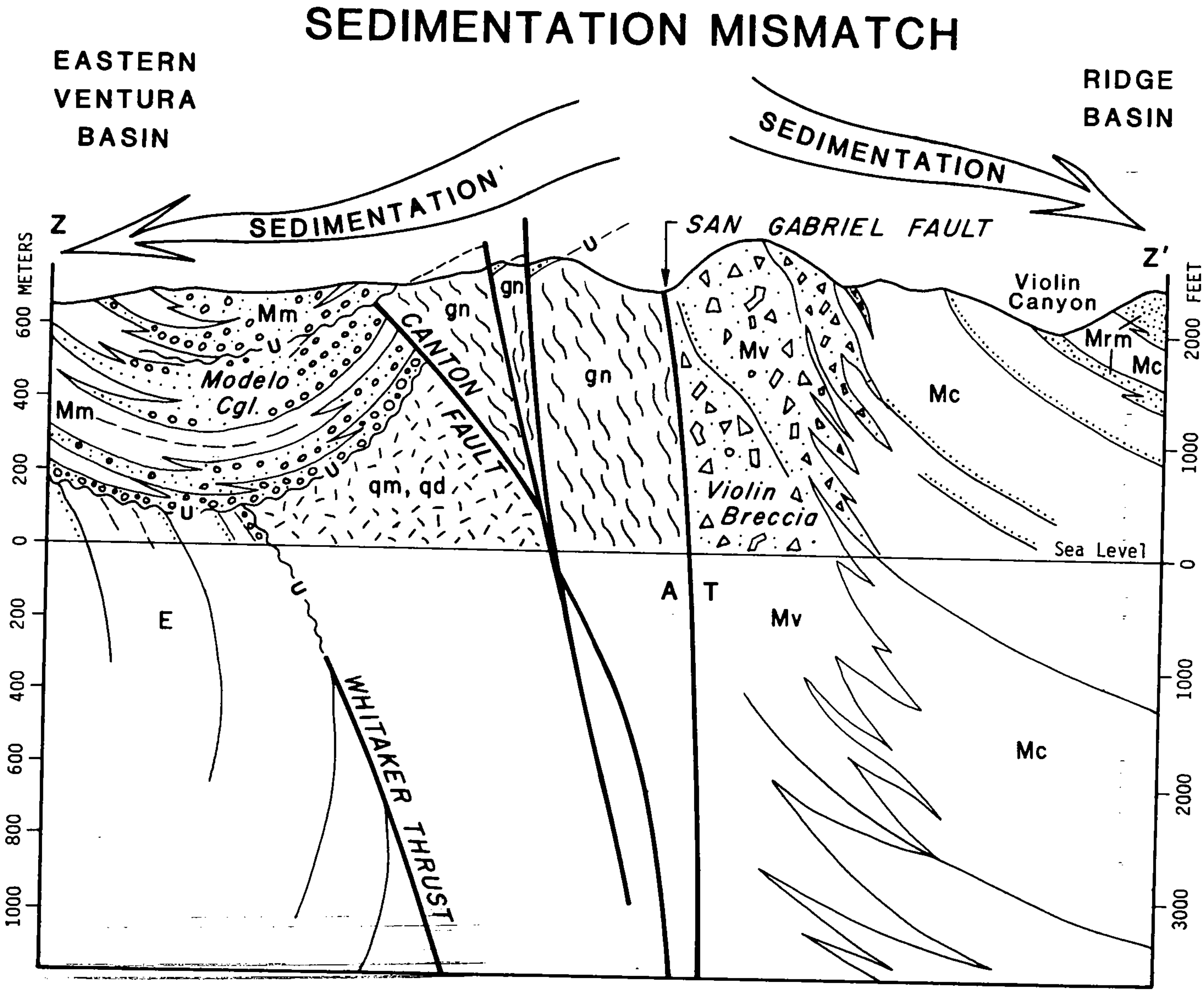

Figure 5. Cross section 2-2' (Fig. 3; Crowell, et al., 1982) showing sedimentation mismatch across San Gabriel Fault. Modelo conglomerates, derived from the northeast, now lie faulted against Violin Breccia, derived from the southwest. The source areas for both have been displaced by right slip on the fault zone. T = Displacement toward observer; A = Displacement away from observer. Symbols as in Crowell, et. al., 1982.

northeastward into Ridge Basin. Although narrow slices of gneiss similar to that contained in clasts of the sedimentary breccia are present within the San Gabriel fault zone, most of the terrane directly to the southwest is still covered by sedimentary rocks of the same age and older, including great thicknesses of Eocene, Oligocene, and older Miocene beds. Stone types within the Violin Breccia match nicely the terrane at Frazier Mountain to the northwest, suggesting that the oldest part of the Violin Breccia has been moved laterally at least 35 km (22 mi). The match between stones and source area would be equally as good with right slip of nearly 60 km (38 mi) (Crowell, this volume).

We thus have a two-pronged argument indicating right slip on the San Gabriel Fault of at least 35 km (22 mi) and as much as 60 km (38 mi). It seems unlikely that the oldest part of the Violin Breccia could have been derived from the gneissic slices within the fault zone itself because the slices are overlapped by the Modelo conglomerate and have apparently only recently been exposed to erosion during very recent uplift of the region and vigorous downcutting of present streams. Moreover, no clasts of the widespread Eocene and younger sedimentary rocks have been noted within the Violin Breccia which surely would have been the case if the region now immediately to the southwest of the breccia had been the source.

The preferred explanation advocated here is that the fault was born in late Miocene time, about 12 or 14 m. y. ago, and began displacement primarily with right slip, but with a dip-slip component. Like other similar faults with lengths measured in many tens of kilometers, these displacements resulted in scarps facing in opposing directions. The San Andreas today, for example, displays a northeastward facing scarp along the margin of the San Gabriel

Mountains, but a southwestern facing scarp at the base of the San Bernardino Mountains. The null region where the fault displays no vertical separation is in the Cajon Pass area. I picture that analogous tectonic events took place along the San Gabriel Fault in late Miocene time. The fault scarp faced to the southwest along the flanks of the San Gabriel Mountains, shedding the Modelo conglomerates into the sea. Beyond the null region beneath the Castaic lowlands, the fault scarp faced into Ridge Basin, shedding the Violin Breccia into a marine embayment.

The characteristics and distribution of the sedimentary Violin Breccia bear significantly on the displacement history of the San Gabriel Fault (Crowell, 1954a,b; 1974a,b; Crowell, this volume). The formation is unusual in that it is over 11,100 m (36,400 ft) thick stratigraphically but extends along the strike for a maximum of only 1500 m (5,000 ft). Along its total extent it very abruptly grades laterally into finer-grained strata of Ridge Basin and everywhere consists of a rubble of gneissic blocks up to 2 m (6 ft) in diameter. These are embedded in a muddy and sandy matrix, at places massive and at others stratified. The formation accumulated as talus or as debris flows and steep alluvial fans at the base of the San Gabriel fault scarp. The great thickness of the sedimentary breccia and the long time (about 7 m. y.) required for its deposition implies continuous or frequently intermittent rejuvenation of the fault scarp. These relations indicate a dip-slip component, down on the northeast into Ridge Basin and elevation of a limited source area.

The time of inception of the San Gabriel Fault is problematic. From what is now known concerning the distribution of basement rock types in both the western San Gabriel Mountains and the Frazier Mountain region, the total right slip on the fault zone is about 60 km (38 mi). This is the same amount allowed by matching the two conglomeratic units with their source areas, but exceeding the minimum required to obtain a satisfactory match between derived conglomerates and their source areas. On the one hand the faulting may have originated at about the same time as the two conglomerates, 12 or 14 m. y. ago, and be displaced the same as the total (60 km or 38 mi). On the other hand, the faulting may have commenced before the deposition of the conglomerates, by an unknown interval of time, so that they do not reveal the total displacement, but only part of it. Since it is doubtful whether it will ever be possible to match with precision basement-rock markers, or conglomerates with their source areas, I prefer a mobilistic view. Such a view would hold that 60 km (38 mi) of slip is quite reasonable within such a splintered transform belt. A regional tectonic synthesis leaning on some of the evolving precepts of plate tectonics may help to resolve this question.

The time of origin of the San Gabriel Fault also hinges on interpretation of the source of very coarse conglomerates in the Canton Canyon area, about 8 km (5 mi) west-northeast from Castaic (Crowell, 1954a, p. 51; 1962, p. 42; Bohannon, 1975, p. 77). Here, within the Sespe Formation of Oligocene Age, clasts of anorthosite, gneiss, Lowe Granodiorite, and other distinctive types present in the San Gabriel Mountains intertongue with finer deposits to the west. The coarse Sespe conglomerates, considered to range between 26 and to 30 m. y. in age are overlapped along the western wall of Canton Canyon by beds of the mid-Miocene Modelo Formation (Luisian Stage) (Woodburne, 1975; Bohannon, 1975). They are apparently derived from a concealed steep slope, presumably a scarp formed along a fault active at that time. Inasmuch as the San Gabriel Fault lies 8 km (5 mi) away, and roughly up the inferred slope from the deposits, I originally used these relations as evidence of displacement on the San Gabriel Fault during the Oligocene Epoch (Crowell, 1962, 1975b). It may be, however, that the coarse debris was shed from another fault, and perhaps one that trends with an east-northeast trend, similar to those within Soledad Basin and the western San Gabriel Mountains (Bohannon, 1976, p. 133, Fig. 73). These faults are known to have been active during deposition on the Vasquez Formation within Soledad Basin, at least 25 m. y. ago, and perhaps earlier. The hypothesis is therefore plausible that debris was shed from such a fault, and perhaps from the one trending northeastward along Big Tujunga Wash in the western San Gabriel Mountains (Jennings and Strand, 1969). The basement rocks in this region include Lowe Granodiorite, anorthosite and its relatives, and gneisses and granitic rocks similar to those in the Canton Canyon deposits. The central San Gabriel Mountains, however, have been strongly uplifted and deeply eroded in late Cenozoic time so that remnants of any basins similar to Soledad Basin have unfortunately been eroded away. Late Miocene right slip of about 55 km (34 mi) on the San Gabriel Fault would then, according to these speculations, bring the buried fault to the Canton Canyon area.

The timing of early movements on the San Gabriel fault zone also depends on interpretations of relations near the unconformities below the Modelo conglomerates and the Canton Fault and other slices in the wide and shattered basement zone. The granitic rocks of the Whitaker Peak massif, unconformably overlapped by the lower Modelo units, are sheared, implying that deformation associated with the nearby San Gabriel Fault had occurred before their deposition (Figs. 3 and 5). The Canton Fault itself clearly acquired much of its displacement during the deposition of the Modelo Formation in this area because it cuts older Modelo beds and is overlapped by younger. The gneissic rocks within the hanging wall are severely broken and shattered, so that at places (such as along the Devil Canyon Truck Trail) it is difficult to determine whether the gneisses are tectonically shattered or have been emplaced by downslope sliding accompanied by brecciation. My preferred interpretation is that the gneissic slice constituting the Canton Fault plate was brought into position during the late Miocene, primarily by being squeezed upward and laterally within the strike-slip zone of the San Gabriel Fault. It is made of rock types similar to those within the San Gabriel Mountains on the southeast and those within the Frazier Mountain region on the northwest, and thus occupies an intermediate position between the two terranes, as would be expected for a strike-slip fault-horse. Other slices of gneissic and granitic rocks within the basement terrane and within the fault zone along the margins of Ridge Basin have probably been emplaced in a similar fashion, such as those along the Whitaker Peak Truck Trail. Some have been squeezed upwards; others have been depressed; still others have travelled along within the strike-slip zone with nearly pure right slip.

The Violin Breccia and the San Gabriel Fault

The strip of Violin Breccia adjacent to the San Gabriel Fault for a length of 35 km (22 mi) reveals considerable information on the nature of the fault's history of displacement, as well as on the origin of Ridge Basin itself. As described in Crowell (this volume), the unit is over 11,000 m (36,000 ft) thick stratigraphically, and it changes facies within a kilometer or so from coarse gneissic rubble into finer beds within the central trough of Ridge Basin. The principal mappable strand of the fault zone marks the boundary between basement terrane on the southwest, and Violin Breccia on the northeast and is therefore the easiest strand to demarcate in the field. It may well be, however, that fault strands within the basement, characterized by gouge and fault breccia, have greater but unproven displacements.

The nature and distribution of the Violin Breccia requires the continuous or closely intermittent rejuvenation of the San Gabriel fault scarp throughout the time of its deposition. This includes a span of time of about seven million years, from about 12 m. y. ago until 5 m. y. ago, when the fault zone was overlapped by beds of the Hungry Valley Formation. During this 7 m. y. interval, although the principal displacement was by right slip, Ridge Basin was always depressed with respect to a limited source area in the environs of Frazier Mountain. The latter region, during the same long interval, was continuously or intermittently elevated to provide a limited source area shedding gneissic debris across the fault into the basin. Since only about 10 or 15 per cent of the sedimentary bulk of Ridge Basin was washed into the basin in this manner, streams courses leading to the fault scarp are pictured as steep and short. Between these steep valleys, much of the debris within the Violin Breccia was deposited in talus and small alluvial cones. Most of the material within the basin, as brought out elsewhere in this volume, entered Ridge Basin from the north and northeast. Nonetheless, inferences gleaned from the manner of deposition of the Violin Breccia tell much concerning the manner of deposition within the basin as a whole.

A conveyor-belt mechanism is envisioned to account for the distribution of the Violin Breccia (Fig. 6) as well as for other units in the basin. Immediately across the San Gabriel fault scarp from the rugged source area lay the Ridge Basin depocenter. Each time there was an earthquake upon the fault, the depocenter was both depressed and moved right-laterally to the southeast with respect to the source area. Displacement might also have occurred by slow creep. At the same time, the source area was elevated and erosionally rejuvenated. This mechanism not only accounts for the total of more than 11,000 m (36,000 ft) of Violin Breccia but also for the northwestward overlap of other sedimentary units within Ridge Basin, as shown in Crowell et al., (1982; isometric cross sections) and in the Frontispiece (this volume). If this kinematic scheme had paused significantly during the 7 m. y. of its operation, the fault scarp would have been quickly reduced by erosion and streams would have worked their way headward to the southwest. They would have

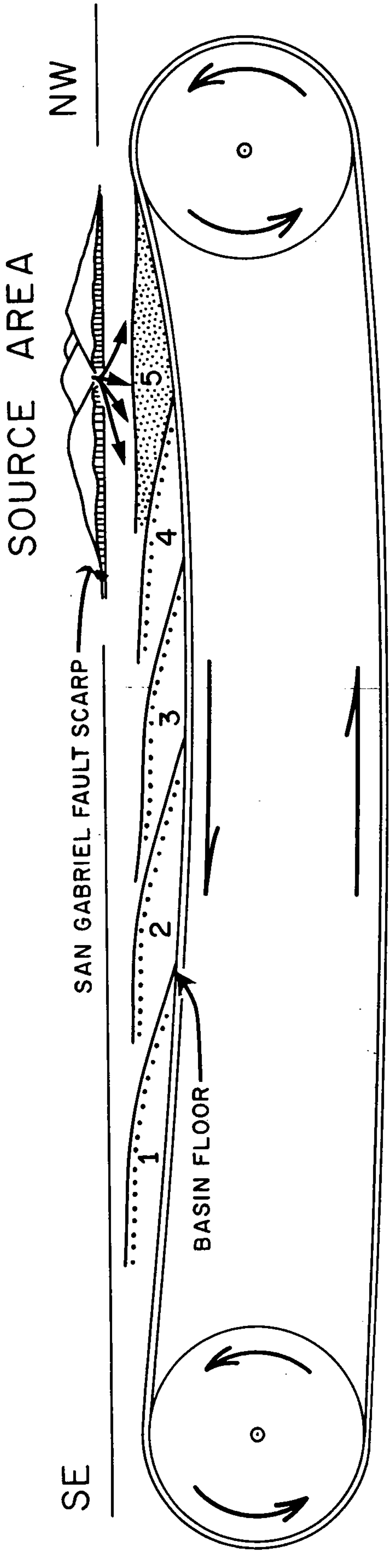

Figure 6. Diagram illustrating conveyor-belt mechanism for deposition of sedimentary units in Ridge Basin, including the Violin Breccia.

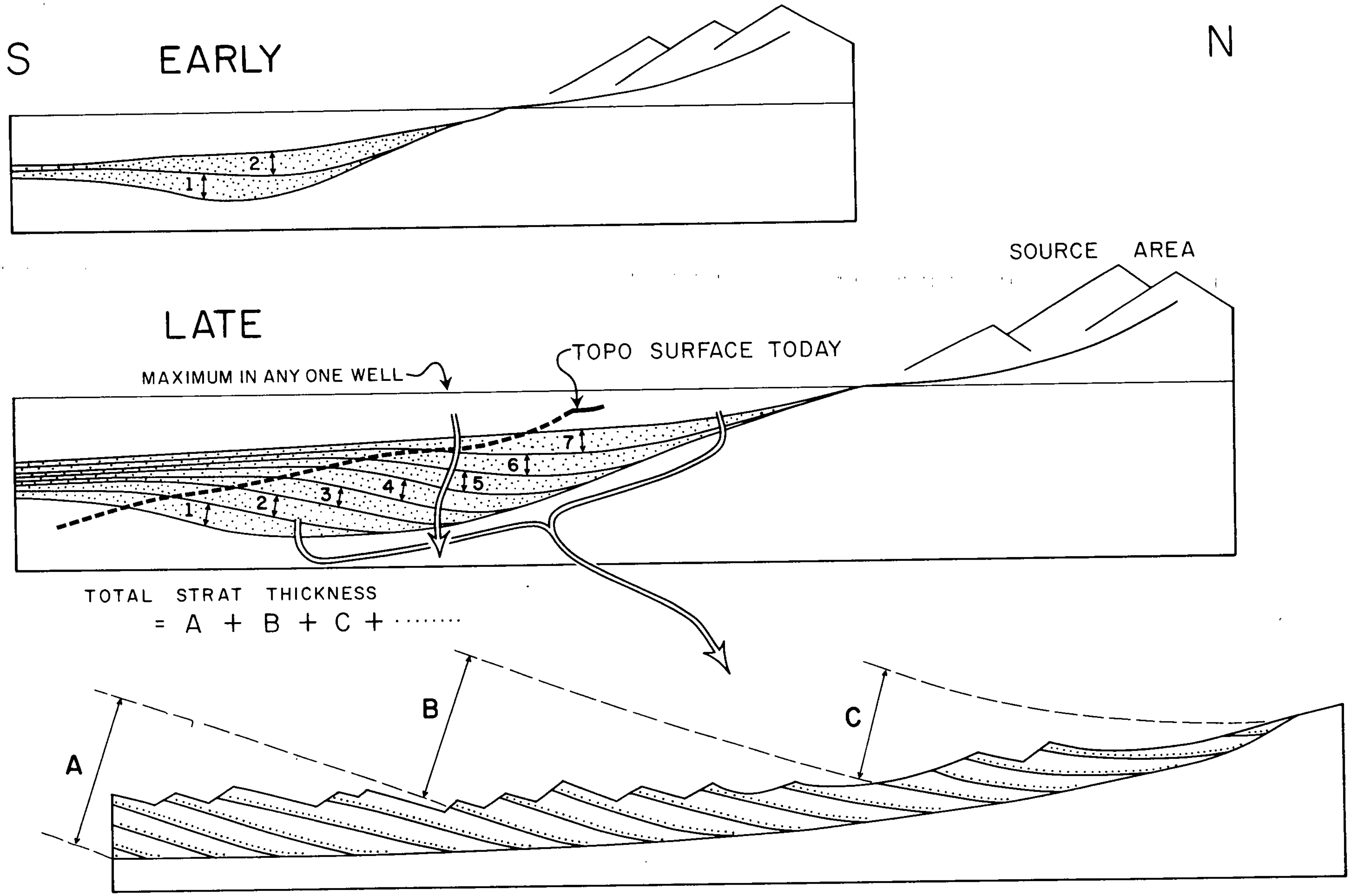

Figure 7. Diagram showing arrangement of stratal units within Ridge Basin, parallel to the depositional trough (see Frontispiece of volume). The depocenter is depicted as migrating toward the principal source area, interpreted to lie to the north of the basin. The upper and center cross sections show the scheme at early and late stages, and the lower cross section shows diagrammatically the way the region is exposed today after uplift and erosion. The total stratigraphic thickness is obtained by adding thicknesses measured along the topographic surface today, and is reproducible by any stratigrapher.

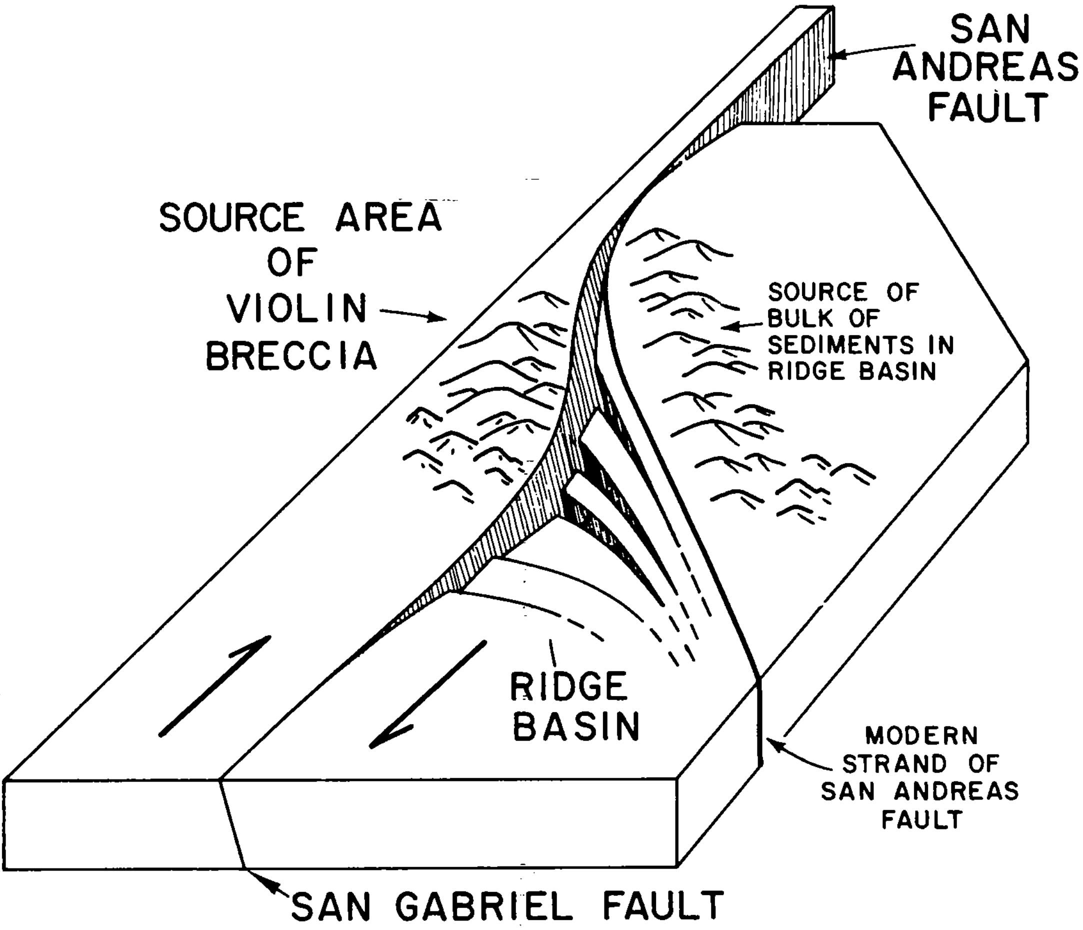

Figure 8. Block diagram illustrating origin of Ridge Basin at a sigmoidal bend in the San Andreas Fault. Modified slightly from Crowell (1974a, Fig. 5).

brought finer-grained sediments to the basin across the San Gabriel Fault, for which there is no evidence until the fault died in mid-Hungry Valley time. Continuous operation of the conveyor-belt mechanism for the 7 m. y. interval is implied.

In Figure 6 a conveyor belt is pictured as carrying Violin Breccia and other Ridge Basin strata laterally and parallel to the San Gabriel Fault. The depocenter is diagrammed as a loading of the flexible conveyor belt. As the older units are transported they acquire a gentle northwestward dip and overlap upon the basin floor. They also are shown thinning laterally and upward in the direction inferred for the outlet of Ridge Basin at times when external drainage prevailed. Because they have been carried laterally in this right-slip system, they overlap like shingles on a roof or scales on a fish. The first-formed and oldest wedge of Violin Breccia has been carried farthest, to the southeast, with successively younger wedges overlapping to the northwest. The last-formed wedge has not been displaced significantly, and remains opposite the source area.

Although the diagram is drawn as pertaining to the Violin Breccia, the same concept applies to the bulk of strata laid down in Ridge Basin (Fig. 7). Most of the debris came from the north and northeast. Note that a well drilled in the basin would penetrate only about one third of the total stratigraphic thickness as measured along the surface before it would reach the basin floor.

These relations show that the San Gabriel Fault is a type of growth fault in that faulting and sedimentation have gone on concurrently. Inasmuch as older units of the Violin Breccia have been displaced more than younger, the San Gabriel Fault has a near infinite number of displacement values. We can speak of the total displacement of the fault or of the displacement of individual beds or time horizons, but not of a single value for the fault as a whole. Moreover, the scheme suggests that the orientation of the slip vector on the fault has changed from place to place as a function of time and position. This inference follows if the floor of Ridge Basin is envisioned as flexible and the source area as bending upward irregularly rather than moving as a rigid block. Older parts of the Violin Breccia and other beds in the basin at the southeast show more shearing and small-scale faulting than younger exposures at the northwest, presumably for this reason and because they have existed for all of the 7 m. y. during repeated displacements.

The History of the San Gabriel Fault Zone

In summary, the San Gabriel fault zone has had a long history involving different types of displacements. It probably originated as a major transform fault at the plate boundary between the Pacific and North American plates about 12 m. y. ago, but perhaps as early as 14 m. y. ago. In drawing this conclusion I am assuming that the Sespe conglomerates in the Canton Canyon area of Oligocene age were deposited near another and much older fault that was not part of the San Gabriel system, but this choice is not proven. As the fault developed in late Miocene time, the walls moved primarily by right slip, but with a low angle of oblique slip. This resulted in the raising of a low straight scarp along the southwestern margin of Ridge Basin, where the terrane stood structurally higher on the southwest. Beneath the Castaic lowlands, displacement was by nearly pure strike slip so that beds were carried along by regional trace slip (Crowell, 1962, p. 13). As a consequence, in this region very little vertical separation developed so that time-stratigraphic units, such as those commonly dealt with in oil field subsurface studies, are apparently easily correlated across the fault zone. Lithologic and facies units are not. Farther to the southeast, the growing fault scarp faced to the southwest and the San Gabriel Mountain block stood structurally higher.

The fault at this time was probably the main strand of the San Andreas fault system, and connected with the San Andreas on the northwest in central California, north of the present "Big Bend" region. Figure 8 is a sketch showing Ridge Basin without sediments within it. The trace of the fault was not completely straight through this connecting region, however, but made a gentle arc swinging eastward around a restraining bend bounding the Frazier Mountain region (Crowell, 1974b, p. 193). Movement of the two plates around this restraining bend squeezed the terrane of Frazier Mountain and elevated it to make a source area for the Violin Breccia. At the same time, the terrane outside of the gentle bend on the northeast was stretched and therefore sagged to make the floor of Ridge Basin. As displacement on the fault continued, wedges of sediment were strewn out and displaced laterally according to the conveyor belt mechanism described above. This method of fault displacement continued smoothly into the beginning of the Pliocene Epoch and came to an end in mid-Hungry Valley time. Along the southeastern reaches of the San Gabriel Fault, although the tectonic history has not yet been satisfactorily worked out, the fault probably extended into the San Bernardino Mountains-San Gorgonio Pass region. It is conceived of as extending into the proto-Gulf region of southeastern California, but not into the head of the present Gulf of California because the gulf did not open until later.

Beginning about 5 m. y. ago, this right-slip displacement on the San Gabriel Fault ended and for a time the fault was dormant. Within the late Pliocene and continuing into the Pleistocene, however, the fault zone was reactivated at places. The Bear Mountain Fault on the north was born and terminated against the San Gabriel Fault resulting in the downdropping and tilting of the western wall of the San Gabriel Fault to the south. As this block was deformed, the main strand of the San Gabriel Fault was deformed and bent and the belt of Violin Breccia was repeated. This repetition took place on a new strand dipping to the southwest, and displacing by oblique slip. Within adjoining Ridge Basin, beds were folded and oversteepened near the fault, presumably at the same time as the Ridge Basin Syncline was formed approximately along the axis of the earlier depositional trough. Along much of the margin of Ridge Basin, the San Gabriel Fault dips to the northeast. Beds within the basin appear to have ridden out and upward with reverse slip primarily.

Across the Castaic lowlands, and at a few places elsewhere along the fault zone, these late reactivations have broken Pleistocene stratigraphic units (Crowell, 1952, p. 2034; Weber, 1979). They are here explained as renewed movements along the major fault zone during later regional deformation because the fault represents a major mechanical discontinuity in the strength properties of the crust. It is no longer a major transform fault. This interpretation is supported by data acquired during and after the San Fernando Earthquake of February 9, 1971 (Grantz et al, 1971; Oakeshott et al, 1975). The zone of rupture during this earthquake dipped beneath the western San Gabriel Mountains and apparently ignored completely the inactive San Gabriel Fault at depth. The epicenter of the earthquake lies to the north of the surface trace of the San Gabriel Fault and ground rupture along the San Fernando Fault lies to the south of it. Tectonic deformation in this general region is pictured as resulting from north-south crustal compression related to the evolution and uplift of the Transverse Ranges. It is apparently not related to the local presence of a major strand of the San Andreas transform system.

CLEARWATER FAULT

Two major fault systems enter Ridge Basin from the east, the Clearwater and the Liebre (Fig. 3; Crowell et al., 1982). From the place where the western end of the surface trace of the Clearwater Fault is overlapped by beds of the Ridge Route Formation, the fault can be traced to its junction with the San Francisquito Fault beneath Bouquet Reservoir and on to the San Andreas Fault, a total distance of about 43 km (27 mi). East of Ridge Basin studies of rocks in either wall suggest a complicated history because the sense of vertical separations differs from place to place: low on the north at the west and high on the north to the east. The fault is probably an oblique- and right-slip fault with at least 1830 m (6000 ft) of right slip and 1200 m (4000 ft) of vertical slip, down on the south (Stanley, 1966). Inasmuch as older and basement rocks along the fault zone are quite different, however, its total displacement is probably much more, but how much more is not yet known. Its straight trace across the rugged mountains, and the broad zone of splintered and shattered rocks suggests that the zone is nearly vertical, although major splays within the zone dip steeply to the north (Fig. 3). The belt of sheared basement rocks along it is clearly truncated locally by Ridge Basin beds, so the origin of the fault is older than their deposition.

On the west, in the region where the Clearwater fault zone is overlapped by Ridge Basin beds, intermittent or renewed displacements on it are indicated. Here a triangular sector of Ridge Route Formation is structurally depressed on the northeast, and truncated by the Clearwater Fault on the south. This same fault, however, is unconformably overlapped by somewhat younger Ridge Basin beds which lap directly upon older rocks in the elevated block to the south of the fault and upon the triangular sector on the

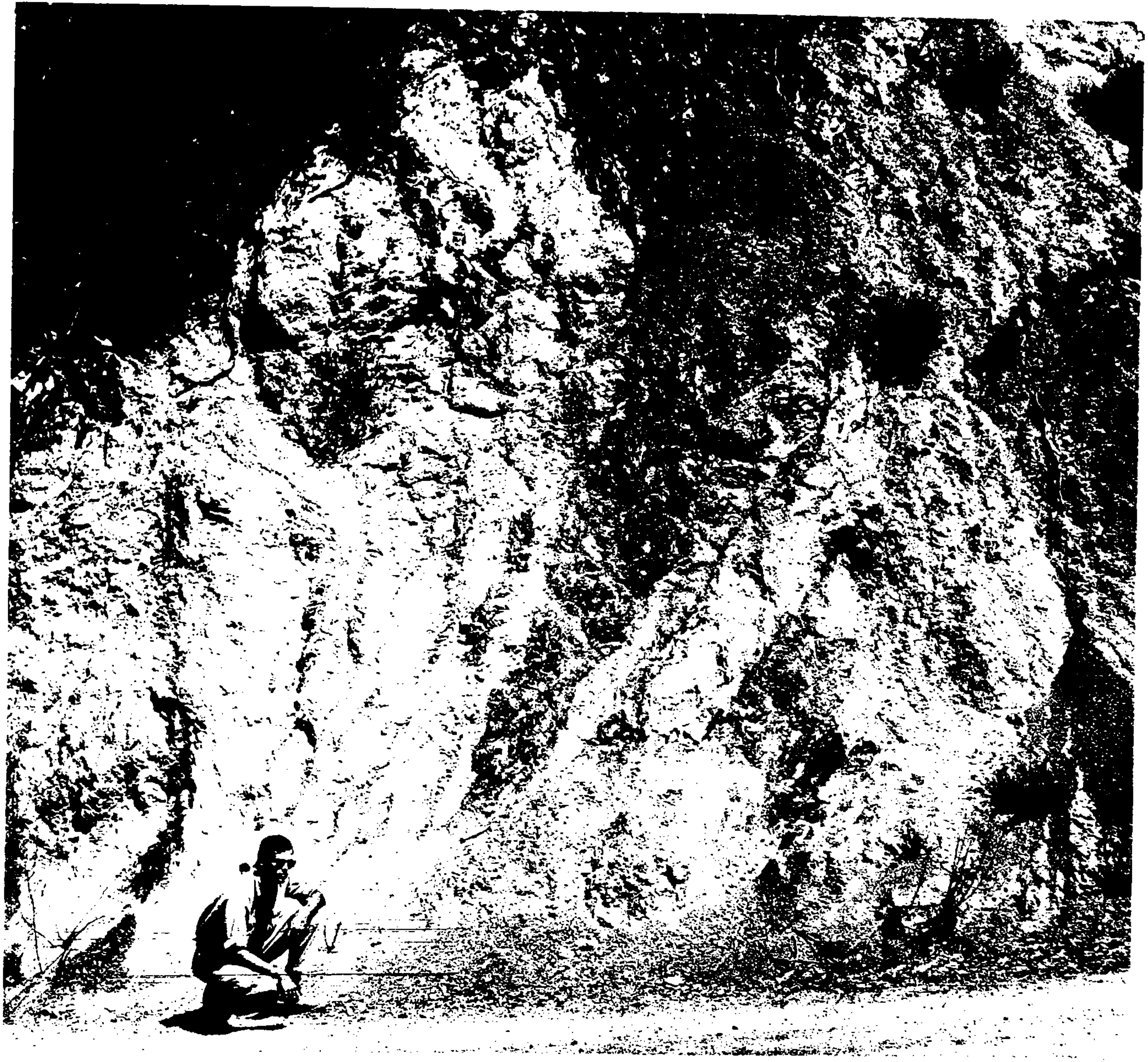

Figure 9. Photograph of Clearwater Fault, Ruby Canyon, east of Ridge Basin. Fractured basement granitic rocks on left (north). Sheared rocks, including phacoids of San Francisquito Formation, in center to right (south). Roger Hope in photograph.

north. Here there is a right separation of at least 4 km (2.5 mi), but the amount of slip and the orientation of the slip vector is not yet known. According to Ensley (1980), based on magnetostratigraphic dating of the sediments, the Clearwater Fault originated here before 8.1 m. y. and died 7.8 m. y. ago. Farther west along the trend of the Clearwater into Ridge Basin, beyond the point of overlap, lies a belt of en echelon folds and faults. These are interpreted as surface expressions of renewed movement on the Clearwater Fault at depth, after the bulk of the Ridge Basin sediments was deposited. Right simple shear between the basement blocks is suggested by the oblique orientation of the overlying folds, and the left-stepping arrangement by some, but not all, of the faults along the trend. The fault is presumably truncated by the San Gabriel fault zone at depth, because the later is clearly in part younger. Its offset counterpart, if it exists, has not been recognized west of the San Gabriel Fault.

In summary, the Clearwater Fault has had several episodes of displacement upon it: 1) some previous to the deposition of Ridge Basin beds near the point of overlap; 2) a time of displacement not long before the time of overlap to account for the depressed segment of Ridge Route Formation, and, 3) local renewed movement along its western course thereafter that did not result in cutting completely through the thick mass of sediments in the central part of Ridge Basin. Instead, these later displacements are pictured as local folds and breaks in the sediments as the result of renewed jostling between the fault walls. Because several thousands of meters of Ridge Basin beds extend southwestward from the Clearwater trace much of the displacement probably occurred during sedimentation within Ridge Basin. Displacements on the fault during the formation of the basin are here interpreted as accounting for most of the shearing in the basement along its trace before the local deposition of Ridge Route beds overlapped it, but some may be much older.

LIEBRE FAULT ZONE

Bordering Ridge Basin on the northwest at least four strands of the Liebre fault zone are overlapped by successively younger beds of the basin sequence (Fig. 3; Crowell, et al., 1982). Faults within the zone were therefore active during deposition of Ridge Basin sediments and coarse conglomerates show that the fault zone played a roll in forming the basin margin. From Ridge Basin the belt of high-angle faults can be traced on eastward for at least 20 km (12 mi) into rugged basement terrane as yet unmapped at large scale.

Along the south flank of Liebre Mountain, two main strands of the fault separate a long slice of gneissic and granitic rocks between granitic rocks within Liebre Mountain on the north, and Ridge Basin beds on the south (Faggioli, 1952). The contact between gneiss and granite appears similar to one north of the fault, about 8 km (5 mi) to the east. If this contact is the same a right separation of this amount is indicated, but not enough petrographic and geochemical work has been done to be sure of the correlation. And it is also not known whether the separation is due entirely to right slip, although the steep dip of the contact as observed in the field suggests that the lateral separation is tantamount to slip.

Overlap relations of strands of the Liebre fault zone are quite well demarcated along the south flank of Liebre Mountain. Just southwest of Kelly Ranch on the east (Crowell, et al., 1982), the southwesternmost strand brings gneiss above a wedge of Ridge Route conglomerate that lies depositionally upon gneiss in the footwall. A few meters higher stratigraphically the base of the Ridge Route Formation can be traced across the fault, which it overlaps, and around a west-plunging anticline. Near Kelley Ranch this contact is truncated by the next strand of the Liebre zone, which, when traced westward, bifurcates into the two main branches of the fault zone. The southern zone is overlapped in turn near the Old Ridge Route, and deep canyons in this region disclose overlap relations clearly. Overlapping beds here wrap around the next plunging anticline, and are truncated by the northern branch of the Liebre Fault. A third repetition of these relations is mappable between the Old Ridge Route and Interstate 5, in the upper reaches of Apple Canyon, and a fourth where the base of the Hungry Valley Formation overlaps the northeasternmost strand of the Liebre fault system off the plunging western end of Liebre Mountain. These four episodes of overlap document the migration northeastern of the principal strand of movement on the fault system during the time of Ridge Basin sedimentation.

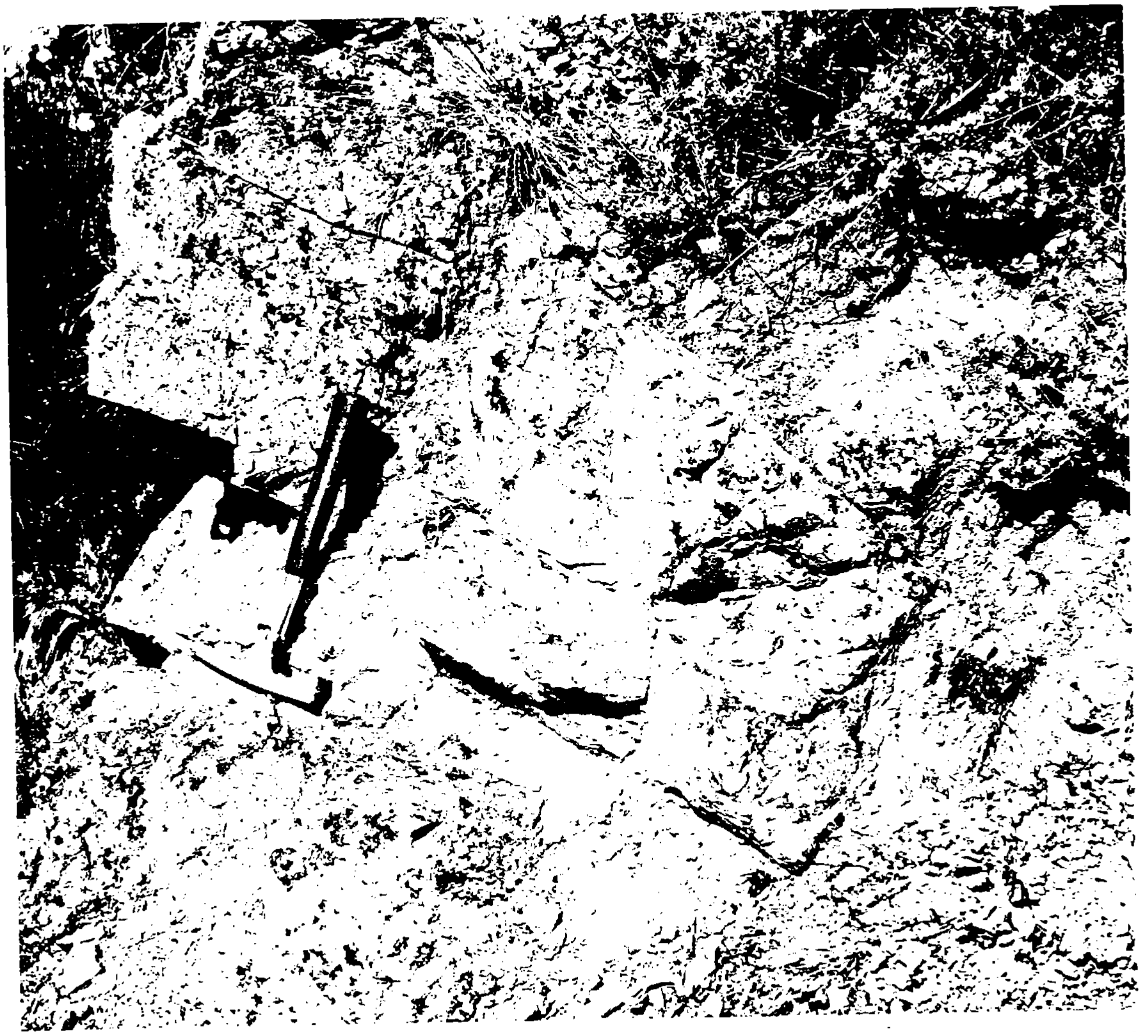

Figure 10. Anorthosite and diorite, Bear Gulch (a tributary to Piru Creek), near the northwestern surface terminus of the San Gabriel Fault.

Structure at depth beneath the plunging western end of Liebre Mountain is puzzling (Fig. 3; Crowell, et al., 1982, Cross section A-A'). The Liebre Mountain fault zone dips steeply to the north, and but 1.6 km (1 mi) to the north, the Sandberg Thrust dips to the south. These two faults therefore converge toward each other with depth, suggesting that the basement-rock mass of the mountain possesses a keel, perhaps detached and separated from underlying sedimentary rocks. Perhaps this part of the mountain has ridden up laterally and obliquely during transcurrent displacements in this complex region where the Liebre and San Andreas systems entangle.

FRAZIER MOUNTAIN THRUST SYSTEM

The northwestern corner of Ridge Basin is bordered by the Frazier Mountain thrust system and related faults. One of the most conspicuous faults of this system, the Dry Creek Thrust, is well exposed about 500 m (1600 ft) up Dry Creek where the creek is crossed by the Forest Service access road to the Gold Hill and Alamo Mountain area, southwest of Gorman and Hungry Valley. Here gneiss cut by many veins of granitic rock is brought up across overturned Hungry Valley beds on a wide gouge zone dipping about 50° to the north. The Dry Creek Thrust merges with another element of the system on the west (Fig. 3; Crowell, et al., 1982). The principal branch of the system trends along the base of the steep flank of Frazier Mountain and also has emplaced Precambrian gneiss upon Pliocene Hungry Valley conglomerate. The thrust joining this fault and the Dry Creek Thrust has been severely folded into antiforms and synforms that have affected not only the gneissic hanging wall but the thin veneer of Hungry Valley beds below, and the basement terrane in turn below these beds. Although total movement on the system apparently has taken place during Pliocene and Pleistocene time, large fanglomerate wedges and terrace deposits of late Pleistocene age have not been cut by it, so it is concluded that the fault system is not now active.

Basement rocks constituting Frazier Mountain are the same or very similar to those beneath the thrust complex along Piru Creek and in Alamo Mountain, and very different from those across the modern San Andreas Fault to the north. It is therefore concluded that the thrusts originated by north-south shortening that has telescoped the rocks to the south of the San Andreas (Crowell, 1950, p. 1645). On the east both the Dry Creek and main Frazier Mountain thrusts terminate with straight faults on the map. These are interpreted as uplifted and rotated segments of the San Gabriel Fault as shown on Figure 3. Under the premise that this hypothesis is correct, increase of thrust slip eastward and rotation of the thrust slabs are required but the total displacement on the system need not exceed about 6 km (3 mi) (Crowell, 1950).

The northern strand of the Frazier Mountain Thrust can be traced westward and around the mountain to intersect the Big Pine Fault in the Lockwood Valley region (Carman, 1964; Crowell, 1964). The mountain mass is therefore completely surrounded by faults, and those on the east, south, and west dip steeply beneath the mountain and bring gneiss out over the top of Hungry Valley and Quatal sedimentary rocks. This geometry suggests a "mushroom" shape for the mountain and emplacement involving uplift followed by sagging around its margins. The gneissic mass is envisaged as yieldable and not strong enough to support itself. The timing of the emplacement of the thrust system is constrained by the timing of deposition of Hungry Valley beds lying stratigraphically several hundred meters above the place where they overlap the San Gabriel Fault, and the age of truncations by both the Big Pine Fault and the San Andreas. The thrust is older than later movements on the Big Pine, which in turn are older than those on the San Andreas. The mountain began to rise, however, during the time of the later deposition of Hungry Valley beds as shown by their lithology in the northwestern part of Ridge Basin (Crowell, this volume). The tectonic scheme in this region apparently fits the concept that the thrust originated and grew as the Big Bend in the San Andreas developed during the Pliocene and Pleistocene epochs.

SAN ANDREAS FAULT

Ridge Basin is sharply truncated on the north by the active San Andreas Fault. Older rocks along the fault zone here are severely broken and where the fault is exposed at the top of Tejon Pass, a meter of black tar-like gouge separates Upper Pleistocene terrace deposits from comminuted and sheared granitic rocks. Slices within the fault zone are out-of-place, and some cannot be easily matched with sources in either wall. Lenses or slices occur in the sinuous fault zone; one of these is conspicuous on the hillside above the town of Gorman (Crowell, 1952b). A mass of broken rock, largely granodiorite (gd on Fig. 3), now lies as a thrust plate south of Gorman and is interpreted as material squeezed from the fault zone. Quaternary movements along the fault zone have resulted in offset streams, shutter ridges, sag ponds, and other fault-landform features (Sieh, 1978; Crippen, 1979; Duebendorfer, 1979). The last displacements occurred during the 1857 Fort Tejon Earthquake, and scarps and other features resulting from this strong event are still fresh (Wood, 1955; Sieh, 1978).

The mismatch in basement and older rock types and histories across the San Andreas Fault in this region suggests great lateral displacement, and the history of sedimentation within Ridge Basin gives some information concerning its timing. Basement rocks formed in the deep crust southwest of both the San Andreas and San Gabriel faults include several types of gneisses and augen gneisses, and several varieties of granitics and other intrusives. These include small bodies of anorthosite, gabbro, and related rocks where Piru Creek first flows to the San Gabriel Fault (Fig. 10). These rocks have been identified northeast of the San Andreas Fault only in the Orocopia Mountains, north of the Salton Sea, about 530 km (200 mi) to the southeast. They are, however, present within the western San Gabriel Mountains and Soledad Basin, southwest of the San Andreas. Their distribution is part of the argument for great displacement on the San Andreas fault system in southern California, suggesting 60 km (37 mi) on the San Gabriel Fault, with an additional 270 km (165 mi) on the San Andreas (Crowell, 1960, 1962, 1975a, 1981; Ehlig and Ehlert, 1975). Figure 11 is a diagram showing the succession and arrangement of rock types in the three critical regions: the Tejon Region west of Ridge Basin, and the Soledad and Orocopia regions to the southeast.

If these large displacements are eventually proven approximately correct, there was a belt of terrane made up of similar rocks in similar sequence across the site of the northern part of Ridge Basin, but before its origin. The belt has been displaced

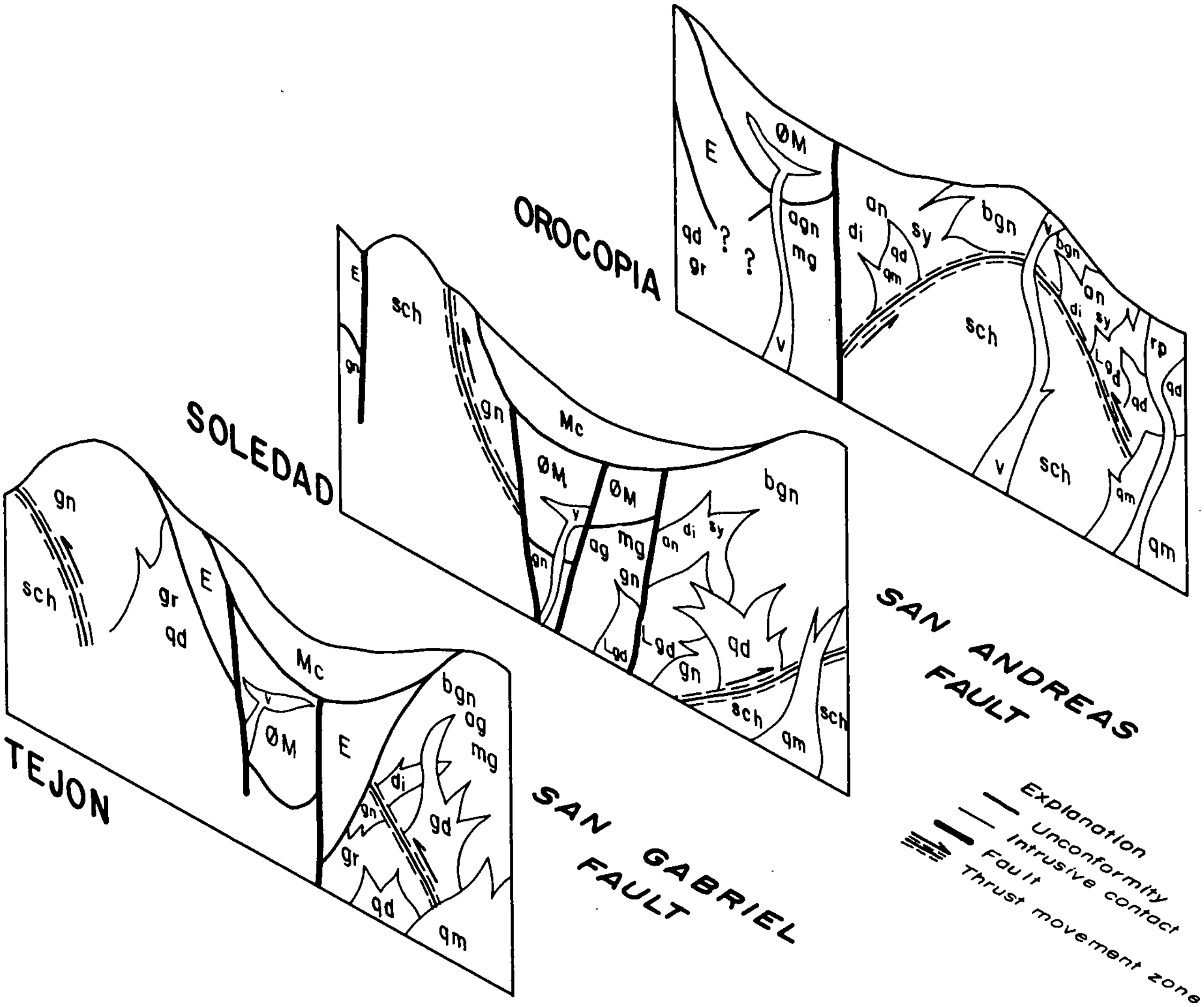

Figure 11. Diagrammatic cross section showing rock types and sequences in three regions displaced by the San Andreas-San Gabriel fault system. Not to scale. Pliocene and younger formations omitted. Symbols: Precambrian:agn=augen gneiss; bqn=blue-quartz-bearing gneiss; mg=migmatite; di=diorite and gabbro; an=anorthosite; sy=syenite; Pre-Tertiary, but mostly Mesozoic:gn=gneiss; Lgd=Lowe granodiorite; gr=granite; qd=quartz diorite; qm=quartz monzonite; sch=Pelona and Orocopia schists; Tertiary: E= Paleocene San Francisquito Formation (the lower part is Cretaceous in age) and other Eocene formations; ØM=Oligocene and Lower Miocene nonmarine conglomerate, sandstone, and shale, with associated volcanic rocks; Mc=Middle and Upper Miocene sedimentary formations, largely continental; rp=rapikivi-textured quartz latite porphyry; v=other volcanic rocks. Reproduced from Crowell (1975a, Fig. 2).

along the San Gabriel fault zone, but these displacements ended within the time of deposition of the Hungry Valley Formation, or about 5 m. y. ago at the beginning of the Pliocene Epoch. Inferences from facies of the formation, and tentative conclusions concerning its provenance, discussed in Crowell (this volume), suggest that the San Andreas Fault began its truncation of Hungry Valley beds at about the same time. As the San Gabriel fault weakened and died the San Andreas took over the transform displacements. Two questions arise: 1) How much displacement has taken place on the fault during this 5 m. y. interval? and 2) Did the San Andreas Fault exist at its present site before the truncation of the Hungry Valley Formation? Complexities off the plunging western end of Liebre Mountain suggest that the San Andreas was active just north of the mountain during and just before Hungry Valley time, but this is not certain. The key that is missing is the identification of offset patches of the Hungry Valley Formation or older units of the Ridge Basin sequence, or their source areas, on the northeastern wall of the San Andreas. Such discoveries would disclose the amount of slip during the interval (5 or 6 m. y.). As yet, however, these patches or a source area have not been recognized. Research on the constituents of the younger Ridge Basin beds and on basement terranes of likely source regions is sorely needed (see Crowell, this volume).

If we speculate that all of the displacement needed to match the basement terranes took place beginning 5 m. y. ago, about 270 (165 m) of right slip is needed. This gives a rate of displacement of 5.4 cm/yr (2.1 in/yr). Although this is an acceptable rate for displacement between the Pacific and North American plates it has not yet been proven by the matching of dated units. Information now in hand makes it unlikely that the total displacement is this young and this much. Such displacement would require the Hungry Valley Formation and immediately underlying beds to be offset from a region about 270 km (165 mi) to the southeast, or from the terrane now underlying the eastern San Bernardino Mountains north of Indio. There is little in the basement rocks here to suggest that this was an appropriate source area. A more attractive hypothesis, based on what we now know, is that the formation has not been offset more than about 150 km (90 mi) from a source area north of the Cajon Pass - western San Bernardino Mountains. I tentatively conclude that the San Andreas Fault existed north of Ridge Basin during the later time that the basin was being filled in; that is 5 or 6 m. y. ago. According to this view, the fault formed its northern margin during the later part of the basin's existence.

OTHER FAULTS AND FOLDS

The Sandberg Thrust, here named for exposures along the Old Ridge Route about a kilometer north of the crossroads at Sandberg, has emplaced broken granitic rocks above white Hungry Valley conglomeratic sandstone. The fault, which roots to the south, has been dismembered by branches of the San Andreas fault zone so that isolated klippen and fensters occur along the northern slopes of Bald Mountain, the western end of Liebre Mountain (Crowell, 1964, p. 37 and Map 9). The basement rocks in the hanging wall are severely broken, sheared, and at places constitute a tectonic breccia. The Sanberg Fault is interpreted as a thrust fault rather than as a composite series of faulted landslides primarily because it can be traced laterally for over 4 km (2.6 mi) and there is no apparent uphill source large enough to account for landslides of this size. The structure is viewed as the lip of a high-angle fault, dipping steeply to the south at depth, and one forming the northern boundary of the keeled Liebre Mountain block. It is considered to be a low-angle oblique-slip fault because of its close association with the San Andreas and Liebre fault zones, but no direct evidence bearing on its slip direction and orientation has yet been recognized. Because the surface lip of the fault has moved downslope by landsliding, many exposures are best interpreted as a combination of thrust and slide: informally I therefore refer to the feature as a "thride" or "slust".

The Ridge Basin Syncline, plunging to the northwest, is the most conspicuous geologic structure through the center of the basin, and Interstate 5 follows along very close to the trace of its axial surface. The structure is therefore viewed by numerous geologists as they speed along the highway. It accounts in part for the northwestern dip of Ridge Basin beds, and the syncline closely follows the thalweg of the depositional trough which received the thick section of shingling strata during the Miocene. Folding has primarily occurred during the Pleistocene and is associated with compression and uplift of the central Transverse Ranges. In the northwestern part of the basin, many other folds and some faults are also interpreted as formed late in the tectonic history of the region. Some of these follow along the buried traces of the Clearwater and Liebre fault zones. Near the southeastern corner of the Frazier Mountain massif, folds trend nearly at right angles to each other: on the west, shortening is primarily directed north-south, whereas within the northwestern part of Ridge Basin, where the stratigraphic section is much thicker, shortening within Hungry Valley beds has primarily been in a northeasterly - southwesterly direction.

ORIGIN OF RIDGE BASIN

Ridge Basin originated at the splintered boundary between the Pacific and North American lithospheric plates some 14 or 12 m. y. ago in the Miocene Epoch. During most of its development, the San Gabriel Fault was the principal strand of the San Andreas Transform and it bounded the basin on the southwest (Fig. 8). The fault was curved slightly in the region where later the "Big Bend" of the modern San Andreas Fault developed. This bend, interpreted as a manifestation on and beneath the continent of the Murray Fracture System coming in from Pacific Ocean regions, served as a constraining bend to easy right-slip along the fault (Crowell, 1974b). Terrane within the bend on the west was squeezed and uplift to make an upland source area for the Violin Breccia. Terrane on the eastern side of the fault zone was stretched so that it sagged to make Ridge Basin. Because such bends are probably sigmoidal in plan view, an uplifted area is pictured on the northeastern side of the fault zone as well, but farther to the north. This region provided the bulk of the material washed into Ridge Basin, lying to the south. At this time the San Andreas Fault is interpreted to have had a straight course at either end of the sigmoidal constraining bend; the southeastern straight part was the San Gabriel Fault.

As the depressed floor of Ridge Basin moved along the side of the sharpest part of the constraining bend, first the Clearwater and then the Liebre fault systems became active. They aided in adjustments between the moving basin floor and the northeastern boundary region of the basin where sediment-source areas lay. Uplands to the north and east, contained within the curvature of the northern bend of the sigmoidal fault zone, provided most of the debris carried into the basin. The moving-system concept suggests that this terrane, as it was carried relatively toward the southeast into developing Ridge Basin, was broken into segments by displacements on the Clearwater and Liebre faults, successively. The segments were depressed as they were carried out of the constraining bend. Fairly late in the development of the basin, the San Andreas Fault as now named, came into being, probably before the Liebre fault system had died completely. The birth of the modern strand of the San Andreas is tentatively considered to have occurred 5 or 6 m. y. ago. As the fault developed, and the associated restraining bend sharpened, the San Gabriel Fault was abandoned. No longer did the huge lithospheric plates slide by each other on this fault; instead, the San Andreas as we know it today took over as the principal strand of movement.

This outline of the tectonic history of Ridge Basin is discernible today because of late uplift of the central Transverse Ranges (Fig. 12). The uplift began in late Pleistocene time and is still actively going on today as shown by geomorphic features and the very existence of the mountains, including perhaps geodetic measurements (Castle, et al., 1976). The growth and reenforcement of the Big Bend in the mobile tectonic scheme are closely associated. The late uplift of the region, accompanied by deep erosional disection, has revealed the overlap relationships between the strata infilling Ridge Basin and the surrounding faults and folds. Similar events where moving depressions have been filled in by sediments are deemed to have taken place elsewhere in California and along other transform plate margins on Earth, but the evidence in most remains burried or has been eroded away.

ACKNOWLEDGEMENTS

In addition to all of those geologists mentioned in the preface of this volume who have aided in understanding Ridge Basin, I am indebted to Martin H. Link for critically reading this paper.

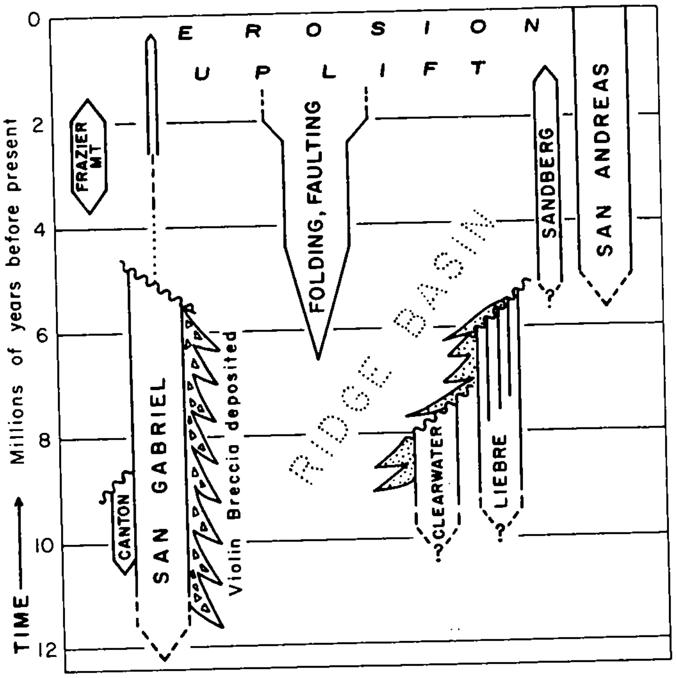

Figure 12. Time graph showing history of fault activity, deposition of sediments within and marginal to the basin, folding, uplift, and erosion, for Ridge Basin, southern California.

San Gabriel Fault, Piru Creek, below Frenchman Flat. Violin Breccia to right of hammer; sheared and broken basement to left.

Air view of Ridge Basin with Frenchman Flat at lower left and Frazier Mountain in background. Old Highway 99 shown along course of Piru Creek before construction of Pyramid Dam.

STRUCTURE AND LATE MIOCENE PALEOGEOGRAPHY OF SOUTHERN RIDGE BASIN, SOUTHERN CALIFORNIA

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ABSTRACT

Subsurface mapping based upon data from 68 exploratory wells drilled in southern Ridge Basin and northwestern Soledad Basin indicates that: southeast-trending structures of Ridge Basin, such as the Ridge Basin Syncline, merge with east- to northeast-trending structures of Soledad Basin near Castaic; major folds such as the Ridge Basin-Dry Canyon Syncline and the Dry Canyon Anticline formed prior to late Pliocene time, and possible isolated Ridge Basin from the eastern Ventura Basin; the late (?) Pliocene marine Pico Formation thins towards and pinches out against the Dry Canyon Anticline, suggesting that this fold formed a topographic high during Pico deposition; the Plio-Pleistocene nonmarine Saugus Formation unconformably overlies the Pico and Castaic formations near Castaic without major folding; the eastern segments of the Dry Canyon Syncline and Dry Canyon Anticline were reactivated after deposition of the Saugus Formation.

Four members or facies in the Castaic Formation are recognized. From west to east they are: 1) The Violin Breccia deposited in neritic to bathyal (?) depths adjacent to the San Gabriel Fault; 2) Mid-basin turbidite sandstones deposited down the axis of late Miocene Ridge Basin, at probably outer neritic to bathyal depths and more than 3000 ft. thick near Castaic; 3) Mudstone facies which intertongues with all other members, and probably deposited at outer neritic to bathyal depths; and 4) Basal transgressive facies comprising shallow marine, probably inner neritic, sandstone which onlap from west to east.

INTRODUCTION

Near the town of Castaic, Ridge, Soledad and Ventura basins are in close proximity to one another (Fig. 1). Although the structural relationship between the Ridge and the Soledad basins is not well understood, the dominant structural trend in Ridge Basin is northwest, whereas structures in Soledad Basin trend mainly east and northeast. There is no conclusive outcrop data to indicate how Ridge Basin structures interface with Soledad Basin structures.

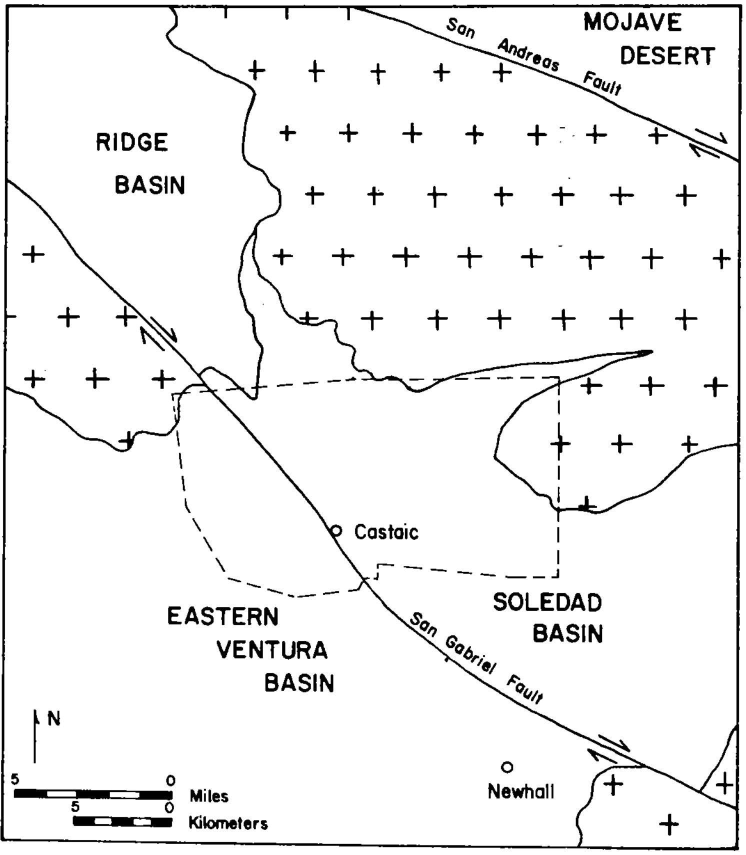

Figure 1. Regional geologic setting of the Castaic area. Areas marked with crosses represent pre-Oligocene units (after Jennings and Strand, 1969). Dashed-line boundary represents the area discussed in this paper.

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in Crowell, J.C., and Link, M.H., eds., 1982, **Geologic History of Ridge Basin, Southern California**, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 43-52.

The Ridge Basin Syncline is covered by alluvium and by nonmarine strata of the Plio-Pleistocene Saugus Formation near Castaic so the southern continuation of the Ridge Basin Syncline cannot be established from surface data alone.

STRATIGRAPHY

Northeast of Castaic, the middle to late Miocene Mint Canyon Formation unconformably overlies the Vasquez and San Francisquito Formations and the Pelona Schist (Figs. 2, 3). The Mint Canyon Formation was deposited in alluvial fan to lacustrine environments (Ehlig, et al., 1975); its thickness in the vicinty of Castaic ranges from 1200 to 2300 ft. Overlying the Mint Canyon Formation are marine rocks of the Castaic Formation. At the surface, the contact between the Castaic and Mint Canyon formations appears to be a low-angle unconformity. Alternatively, the contact may be a disconformity, and in the subsurface the two formations may locally intertongue (Stitt, in prep.).

Skolnick and Arnal (1959) described the mircrofauna which characterize the Castaic Formation. It is referred to informally as the Charlie Canyon fauna (CCF) by the petroleum industry, and is assigned to the Mohnian and Delmontian (?) Stages of Kleinpell (1938). The questionable Delmontian faunas suggest that the age of the Castaic Formation may extend into Pliocene time. However, Stanton (1966) concluded that the megafauna of the Castaic Formation is exclusively late Miocene in age. The thickness of the Castaic Formation in southern Ridge Basin ranges from approximately 1700 ft. to approximately 9000 ft. East of San Gabriel Fault, the Pliocene Pico Formation is only present in the subsurface where it overlies the Castaic Formation with angular unconformity. The Pico is overlapped by the Saugus Formation so that Saugus rests upon Castaic Formation at the surface. The Pico pinch-out is productive in Tapia and Wayside Canyon oil fields (Miller and Turner, 1959; Dosch and Beecroft, 1959; Mefferd and Johnson, 1967). The zero edge of the Pico Formation east of the San Gabriel Fault is shown on Figure 5. Near Castaic, the Pico Formation is mainly of shallow-marine origin, and the thickness ranges from 0 to 800 ft. East of the San Gabriel Fault, the Plio-Pleistocene Saugus Formation unconformably overlies both the Pico and Castaic Formations. The base of the Saugus Formation locally comprises marine to brackish water deposits of the Sunshine Ranch Member (Winterer and Durham, 1962). The remainder of the Saugus Formation is nonmarine and represents alluvial fan deposition. The maximum thickness of the Saugus Formation east of the San Gabrield Fault is approximately 2000 ft.

Structure of Southern Ridge Basin

Sixty-eight relatively deep exploratory wells for oil and gas have been drilled in southern Ridge Basin and northwestern Soledad Basin. The major objective for most of these wells was to test sandstones of the Castaic Formation (Stitt, in prep.). Thirty-five of the wells penetrated the Castaic Formation and drilled into the underlying Mint Canyon Formation. Using data from these wells, the top of the Mint Canyon Formation is contoured to illustrate the pre-Saugus structure of southern Ridge Basin (Fig. 4). Bedding attitudes from outcrops of the Castaic Formation guided contouring where there were no subsurface data. Structure is mainly inferred along the San Gabriel Fault because no wells penetrated into the Mint Canyon Formation.

Based on this subsurface mapping, it can be shown that the Ridge Basin Syncline changes trend from northwest to east near Castaic and that it merges with what is mapped at the surface as the Dry Canyon Syncline of Stitt (1980). The Dry Canyon Anticline also exhibits this change from northwest to east trend. Further northeast of Castaic, Sams (1964) mapped a number of minor folds at the surface

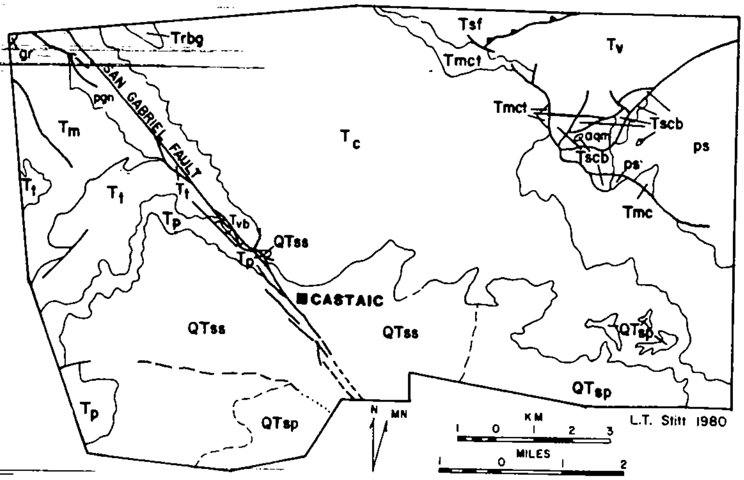


Figure 2. Generalized geologic map of the Castaic area (after Stitt, 1980, pl. 1); gr:granite; pgn:Palomas Gneiss; aqm:aplitic quartz monzonite; ps:Pelona Schist; Tsf:San Francisquito Fm; Tv:Vasquez Fm; Tscb:San Francisquito Canyon Breccia; Tmc:Mint Canyon Fm; Tmct:Taylor Fanglomerate member of the Mint Canyon Fm; Tc:Castaic Fm; Tvb:Violin Breccia; Tm:Modelo Fm; Tt:Towsley Fm; Trbg:Ridge Basin Group; Tp:Pico Fm; QTss:Saugus Fm (San Francisquito ss clasts); QTsp:Saugus Fm (Pelona Schist clasts).

which change trend from northeast to northwest as they approach Ridge Basin Syncline (Fig. 4).

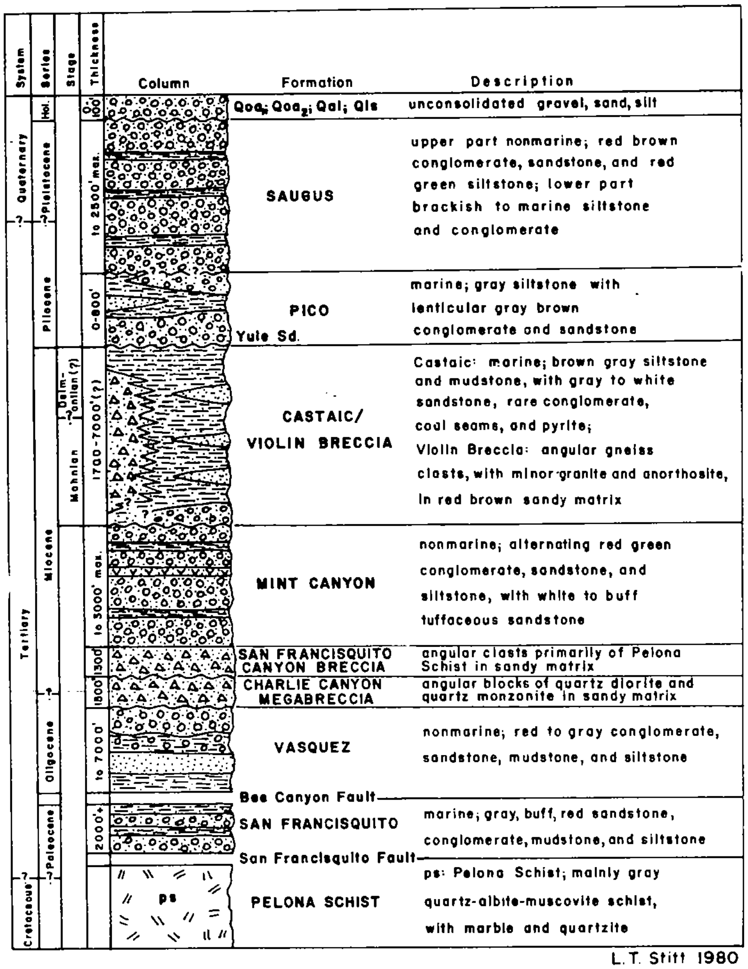
Subsurface investigations also show that folding of the Ridge Basin Syncline occurred prior to deposition of the Saugus Formation (Fig. 5). Although Weber (1979) indicated possible synclinal folding of terrace material near Castaic, there is only very minor warping of the Saugus Formation along the Ridge Basin Syncline at this locality (Fig. 5). Also, folding of the Saugus Formation along the Dry Canyon Syncline and Dry Canyon Anticline dies out from east to west so that the base of the Saugus Formation is essentially a southwest-dipping homocline near Castaic.

The pre-Saugus folding of the Ridge Basin Syncline is also shown in cross section A-A' (Fig. 6; location of cross section shown on Figs. 4, 7). Electric log markers in Texaco Yule #1 and Henry King Urtasun #1 show the existence of the Ridge Basin Syncline, whereas the Saugus Formation is unfolded. The subsurface pinch-out of the Pico Formation is also shown on cross section A-A'. It appears that the Dry Canyon Anticline may have been a topographic high during deposition of the Pico Formation inasmuch as isopachs of the Pico Formation east of the San Gabriel Fault are parallel to the trend of the Dry Canyon Anticline and show a wedge-out of the Pico Formation against this structure (Stitt, 1980, pl VI).

PALEOGEOGRAPHY OF THE CASTAIC FORMATION

Stanton (1960) recognized four members of the Castaic Formation, listed here from bottom to top, they are: 1) Basal member which comprises the eastern and northeastern outcrops of the Castaic Formation; shallow-marine deposition; includes inhomogeneous strand deposits of conglomerate and sandstone with abundant megafossils; basal member onlapped from west to east during late Miocene marine transgression; 2) Mudstone member which generally occurs southwest of outcrops of the basal member; probable outer neritic to bathyal paleobathymetry; 3) Mid-basin sandstone member intertongueing with the mudstone member; limited to the central, thickest portion of the Castaic Formation; sandstones exhibit graded bedding, mudstone rip-ups, contorted bedding, southwestward-directed slump structures; possible outer neritic to bathyal paleobathymetry; 4) Violin Breccia which occurs along the southwest margin of the basin adjacent to the San Gabriel Fault; fault-scarp talus breccia; paleobathymetry unknown, possibly bathyal (?).

Stanton (1960) considered the mid-basin sandstone member to have been deposited as the late Miocene sea regressed or as the basin shoaled. It is more likely that Stanton's mid-basin sandstones are turbidites (Link and Osborne, 1978). These turbidity currents may have flowed down the axis of Ridge Basin parallel to and adjacent to the San Gabriel Fault.



L.T. Stitt 1980

Figure 3. Generalized columnar section east of the San Gabriel Fault (not to scale).

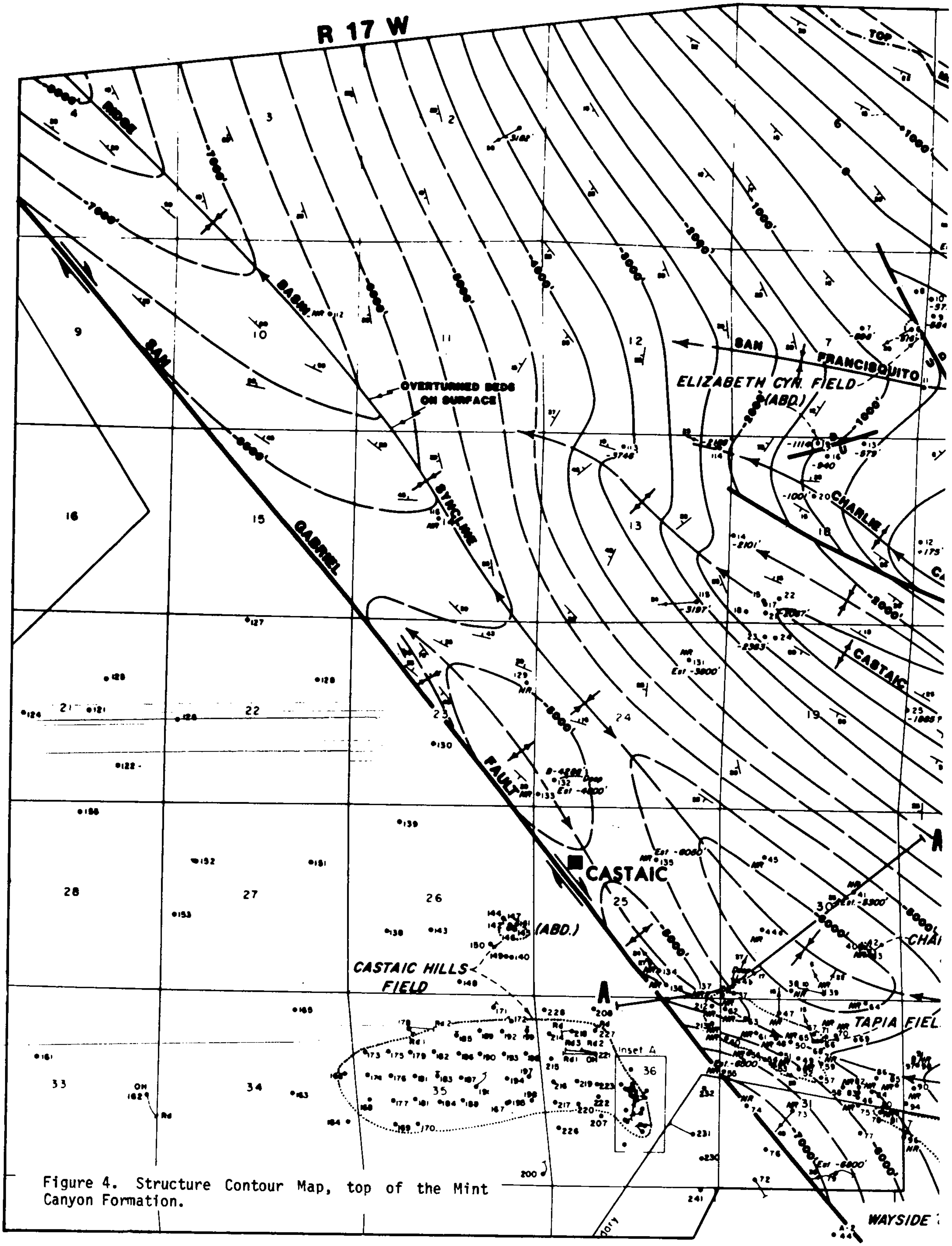
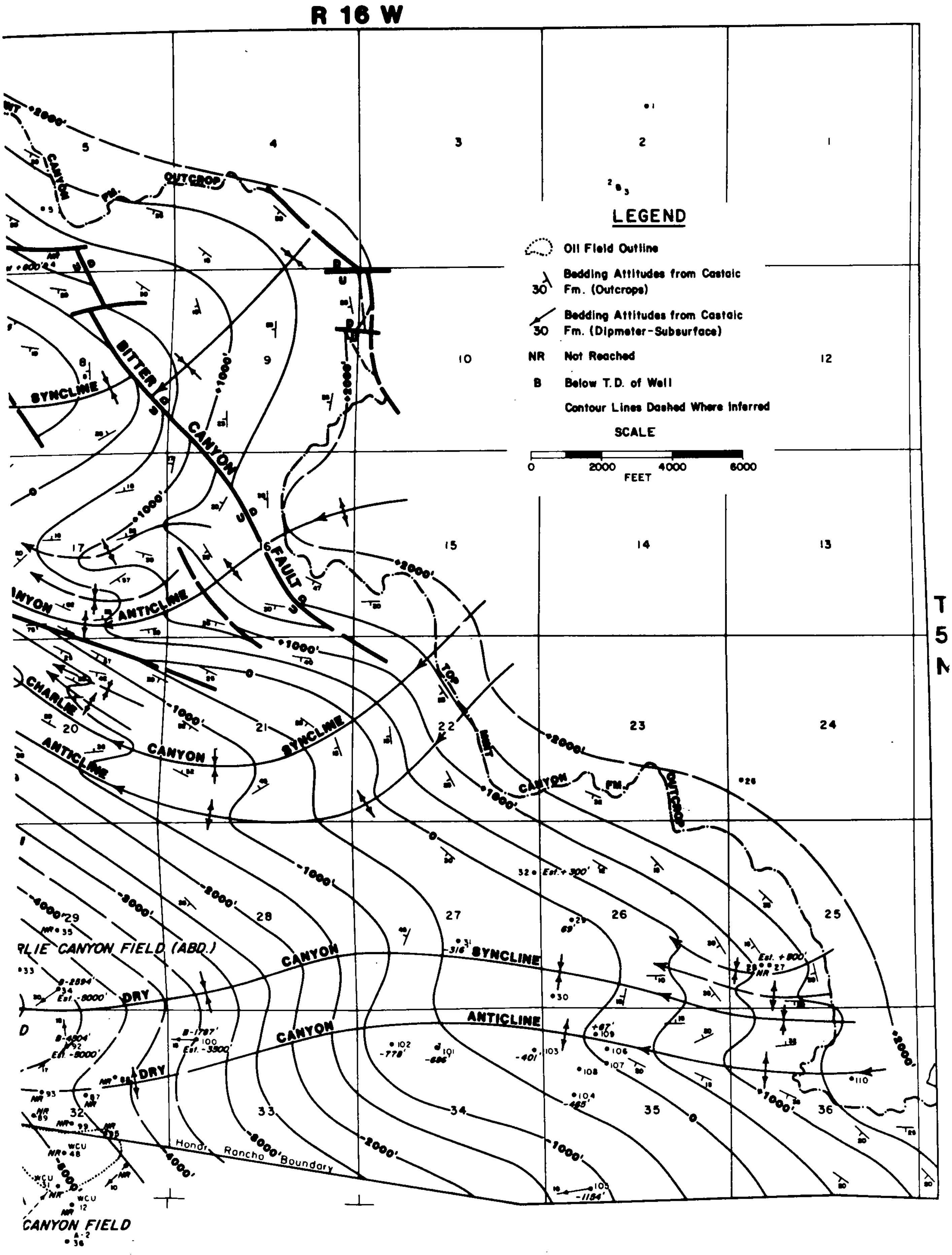


Figure 4. Structure Contour Map, top of the Mint Canyon Formation.


R 16 W
T
5
N
LEGEND
Oil Field Outline
30 Bedding Attitudes from Castaic Fm. (Outcrops)
30 Bedding Attitudes from Castaic Fm. (Dipmeter-Subsurface)
NR Not Reached
B Below T.D. of Well
Contour Lines Dashed Where Inferred
SCALE
0
2000
4000
6000
FEET
OUTCROP
CANYON FM.
BITTER CANYON FAULT
SYNCLINE
CANYON ANTICLINE
CHARLIE CANYON ANTICLINE
CANYON SYNCLINE
TOP
CASTAIC
CANYON FM. OUTCROP
CHARLIE CANYON FIELD (ABD.)
DRY CANYON SYNCLINE
DRY CANYON ANTICLINE
Honor Rancho Boundary
CANYON FIELD
2000'
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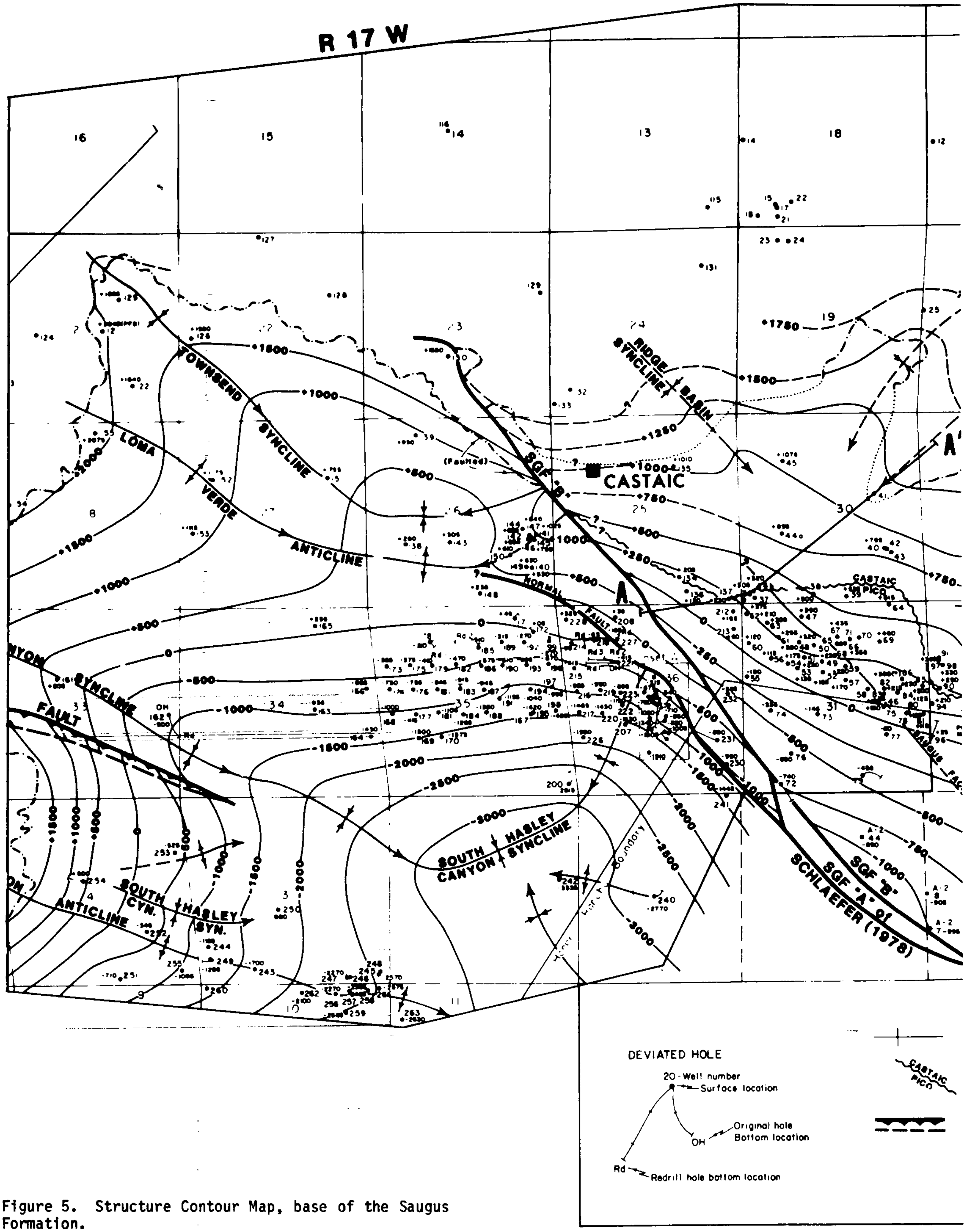
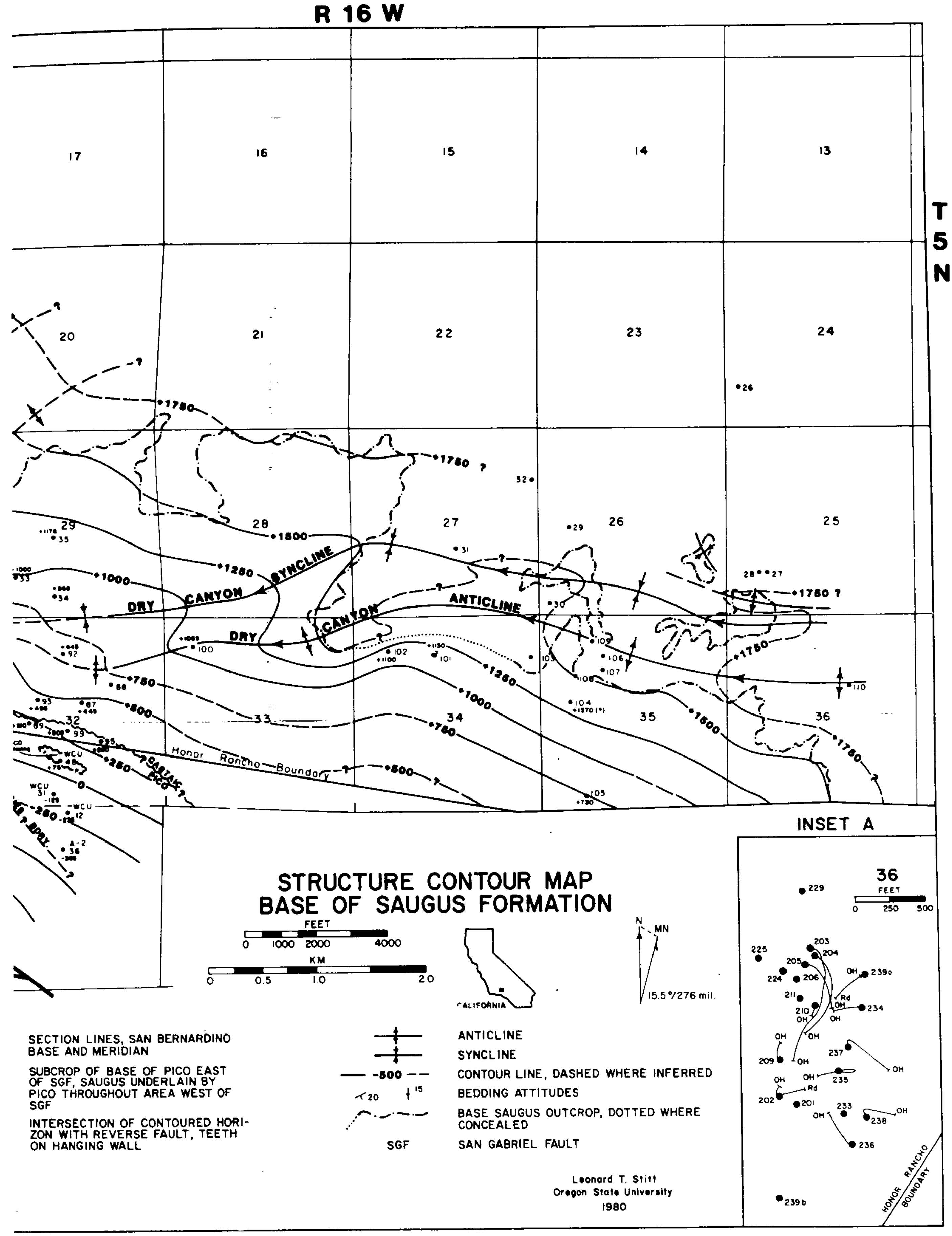


Figure 5. Structure Contour Map, base of the Saugus Formation.



STRUCTURE CONTOUR MAP
BASE OF SAUGUS FORMATION

SECTION LINES, SAN BERNARDINO BASE AND MERIDIAN	
SUBCROP OF BASE OF PICO EAST OF SGF, SAUGUS UNDERLAIN BY PICO THROUGHOUT AREA WEST OF SGF	
INTERSECTION OF CONTOURED HORIZON WITH REVERSE FAULT, TEETH ON HANGING WALL	
	ANTICLINE
	SYNCLINE
-500	CONTOUR LINE, DASHED WHERE INFERRED
20 15	BEDDING ATTITUDES
	BASE SAUGUS OUTCROP, DOTTED WHERE CONCEALED
SGF	SAN GABRIEL FAULT

Leonard T. Stitt
Oregon State University
1980

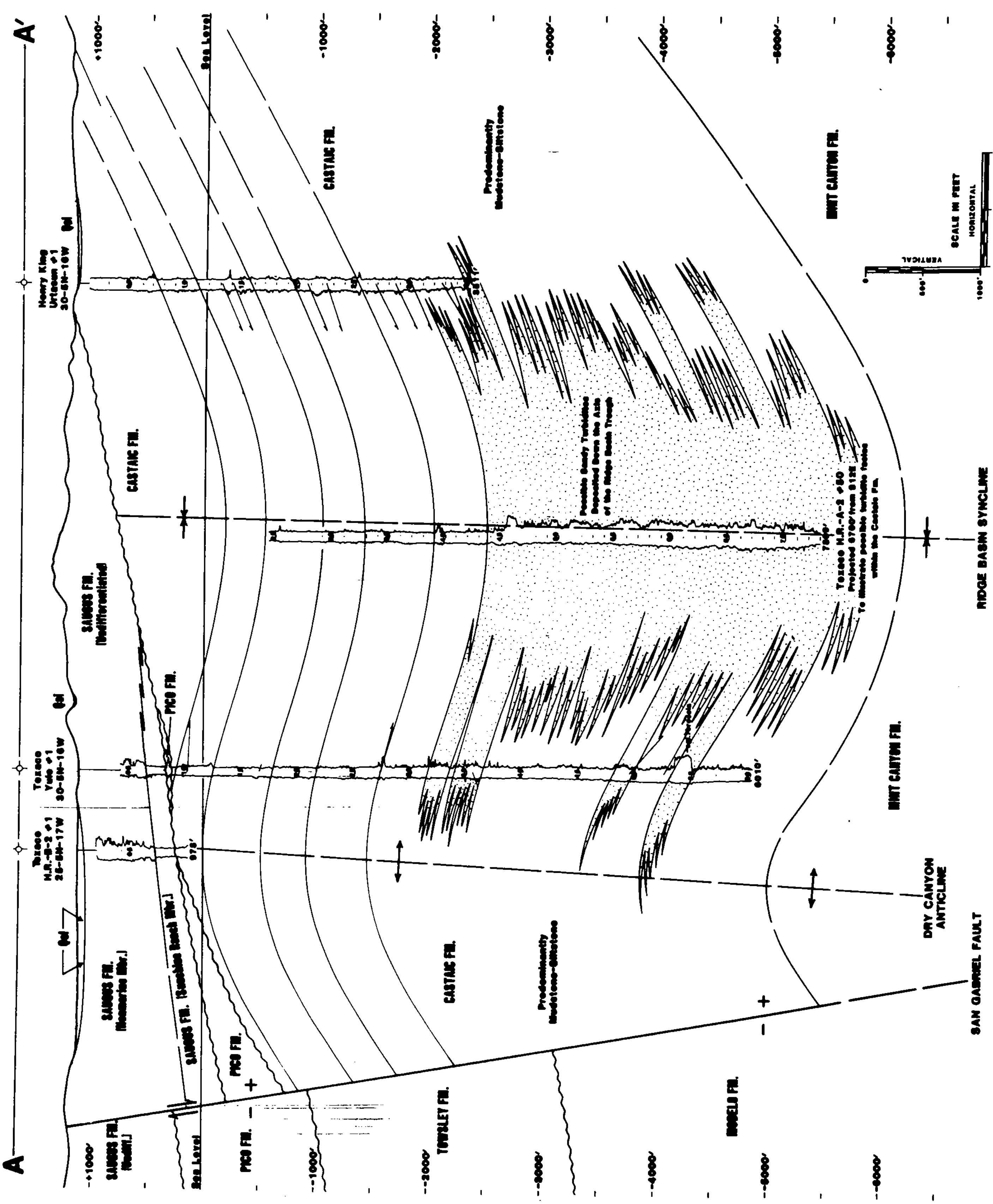


Figure 6. Cross section demonstrating pre-Saugus folding of Ridge Basin Syncline. Pico Formation deposition possibly controlled by the Dry Canyon Anticline. Possible facies within the Castaic Formation also are illustrated (Stitt, in prep.).

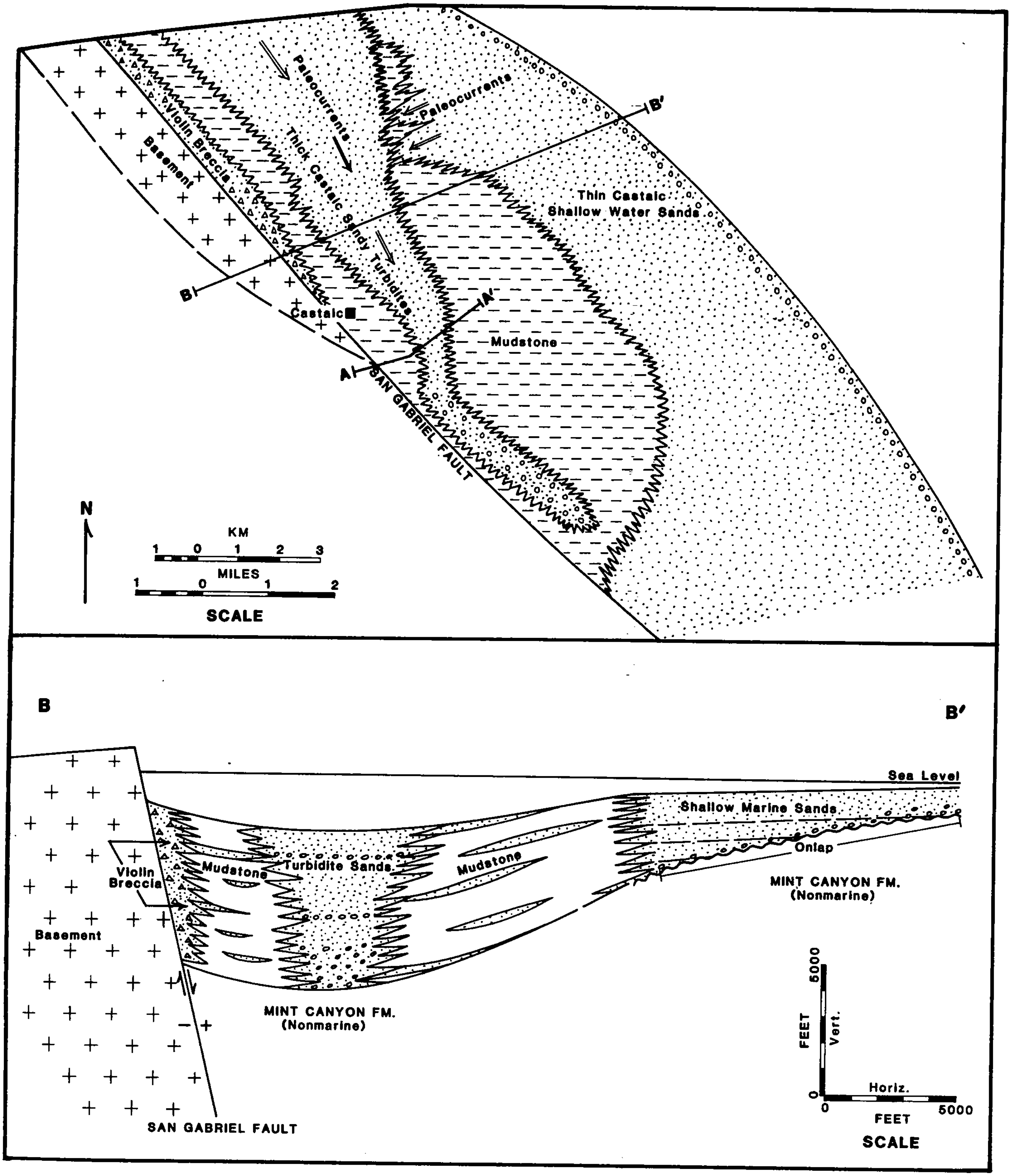


Figure 7. Paleogeography of the Castaic Formation during Mohnian (Late Miocene) time. Area north of Castaic modified after Stanton (1960) and Link and Osborne (1978).

In the subsurface near Castaic, a number of wells penetrate a sandstone- and conglomerate-rich portion of the Castaic Formation. One of these wells, Texaco Honor Rancho A-NCT-2 #50 (well #79 in 31-5N-16W, Figs. 4, 5) is projected onto cross section A-A' (Fig. 6) to illustrate possible facies relationships within the lower part of the Castaic Formation. Figure 7 illustrates diagrammatically possible facies relationships and paleogeography of the Castaic Formation during Mohnian (late Miocene) time (Stitt, in prep.), and is consistent with the interpretation of Link and Osborne (1978).

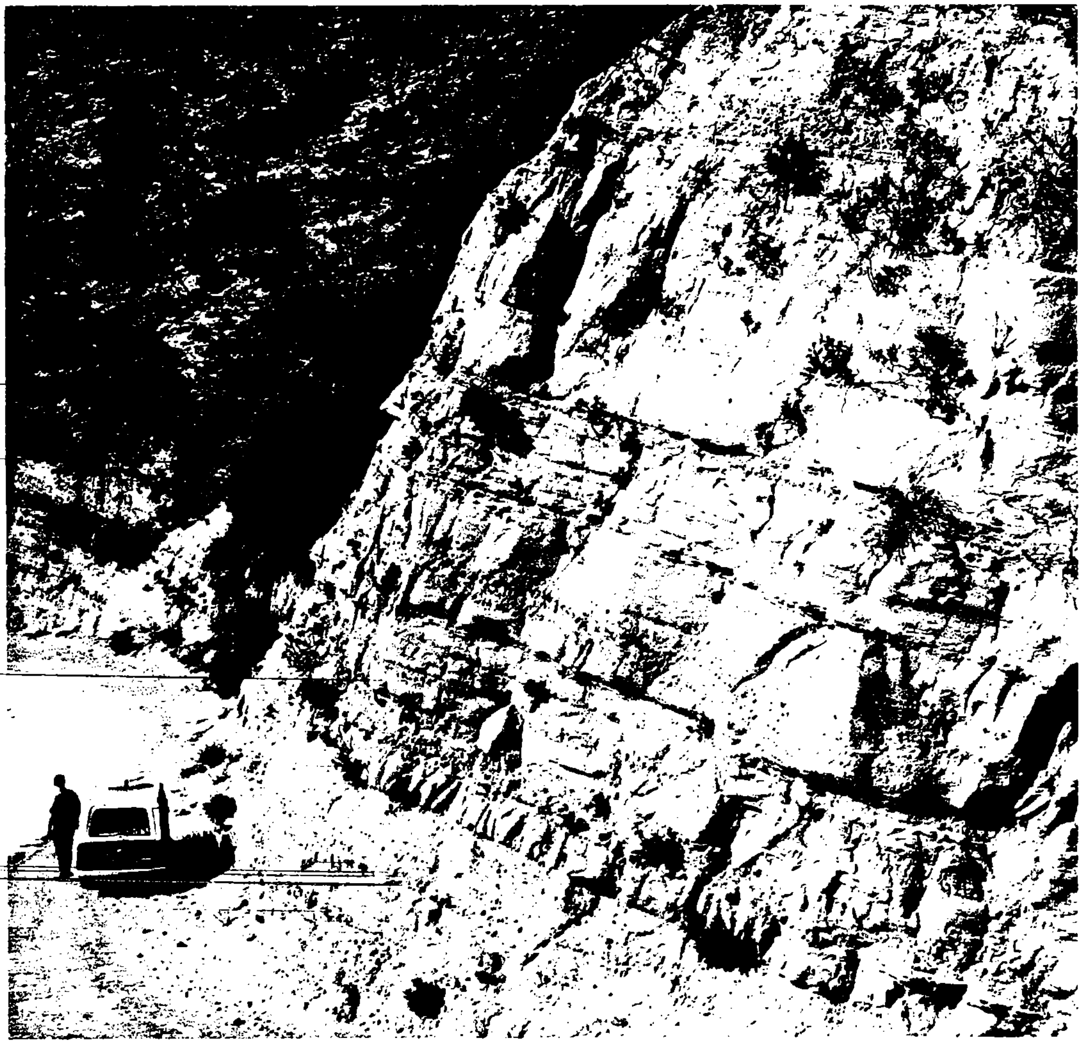
SUMMARY

Subsurface investigations demonstrate that northwest-trending structures of Ridge Basin merge into east- and northeast-trending structures of Soledad Basin near the town of Castaic. Southern Ridge Basin and northwestern Soledad Basin were part of the same province during deposition of the late Miocene Castaic Formation.

Major folds such as the Ridge Basin-Dry Canyon Syncline and the Dry Canyon Anticline formed prior to late Pliocene time, and possible isolated Ridge Basin from eastern Ventura Basin. The late Pliocene marine Pico Formation thins towards and wedges out against the Dry Canyon Anticline, and the Plio-Pleistocene nonmarine Saugus Formation unconformably overlies the Castaic and Pico formations without major folding near Castaic. The eastern segments of the Dry Canyon Anticline and Dry Canyon Syncline were reactivated after deposition of the Saugus Formation.

ACKNOWLEDGEMENTS

Most of this work was done as part of a master's thesis at Oregon State University. That project was funded under U. S. Geological Survey Contract 14-08-0001-16747 of the Earthquake Hazards Reduction Program. The paper was reviewed by Robert S. Yeats of Oregon State University and Kevin J. Lant of Conoco. Drafting of Figures 4 and 5 was done by Adolph Losawyer of Conoco. Steve Wiler of Conoco also provided drafting assistance.



Exposure of Ridge Route Formation along Old Ridge Route, Ridge Basin, southern California

SEDIMENTARY FACIES OF RIDGE BASIN, SOUTHERN CALIFORNIA

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ABSTRACT

Ridge Basin is a linear trough formed in a wrench-fault setting and filled with about 13,000 m (44,000 ft) of marine and nonmarine sedimentary rocks. The water depth in the basin was shallow and was never more than 100 m deep. First marine and then nonmarine facies were deposited in Ridge Basin. The Ridge Route Formation on the northeast, the Hungry Valley Formation on the north, and the Violin Breccia on the southwest comprise the basin margin units. These marginal strata consist of alluvial fan-fluvial, fluvial-deltaic, and shoreline facies which intertongue and are transitional into one another. Basin margin sedimentation was toward the center of the basin with periodic progradation across the basin. Source terranes of the northeast side of the basin yielded the greatest volume of sediment. The Castaic and Peace Valley Formations as well as certain members of the Ridge Route Formation are basin axis deposits, which include offshore, fluvial-deltaic, and turbidite facies. Axial sedimentation includes mudstone derived from the margins of the basin, and fluvial-deltaic and turbidite deposition down the axis of the basin to the south-southeast. The Hungry Valley Formation marks the termination of sedimentation in Ridge Basin. These dominantly alluvial and fluvial deposits filled the basin from the north.

The duration of Ridge Basin was at least 6 m. y., and it is characterized by one of the world's highest sediment accumulation rates (~2 m/1000 years). The source terrane for the Violin Breccia is to the southwest and consists of nearly equal amounts of granitic and metamorphic rocks. The Ridge Route, Castaic, and Hungry Valley Formations' source terranes are to the northeast and north and consist of granitic, metamorphic, volcanic, quartzite, and older sedimentary rocks.

INTRODUCTION

Ridge Basin is the best exposed basin along the San Andreas transform belt and affords an excellent opportunity to observe marine and nonmarine facies in a wrench-fault setting. Ridge Basin is filled with over 13,000 m (44,000 ft) of marine and nonmarine sedimentary rocks deposited in a relatively short period of geologic time (6 m. y.). Since the strata in Ridge Basin are characterized by numerous rapid facies changes, the purpose of this paper is to summarize the general facies relationships in Ridge Basin and to update an earlier paper (Link and Osborne, 1978). The reader is referred to later papers in this volume dealing with individual formations and members for more complete descriptions and interpretations.

Sediments of Ridge Basin were deposited in marine and nonmarine environments adjacent to areas of high relief which at times contributed large quantities of clastic material. Two major styles of sedimentation occur in the Ridge Basin: basin margin and axial deposition. The facies, lithologies, sedimentary features and fossils which characterize these styles of sedimentation are summarized in Figure 1. Basin margin deposition includes alluvial fan-fluvial, fluvial-deltaic, shoreline, and nearshore facies, whereas axial deposition includes turbidite, offshore, and fluvial-deltaic facies. The marine and nonmarine environments have similar sedimentation styles, facies and lithologies and merely reflect a change in fauna, salinity, and water depth in the basin. Where applicable, the marine and nonmarine environments are discussed together.

Marine and Nonmarine Sedimentation

Sedimentation began in Ridge Basin within a marine embayment which connected to the south with the Ventura Basin by way of a connection now eroded away from the high country northwest of Castaic. The marine Castaic Formation and lower part of the Ridge Basin Group were deposited in the central and northeastern part of the basin, whereas the Violin Breccia accumulated adjacent to the San Gabriel Fault on the southwest side of the basin. At this time, Ridge Basin was a relatively shallow-marine basin (<100 m water depth) in which turbidite and offshore facies accumulated in the center of the basin and shoreline facies formed along the margins (Fig. 2A). Alluvial fans extended into the basin from the east (Mint Canyon Formation) and west (Violin Breccia). The Mint Canyon Formation was onlapped early in its history by the Castaic Formation. The Violin Breccia continued to accumulate along the San Gabriel Fault, interfingering first with the Castaic Formation and second with the various other members of the Ridge Basin Group.

With continued strike-slip displacement along the San Gabriel Fault, the marine embayment to the south was either blocked and the basin shoaled, filling with nonmarine rocks, or the basin was moved inland along the San Gabriel Fault, being filled continuously with nonmarine rocks. At this time, Ridge Basin was a low-lying nonmarine coastal basin with drainage to the south-southwest. Nonmarine deposits, namely alluvial fan-fluvial, fluvial-deltaic, and lacustrine shoreline, nearshore, offshore and turbidite facies filled the basin (Fig. 2B). Small alluvial fans extended into the basin from the southwest (Violin Breccia) with much larger alluvial fan-fluvial complexes entering the basin from the northeast side (Ridge Route Formation). Basin margin fluvial-deltaic, lacustrine shoreline and nearshore facies interfinger with axial lacustrine offshore, turbidite, and fluvial-deltaic facies on the west side of the basin adjacent to the San Gabriel Fault. These nonmarine rocks make up most of the Ridge Basin Group. Sediment transport was primarily to the west-southwest toward the San Gabriel Fault but with moderate transport to the south-southeast down the basin axis. With the termination strike-slip displacement upon the San Gabriel Fault, coupled with infilling of the basin by the Hungry Valley Formation, Ridge Basin ceased to be a depositional basin.

in Crowell, J.C., and Link, M.H., eds., 1982, **Geologic History of Ridge Basin, Southern California**, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 63-78.

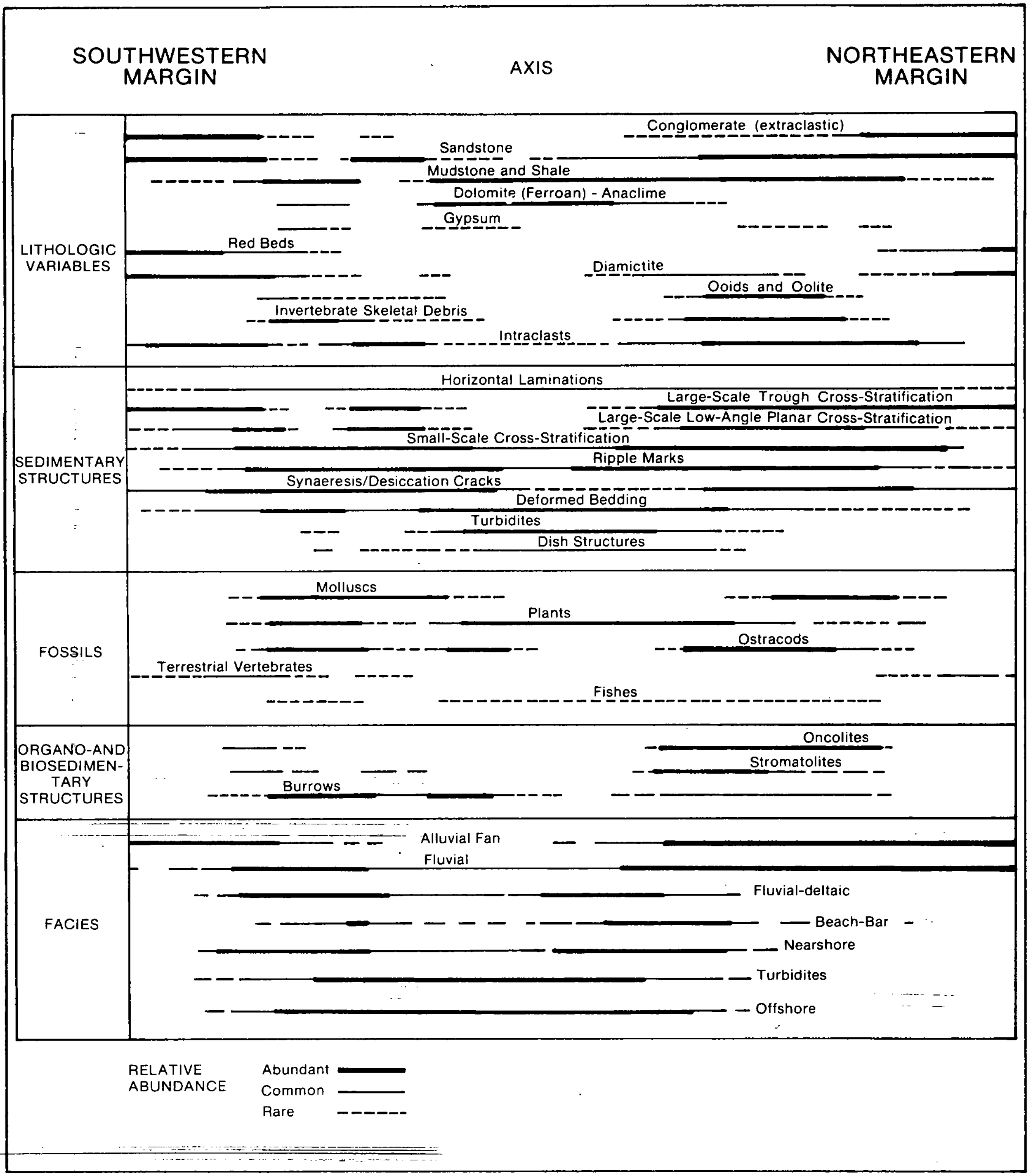

Figure 1. Lithologies, sedimentary structures, fossils and facies relationships.

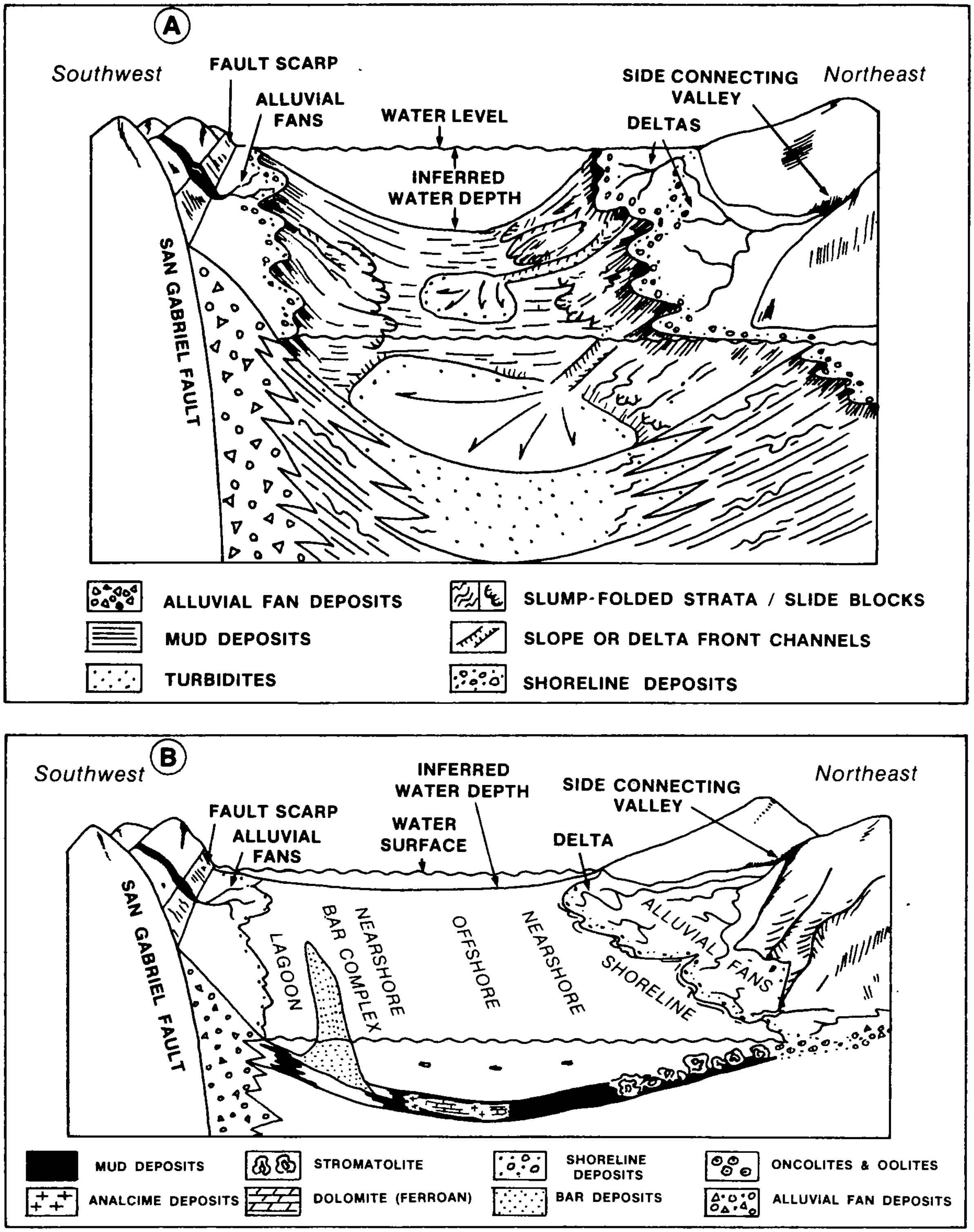

Figure 2. Deep-water and shallow-water depositional environment diagrams. A, deep-water; and B, shallow-water facies relationships.

Basin Margin Deposition

Alluvial Fan-Fluvial Facies

Coarse-grained fluvial sediments were deposited as a series of coalesced alluvial fan complexes along the northeastern and southwestern margins of Ridge Basin (Fig. 2). Thick accumulations of interstratified conglomerate, sandstone and breccia characterize the alluvial fan complexes that interfinger with finer-grained mudstone, shale and sandstone toward the center of the basin. The eastern and western alluvial fans differ markedly with respect to size, composition, texture and distribution, therefore, these will be discussed separately.

Eastern alluvial fan-fluvial facies. Along the eastern margin, the alluvial fan-fluvial facies (Ridge Route Formation) is very coarse-grained (gravel- to cobble-sized clasts are common), contains well-rounded to angular clasts, is moderately to poorly sorted, lacks well-defined stratification, and is devoid of fossils (Figs. 3,4). Well-rounded, friable basement clasts (Fig. 3A) are a common constituent. Near the shoreline, the alluvial fan facies is finer grained (sand- to gravel-sized clasts predominate), well stratified (Figs. 3B,C) and moderately to well sorted; includes tabular beds of cross-stratified and horizontally-laminated sandstone (Fig. 3D), erosive channels filled with conglomerate (Fig. 3E) and cross-stratified sandstone; and contains load and flame structures (Fig. 3F), convolute laminae, ripple marks, mudcracks, and a few vertebrate remains.

At times, perhaps during seasonal flooding, coarse-grained sediment was transported by braided streams on the alluvial fan complexes into the basin. These streams greatly modified the shoreline lacustrine and marine environments (Fig. 5). Such sediment influxes are indicated by channeling and by Basal conglomerate or sandstone containing intraclasts of stromatolites, mudstone, siltstone, and sandstone (Figs. 3G,H,I; 5B). Thinning- and fining-upward cycles characterize these deposits (Fig. 4C).

The eastern alluvial fan-fluvial facies consists of light-colored granitic debris and minor amounts of volcanic, quartzite, and sandstone clasts derived from adjacent basement terrane and from Cretaceous to Paleocene strata along the eastern side of the basin. These clasts may have been derived even farther east from the Orocopia Mountains region (Ehlert and Ehlig, 1977). Local red fanglomerate and paleosoils are found around basement highs and along major vertical faults on this side of the basin. These sequences thicken- and coarsen-upward (Fig. 4C). This facies contains the greatest volume of sedimentary rock in Ridge Basin and a great abundance and diversity of sedimentary structures. The well-rounded nature of some of the basement clasts suggest reworking or moderate transport on large alluvial fan-fluvial complexes or by a large fluvial channel(s) from the east (Ehlert and Ehlig, 1977).

Western alluvial fan. Along the western margin, the alluvial fan facies (Violin Breccia) consists of narrow but continuous exposures of dark-colored breccia, conglomerate and diamictite (Figs. 6 7) (Crowell, this volume). The clasts in the Violin Breccia consist primarily of gneiss and granitic basement. Near the San Gabriel Fault, the Violin Breccia lacks well-defined stratification (Fig. 6A); is poorly sorted (Fig. 6B); sheared and fractured; devoid of fossils; displays few sedimentary structures, and the clasts are angular and large (up to 2 m in diameter). This formation is extremely thick (11,000 m or 36,000 ft) and extends as a band along the San Gabriel Fault for about 30 km with a width of 1,500 m. The relatively small fans of the Violin Breccia dramatically interfinger with axial sediment a short distance (500-1000 m) into the basin where it changes into finer-grained facies (Figs. 6C,D). These fine-grained strata are dark gray to red, mudcracked, cross bedded, parallel laminated, burrowed, and contain occasional stromatolitic horizons.

The Violin Breccia formed as talus, debris flow, and landslide deposits along the San Gabriel Fault scarp (Fig. 7). With continued strike-slip to oblique-slip displacement along this fault, sediment was deposited continuously along this narrow belt. The western side of the basin was near the axis of the Ridge Basin trough and shows the greatest composite thickness of sedimentary rock which probably was related to local subsidence along this dominantly strike-slip margin (Crowell, this volume).

Shoreline facies

The shoreline environments in Ridge Basin are interrelated and complex. Beach, nearshore, bar, and fluvial-deltaic facies have been identified. These facies occur on both sides of the basin and are repetitive within this thick section. Examples of each environment are discussed below.

Beach facies. Terrigenous and carbonate sediments were deposited and locally reworked along the margins of Ridge Basin. The beach facies is especially common and well defined along the east side of the basin where broad alluvial fans from the east extended into the center of the basin (Fig. 5). Well-sorted, horizontally-laminated and low-angle, large-scale, cross-stratified sandstone interbedded with and transitional to basin mudstone characterize this facies (Figs. 8C,E,F). The following features are common in the beach facies: small-scale cross stratification; climbing ripples; ripple marks (Fig. 8C); flaser bedding; desiccation cracks; ooids and pisoliths; and lag conglomerate. Associated fossils include molluscs (some in growth position), ostracodes, vertical burrows, oncolites, rootlets and other plant remains (Fig. 8F). Wave agitation and local reworking is indicated by cross-stratified and horizontally-laminated sandstone and intraclast lag conglomerate. A narrow sandy beach is suggested because of the high local relief in Ridge Basin and the general coarseness of the detrital sediment.

Nearshore facies. Carbonate and terrigenous sediments deposited in nearshore environments display the greatest lithologic diversity of any facies (Figs. 3G,H,I; 4A,B). This facies commonly includes blue-gray mudstone, fine- to medium-grained sandstone, fossils, ooids, pisoliths, oncolites, stromatolites (Figs. 3H,I), and intraclasts. Strata are horizontally laminated, cross stratified and massively bedded, and are from poorly to well-sorted and commonly burrowed. Conglomerate- and sand-filled channels and desiccation cracks are present. Stromatolites are most common in this facies (Link, et al., 1977) and are associated with molluscan, ostracode and plant remains. Ooids, pisoliths, oncolites, and ostracode-bearing strara are interbedded with the stromatolites and are transitional

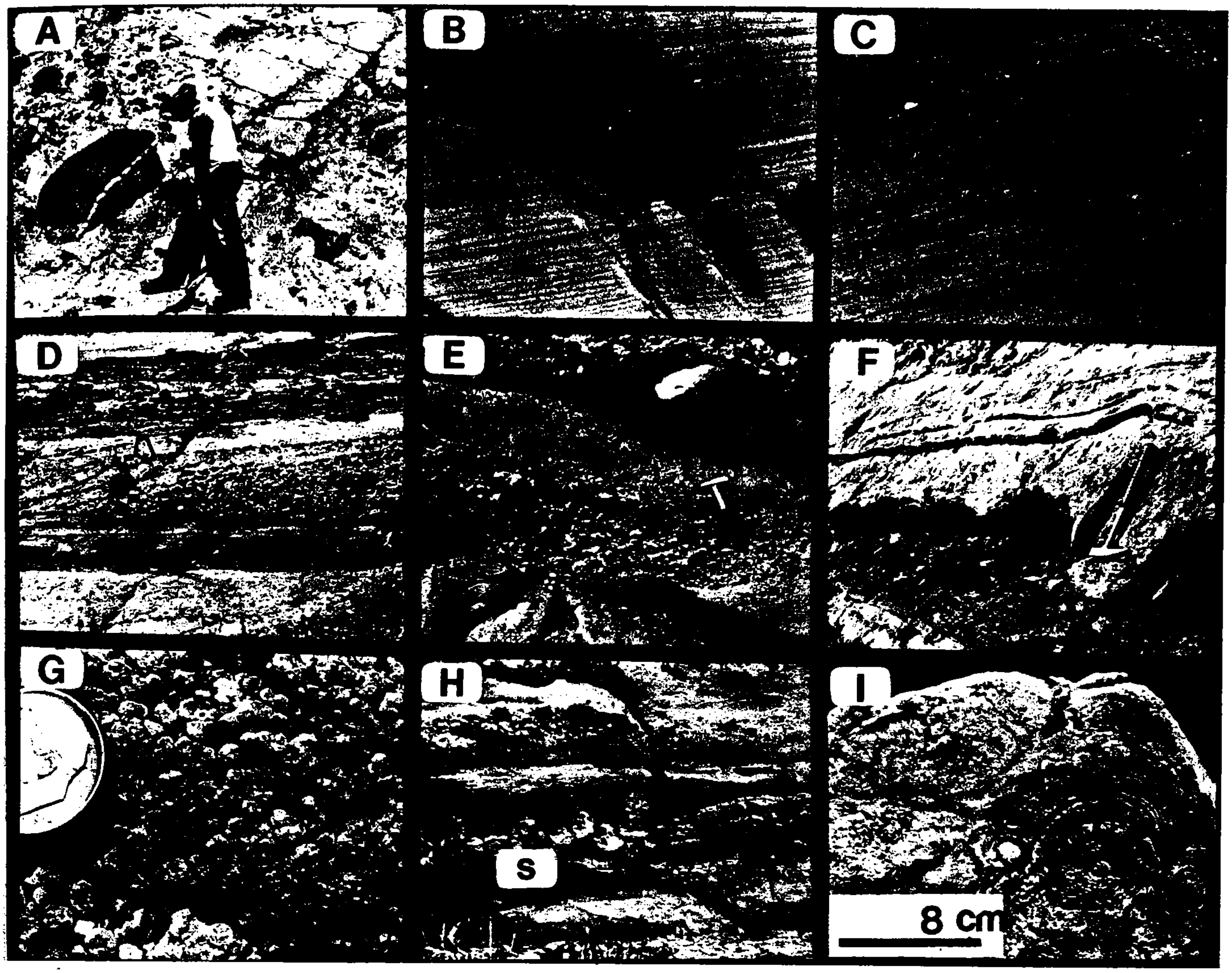

Figure 3. Photographs of eastern alluvial fan-fluvial and shoreline facies. A, basal conglomerate with subordinate breccia; B, parallel-laminated sandstone; C, low-angle tabular and festoon cross bedding in conglomeratic sandstone; D, tabular cross-bedded conglomerate cutting into mudstone; E, U-shaped channel conglomerate cutting into sandstone; F, load structures in lower part of a sandstone; G, ooid and oncolites; H, stromatolites (S) in an oncolitic sandy mudstone overlain by a conglomeratic sandstone; and (I) domal stromatolite with concentric layers.

into adjacent facies. Locally, slump-folded strata and load features are common.

The nearshore facies is transitional between the shoreline and offshore facies, and is characterized by at least three depositional features (Fig. 8B): (1) Coarse-grained sediment influx by fluvial processes into the alluvial fan and shoreline facies is indicated by channeling, textural changes, and the incorporation of stromatolites, ooids, pisoliths, and oncolites into dominantly terrigenous sediment. (2) Wave agitated, shoreline sedimentation in the nearshore facies is indicated by cross-stratified, well-sorted strata; oscillation ripple marks; the formation of ooids, pisoliths and oncolites; and the presence of intraclasts, which in part may reflect storm activity or fluctuations in water depth. (3) Subaerial exposure of at least the shallower parts of this facies is indicated by mudcracks, desiccation cracks within stromatolites, and the occurrence of intraclasts. Repetition of these three factors in response to climatic and tectonic conditions produced the shoreline facies and nearshore sedimentary sequences illustrated in Figure 26B.

Bar facies. Mainly along the western margin thin- to thick-bedded isolated packages of well-sorted, cross-bedded, white sandstone occur within thick molluscan-rich mudstone (Figs. 6E,F,G; 9A). These sandstone bodies are commonly convex upward with low-angle cross bedding dipping both on and offshore and contain mudstone rip-up clasts, pebble lags, molluscan debris, and burrows (Fig. 9A). Individual bar complexes have erosional to gradational lower contacts with low-angle accretionary (Fig. 8E), tabular, and festoon cross-bedding which grades upward into pebble and mudstone chip lags with small-scale cross-stratification (climbing ripples) and shell debris (Fig. 8F). The upper contact is commonly convex upward (Fig. 8G) with low-angle cross-bedding dipping both onshore and offshore, and gradational upward into blue-gray mudstone. These sandstone bodies are up to 3 m thick and at least from 10 to 30 m long. It is difficult to tell what

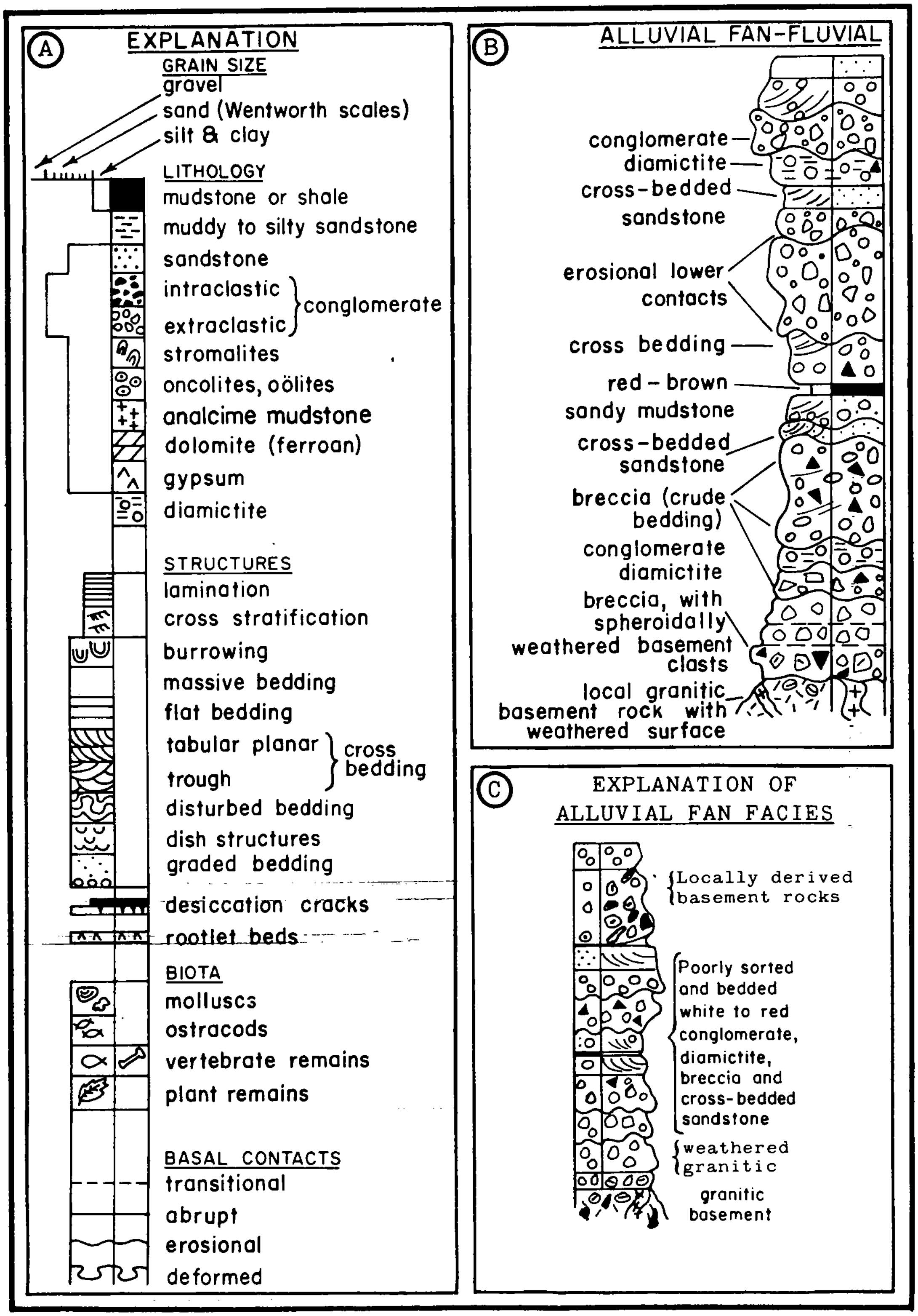

Figure 4. Explanation and schematic section of eastern alluvial fan-fluvial facies. A, explanation; B, alluvial fan-fluvial diagram; and C, explanation of alluvial fan facies.

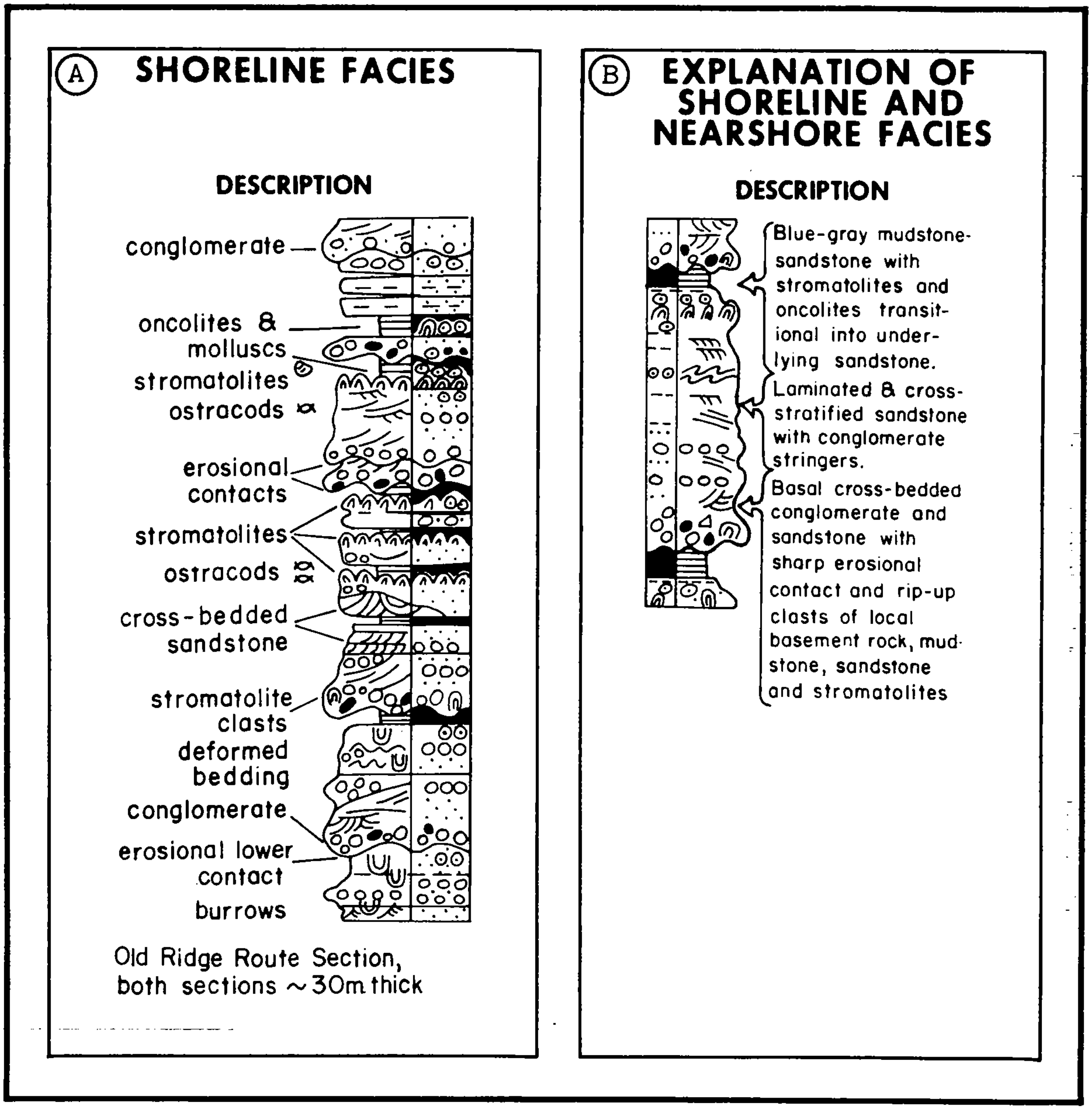

Figure 5. Schematic section of eastern shoreline facies. A, shoreline facies description; and B, explanation of shoreline and nearshore facies.

type of bar complex these represent, but their geometry and stratigraphic position within mollusc-bearing mudstone suggest these are most likely bar complexes deposited parallel to shoreline.

Fluvial-deltaic facies. Several well-developed fan deltas and/or modified Gilbert deltas are present in Ridge Basin (Figs. 8A,B; 9,B). They enter the basin from the east (Ridge Route Formation and Apple Canyon Sandstone Member) and pinch out into Peace Valley Formation. Individual depositional lobes are from 30 to 90 m thick and consist of sandstone interbedded with mudrock (Fig. 9B). These sequences commonly thicken- and coarsen-upward. The sandstone is white to blue-gray, fine- to coarse-grained, cross-bedded, ripple marked (Fig. 8C), burrowed, and commonly amalgamated. Slump folds, pull-apart beds, load structures, mudcracks (Fig. 8D) and rootlets are present. The mudstone is blue-gray to black, ripple marked, mudcracked, burrowed, and contains abundant organic material and rootlets. A typical vertical sequence (Fig. 9C) displays a few thin sandstones overlain by composite thick cross-bedded channel sandstone. This sequence is, in turn, overlain by massive and gradational sandstone and mudstone sequences which are commonly mudcracked. These cycles may be interrupted or repeated several times. These fluvial-deltaic complexes clearly entered shallow water and built into the basin, prograding over initial bottom set and foreset (?) strata. Locally thick fluvial sequences cap these complexes and with delta shifting or abandonment, thick deposits of interdistributary mud and sand accumulated. Slumping, contemporaneous faulting and the generation of turbidites occurred off the front and flanks of these prograding fan deltas.

Axial Deposition

The axial environments of Ridge Basin include three facies: (1) turbidite, (2) offshore, and (3) fluvial-deltaic. Each of these facies accumulated at different times in or near the center of the basin.

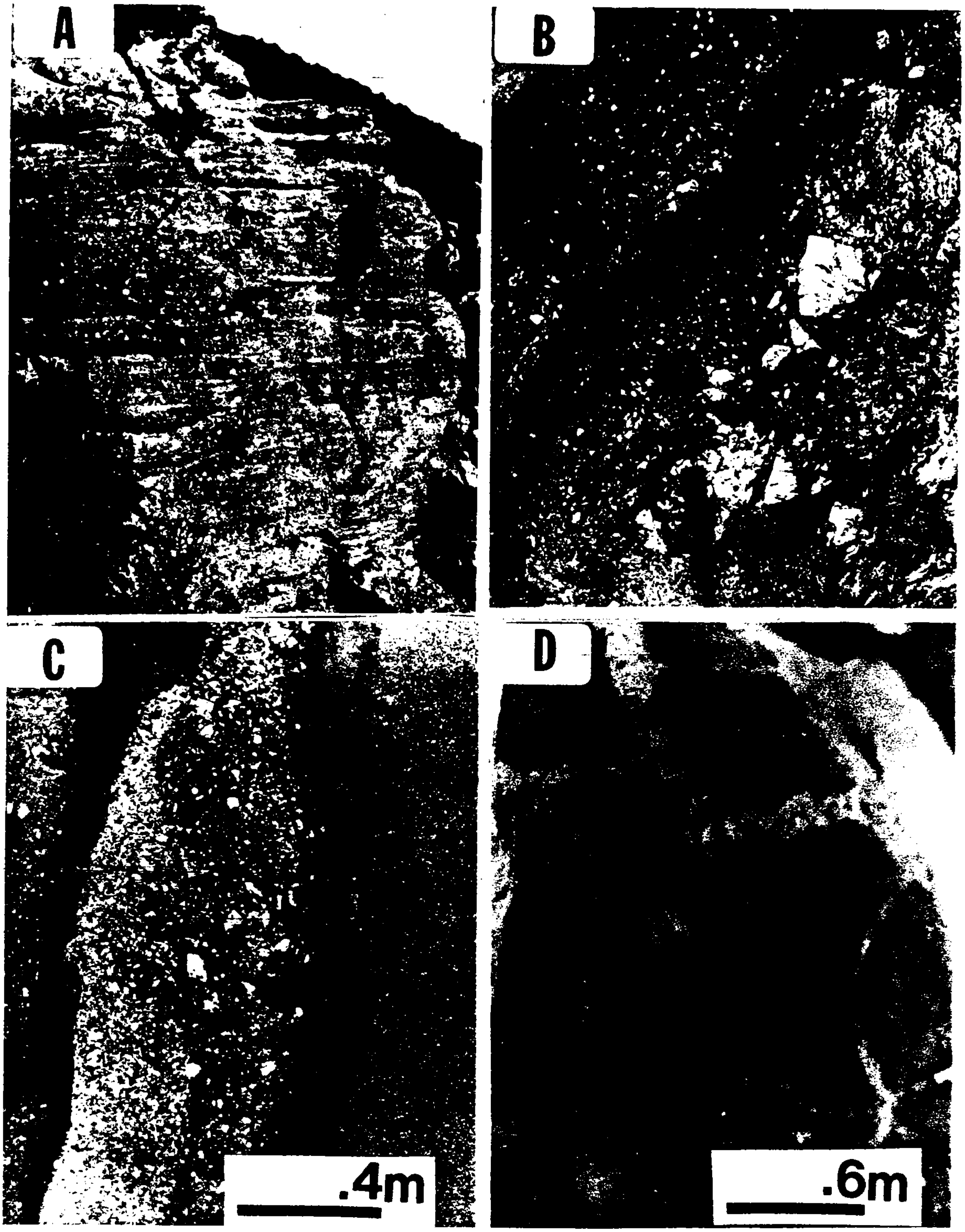

Figure 6. Photographs of Violin Breccia. A, massive to poorly bedded outcrop of Violin Breccia; B, angular clasts in a diamictite; C, alternating gravel, sandstone, and mudstone beds which are locally burrowed; and D, mudcracks on sole of bed.

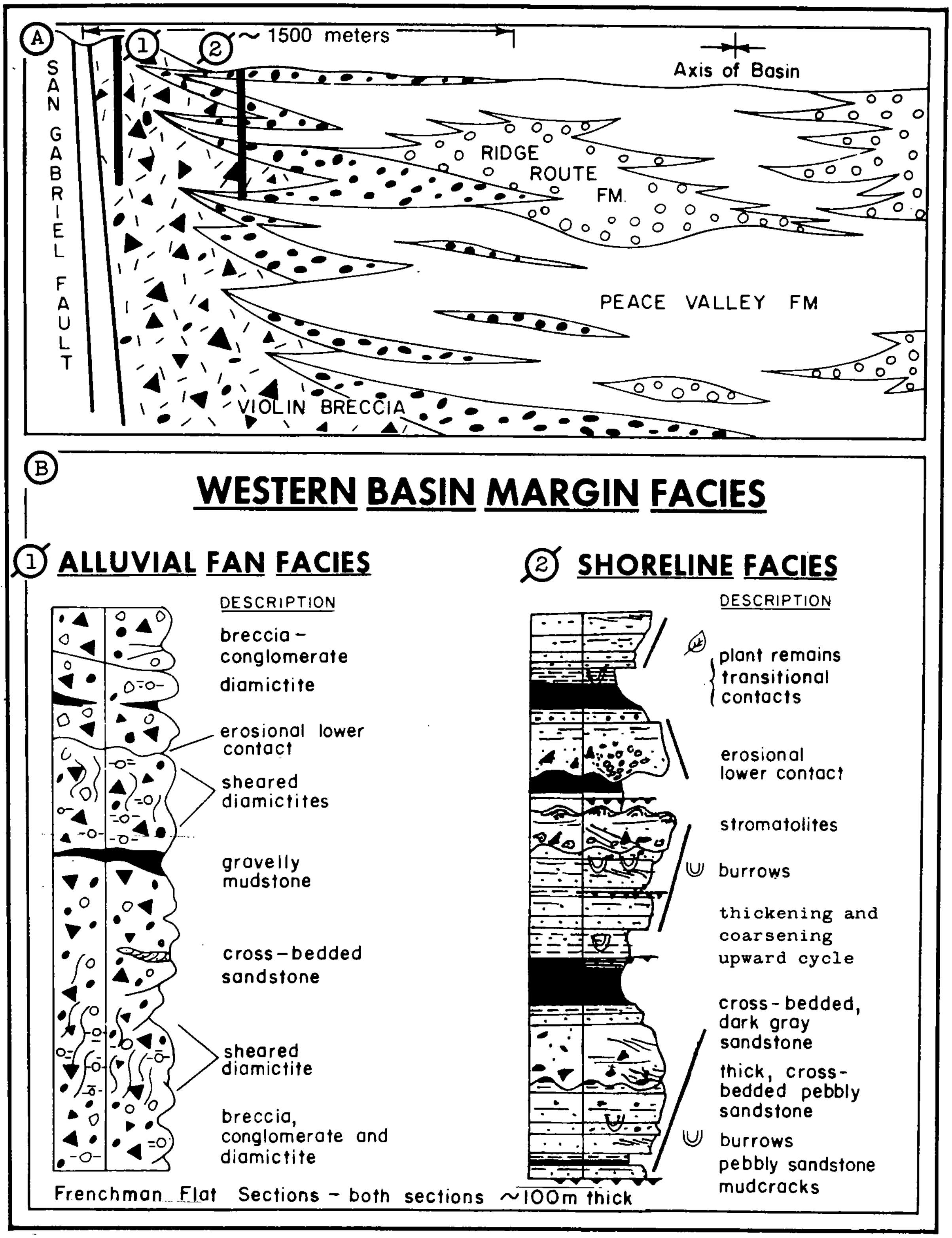

Figure 7. Diagram and schematic section of alluvial fan and shoreline facies of Violin Breccia. A, Stylized cross-section of Violin Breccia stratigraphic relationships and B, western basin margin facies showing (1) alluvial fan facies, and (2) shoreline facies.

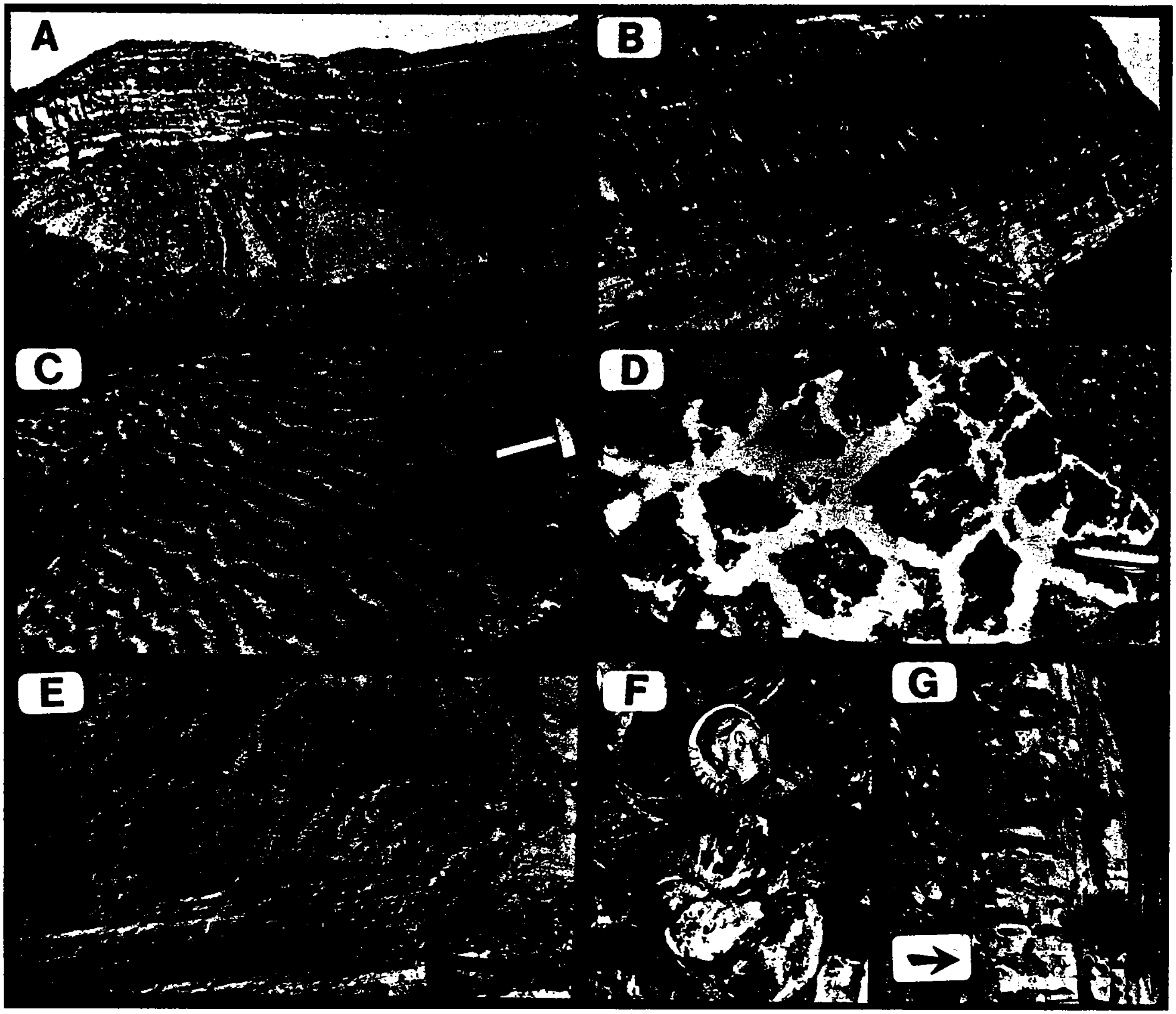

Figure 8. Photographs of western margin and axial shoreline and fluvial-deltaic facies. A, interfingering axial sandstone and mudstone sequence (Piru Gorge and Cereza Peak members); B, thinning- and fining-upward channel and interchannel sequence (Piru Gorge member); C, ripple marks; D, mudcracks; E, low-angle (accretionary) cross bedding in sandstone; F, *Anodonta* sp. (bivalve) and *?Stagnicola* (gastropod); and G, convex-upward sandstone interpreted as a bar complex (arrow points to the top of the bed).

Turbidite facies

This facies is characterized by thin- to thick-bedded sandstone interbedded with organic-rich mudstone and shale (Figs. 10,11). Strata are commonly graded; contain mudstone rip-up clasts, dish structures (Nilsen et al., 1977), sole marks, and slump-folded strata (Fig. 10). Individual turbidite sandstone units are intercalated with the offshore facies and are graded and amalgamated, contain Bouma Ta, Tab, Tbc, and Tcde intervals, sole marks, mudstone rip-up clasts, and may be bioturbated or rippled marked (Fig. 10). Thick-bedded turbidite (>1 m) sandstone units are best exposed in the Castaic Formation and Marple Canyon Sandstone Member of the Ridge Route Formation (Fig. 11B). This package contains the marine-nonmarine boundary in Ridge Basin and probably represents prodelta fan deposits accumulating in marine (<100 m water depth) or paralic environments (Fig. 2A). This sandstone package is over 1300 m (4250 ft) thick and is thickest in the central part of the basin where it interfingers and changes facies to coarser-grained shoreline facies along the flanks of the basin. In this lens-shaped unit paleocurrent directions suggest sediment transport to the southwest across the basin axis and southeast along the axis of the basin. The turbidite beds are interbedded with massive to laminated mudstone which may be slump folded (Figs. 10A,C), pulled apart (Fig. 10B), and locally inclined. Molluscs, pebbly mudstone, and small-scale cross stratification occur in the mudstone sequences. The turbidite sandstone consists of facies B, C, and D, and locally of facies A and F of Mutti and Ricci Lucchi (1978). Several thickening- and coarsening-upward cycles occur in the lower part of Paradise

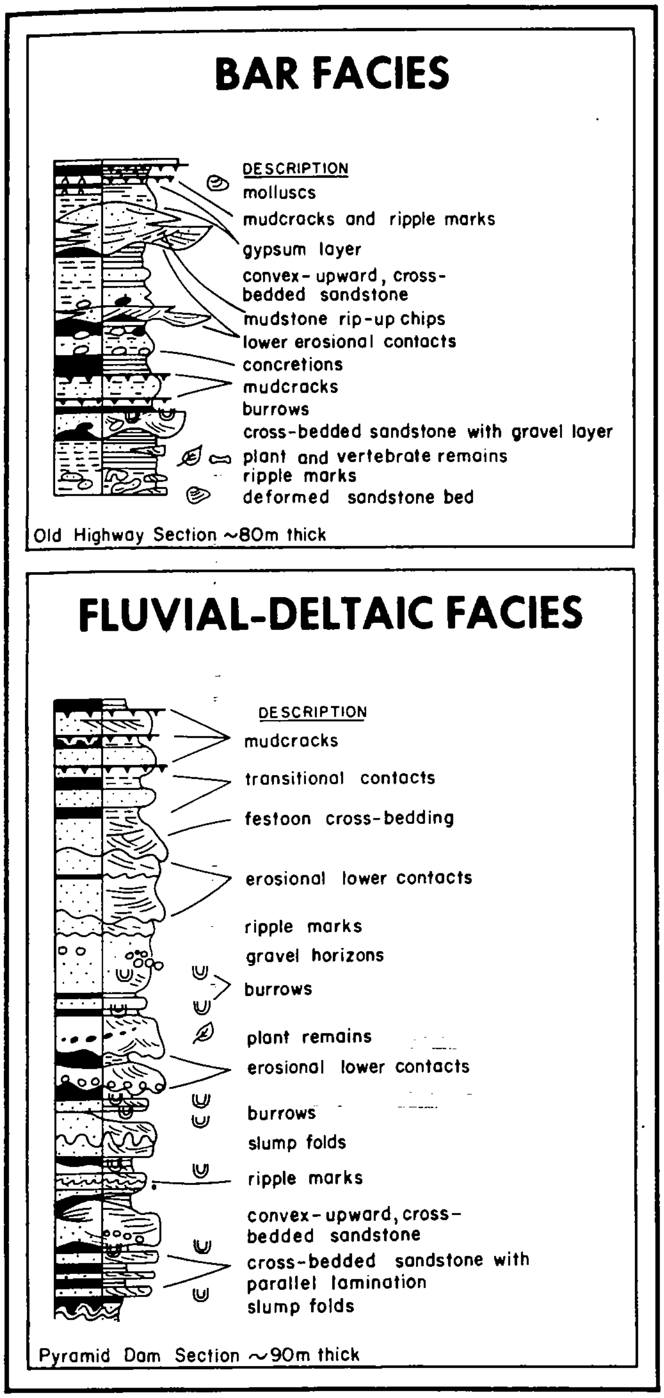

Figure 9. Schematic sections of bar and fluvial-deltaic facies. A, bar facies and description; and B, fluvial-deltaic facies and description.

Ranch Shale and Marple Canyon Sandstone Members and in the Castaic Formation. These sequences are interpreted to be depositional lobes consisting of facies C and D (Fig. 11B). Thinning- and fining-upward sequences are the most common cycles observed in the Castaic Formation and Marple Canyon Sandstone Member. They consist of facies B and locally facies A turbidites and are interpreted as channel deposits (Fig. 11B). These channel deposits are laterally and vertically transitional into interchannel deposits consisting of mudstone and facies B, C, D, and F. These interchannel deposits are highly slump folded and locally inclined away from the channels. No vertical cycles were observed in the interchannel deposits.

Thin-bedded (<1 m) turbidites are best exposed in the various members of the Peace Valley Formation (Fig. 11A). They form laterally persistent beds, which are graded, contain Bouma Ta, Tab, Tabc, Tbcde, Tabcde sequences, dish structures, sole marks, rip-up clasts, and flame structures. They consist of facies C, D, and G turbidite associations. These thin-bedded turbidites are commonly reworked by traction currents, bioturbated, and have rare animal tracks and mudcracks on the top of the bed. Laminated, varve-like graded beds in predominantly mudrocks suggest turbidite deposition occurred as interflows or surface flows of silt in Lake Mead-like currents (Gould, 1951). Other thin-bedded turbidites were derived from fluvial-deltaic sequences and represent density underflow deposits. The fluctuating lake level may account for animal tracks, mudcracks, and traction current features superimposed on the turbidites. Many of these turbidites are in the axial part of the basin and generally do not form any distinctive vertical sequences. They are similar to the basin-plain turbidites of Mutti and Ricci Lucchi (1978).

Offshore facies

This facies is characterized by massive to horizontally-laminated mudstone, shale, claystone, clayshale, dolomicrite, and siltstone, which contain ferroan dolomite, other iron carbonates, analcime, pyrite, jarosite, and gypsum (Figs. 12A,B; 13). The facies occurs in the axis of basin and includes the Castaic Formation and the various members of the Peace Valley Formation. Sedimentary structures include parallel, varve-like laminations (Fig. 12B), concretions (Fig. 12C), soft sediment deformation (Fig. 12E), small-scale internal faulting (Fig. 12E), injection features (Fig. 12D) and locally graded and brecciated beds. Biological constituents include plant remains, fish, insects, peloids, shell debris, and minor burrowing (Fig. 12F). Total organic carbon (as C percent) ranges from 0.16 to 2.78 and bulk ferrous iron content ranges from 1.5 to 9.5 percent FeO. The analcime and ferroan dolomite both appear to be early diagnetic with the analcime formed from the reaction of smectite and kaolinite with saline, alkaline waters and the ferroan dolomite formed by diagenesis of a high-Mg calcite precursor phase in a reducing environment (Irvine, 1977; Smith, 1981). The association of black, highly organic mudstone and shale, various carbonates, analcime, pyrite, and paucity of in situ fossils and burrowing may suggest reducing conditions at least within the sediment. These reducing conditions may have been enhanced by chemical and/or thermal stratification within the water column. Burrows also indicate oxic conditions along the sediment-water interface. Periodic subaerial exposure of these sediments is suggested by desiccation features, brecciated beds and animal tracks.

Smith (1981; this volume) recognized three major subdivisions of the mudrocks of this facies: (1) deep-water, (2) shallow-water, and (3) deep-water brackish phases. It appears that these facies or phases can be applied to both the marine and lacustrine deposits in Ridge Basin (Fig. 14). The deep-water facies consist of clayshale, shale, and mudstone interbedded with turbidites. Mudcracks, bioturbation, wave ripples, vertebrate tracks, analcime, dolomite, gypsum nodules, and varve-like

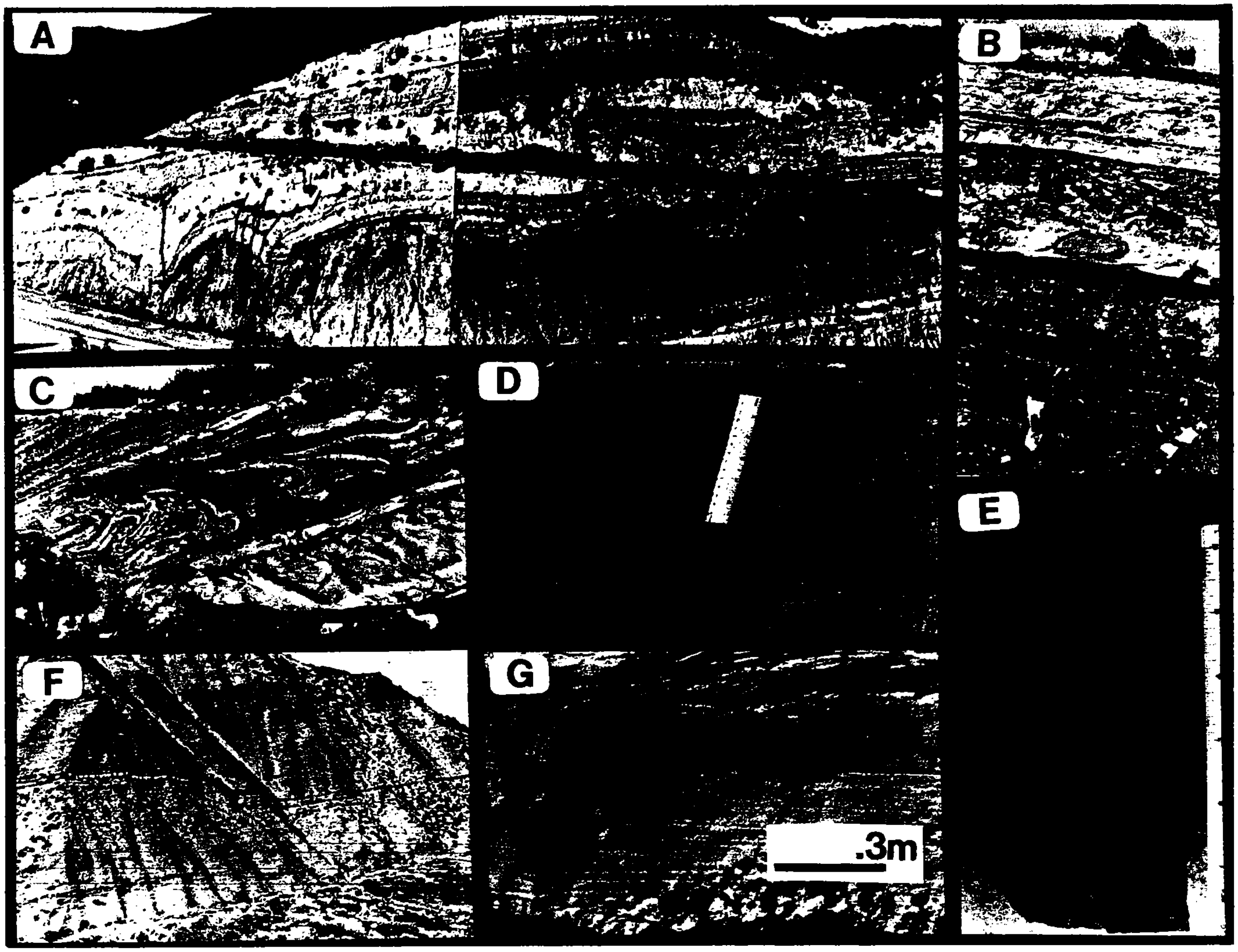

Figure 10. Photographs of turbidite facies. A, large-scale slump-folded strata showing growth faults with laterally continuous overlying and underlying strata; B, brecciated bed bounded by laterally continuous strata; C, small-scale slump-folded and disrupted strata; D, dish structures; E, groove casts on sole of bed; F, thin-bedded turbidites in shale sequence (Osito Canyon member); and G, graded sandstone bed showing Bouma Ta, b, c intervals and a erosional base.

deposits are absent (Smith, 1981). Estimated water depth is 25 to 100 m with anoxic bottom conditions. The marine Castaic Formation and the lacustrine Paradise Ranch Shale Member are in this group (Fig. 11B). The shallow-water facies consists of claystone and mudstone interbedded with turbidites and fluvial-deltaic deposits. Mudcracks, bioturbation, wave ripples, and vertebrate tracks are common (Fig. 11A). Analcime, dolomite, and gypsum nodules are rare; varve-like deposits are absent. Estimated water depth is 4 m with oxic bottom conditions. The Osito Canyon and Cereza Peak Shale Members would be in this facies (Smith, 1981). Deep-water brackish facies consist of dolomicrite, mudstone, shale, and siltstone interbedded with turbidites. Mudcracks, bioturbation, wave ripples are rare to absent. Vertebrate tracks do not occur. Analcime, dolomite, and varve-like deposits are common. Estimated water depth is 25 m with alternating anoxic and oxic conditions. The Posey Canyon Shale, Alamos Canyon Siltstone, and the middle Marple Canyon Sandstone members are in this group.

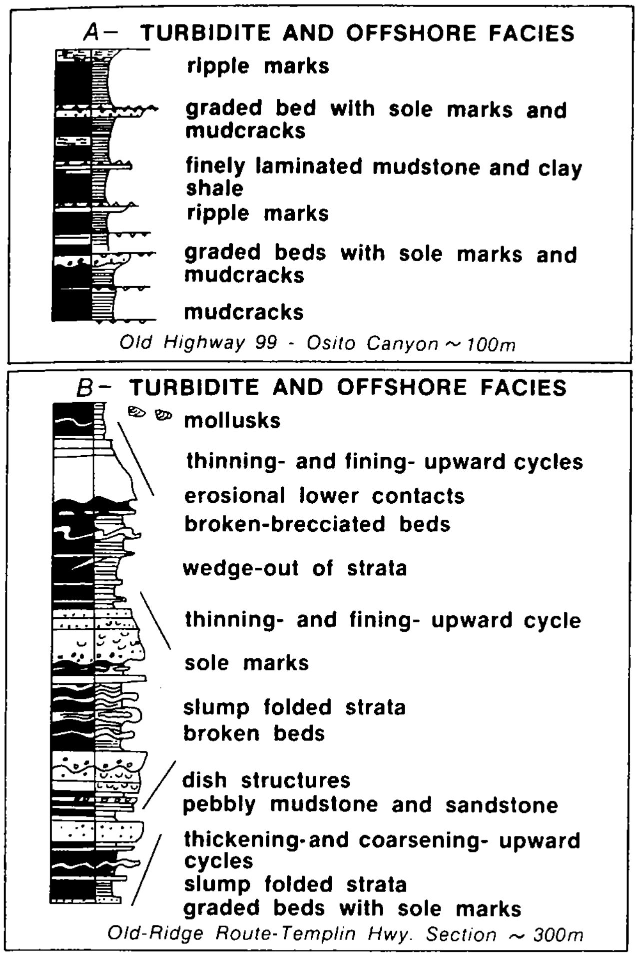

Figure 11. Schematic sections of turbidite and offshore facies. A, thin-bedded turbidite and offshore facies in shallow-water environments; and B, thick-bedded turbidites and offshore facies in deeper-water environments.

Fluvial-Deltaic Facies

Five major fluvial-deltaic deposits occur in the axial part of Ridge Basin. The upper part of the Marple Canyon, Fisher Spring, Frenchman Flat, Piru Gorge, and Apple Canyon Sandstone members all thicken in the axial part of the basin and thin and interfinger with the Violin Breccia to the west and the Ridge Route Formation to the east. These members form lobe-shaped deposits that generally thicken- and coarsen-upward (Fig. 9B). They may consist of a (1) lower turbidite sequence, overlain by, (2) multistacked channel-levee cycles, and (3) interchannel deposits. These sequences are arranged into several megasequences, each is up to 60 m thick. The lower turbidite interval is locally up to 10 m thick and grades laterally and vertically into the offshore facies. The channel-levee sequence is up to 30 m thick and consists of cross-bedded channel sandstone, slump-folded strata, and inclined levee deposits arranged in thinning- and fining-upward cycles. It makes up the middle and upper parts of each megasequence. Inter-channel deposits are interbedded with and laterally transitional with the channel-levee sequences. They consist of thick intervals of organic rich, bioturbated, mudcracked mudstone and sandstone which locally have rootlets and animal tracks. All the fluvial-deltaic sequences are interpreted to be freshwater, oxic, near surface deposits (Fig. 14).

Sedimentation Rates

Sedimentation rates in Ridge Basin are extremely high inasmuch as about 12 km of sedimentary rocks accumulated in it in 6 m. y. It therefore has one of its highest known rates yet documented. The precise age and duration of these rocks is inferred from magnetostratigraphic reversal correlations (Ensley, 1980; Ensley and Verosub, 1982), mollusc and benthic foraminifera stages, mammal stages, and projected sediment accumulation rates. No isotopic dates for the sedimentary sequence have come out of Ridge Basin as yet. The sedimentation rate in Ridge Basin is interpreted to be about 2 m/1000 years. Ensley (1980) and Ensley and Verosub, this volume) found sedimentation rates varied in the basin from 0.2 m/1000 years to 3.2 m/1000 years, averaging 1.8 m/1000 years. Figure 14 shows the inferred sedimentation rates for the various units in Ridge Basin. It is based on the ages established by Ensley (1980), strata thicknesses, and calculated rates of sediment accumulation. It appears that the shale members accumulated at a faster rate (3.0 m/1000 years) than the sandstone members (0.2 m/1000 years) (Ensley and Verosub, 1982).

CONCLUSIONS

Ridge Basin formed as a narrow, small basin in a tectonically active region and in which a great quantity of coarse-grained terrigenous sediment accumulated in alluvial fan-fluvial complexes which flanked the basin. Fluvial-deltaic and shoreline environments were transitional between its margins and its center. These environments were subject to major terrigenous influx, subaerial exposure and wave agitation. In the basin axis, thick accumulations of organic-rich mudstone turbidites, and fluvial-deltaic deposits formed.

The physical and chemical nature of Ridge Basin changed considerably through its 6 million year history. Ridge Basin was born in Castaic Formation time (10-12 m. y. ago) and started its history as a relative shallow-marine embayment. Local shorelines formed on the northeast side of basin and thick deposits of turbidites and offshore mudstone accumulated in its center. With continued right slip along the San Gabriel Fault, the marine basin was restricted and eventually shoaled, becoming nonmarine. Lacustrine, fluvial-deltaic, and alluvial-fan phases of Ridge Basin followed. A sequence of shallow-lakes of variable water depth and chemistry suggest generally external conditions of drainage to the southeast into the Ventura Basin. Thick chemical deposits in the upper lake deposits suggest either local closed basin, internal drainage or chemical stratification in an externally drained system. The final event in the history of Ridge Basin was the termination of strike-slip displacement on the San Gabriel Fault coupled with infilling of the northern end of basin with fluvial and alluvial deposits of the Hungry Valley Formation, ending about 4 m. y. ago.

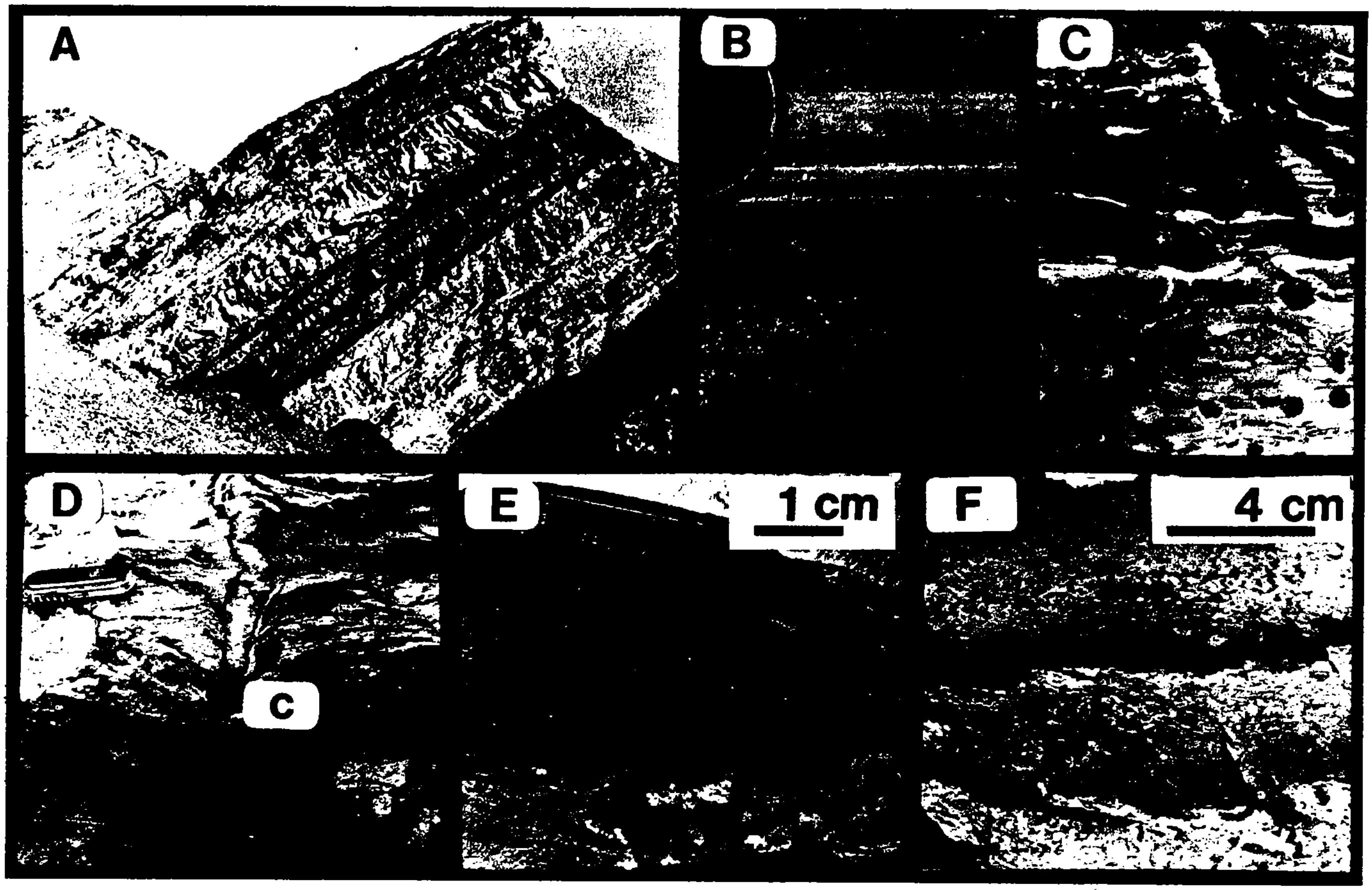

Figure 12. Photographs of offshore facies. A, interbedded resistant dolomicrite and softer mudstone; B, varve-like laminae of dolomicrite, analcime, and organic-rich shale; C, gypsum nodules in laminated siltstone and shale; D, injection concretionary beds (C); E, small-scale sluried and slump-folded strata; and E, mottled burrowed and internally faulted mudstone and shale.

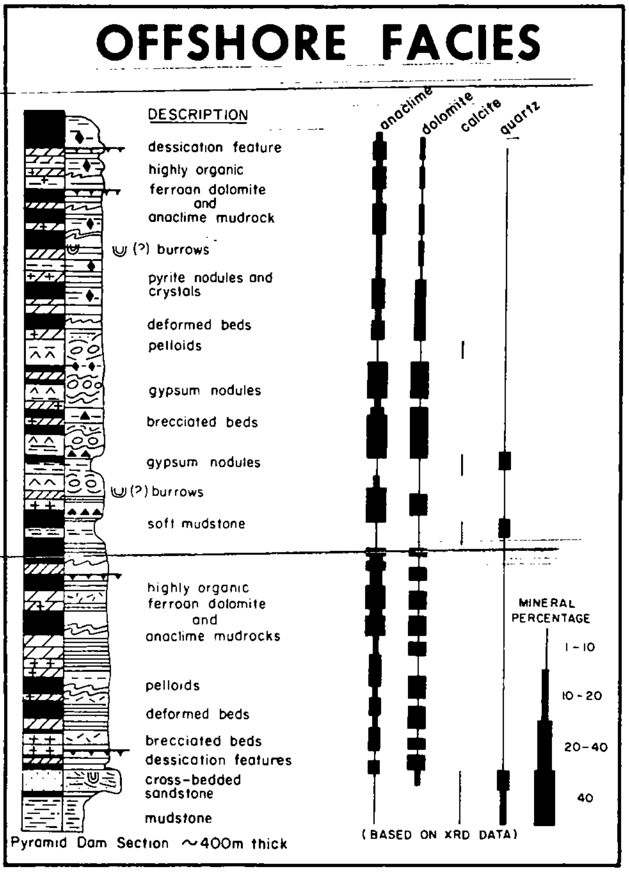

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Figure 13. Schematic section of offshore facies. Modified after Irvine (1977).

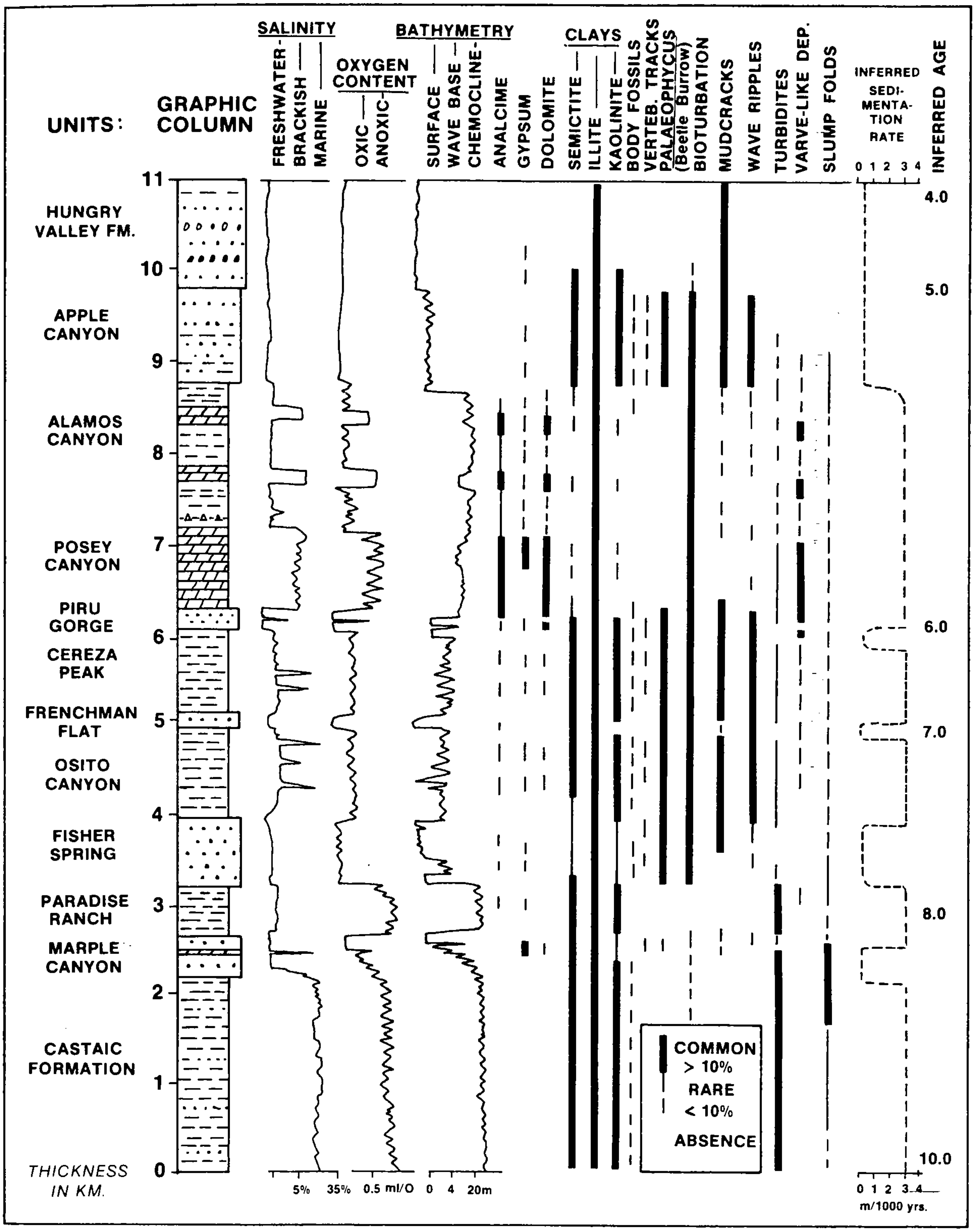
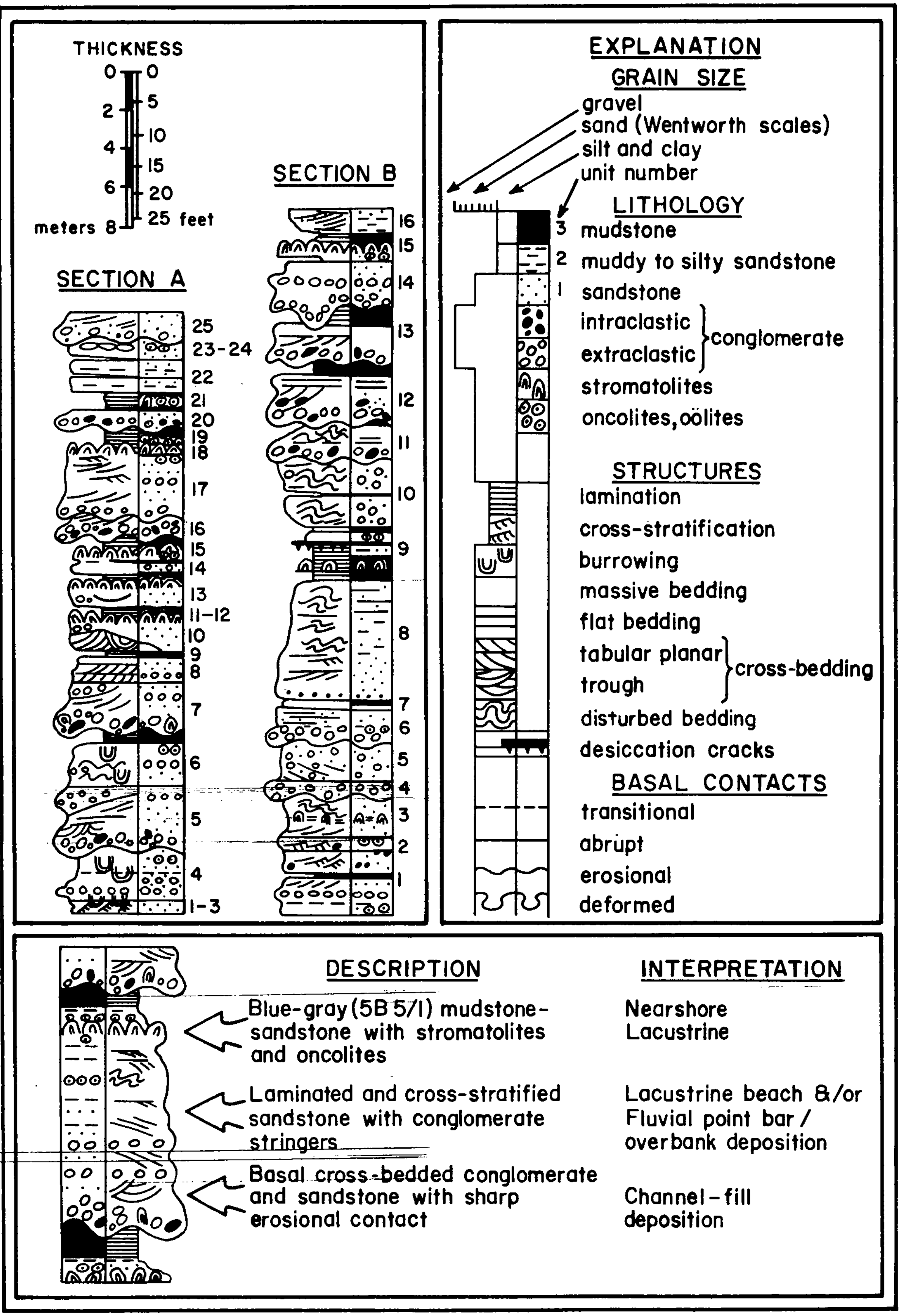


Figure 14. Paleobathymetry, paleosalinity, inferred sedimentation rates and oxygen content diagram. Modified after Smith (1981); see Smith (this volume) for discussion of the criteria for salinity, oxygen, and bathymetry determination; and Ensley and Versoub (this volume) for sediment accumulation rates.



DESCRIPTION	INTERPRETATION
Blue-gray (5B 5/1) mudstone-sandstone with stromatolites and oncolites	Nearshore Lacustrine
Laminated and cross-stratified sandstone with conglomerate stringers	Lacustrine beach &/or Fluvial point bar / overbank deposition
Basal cross-bedded conglomerate and sandstone with sharp erosional contact	Channel-fill deposition

Stratigraphic section along the Old Ridge Route showing interfingering fluvial, shoreline, and near shore facies.

THE VIOLIN BRECCIA, RIDGE BASIN, SOUTHERN CALIFORNIA

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ABSTRACT

A narrow but continuous belt of Upper Miocene sedimentary breccia, the Violin Breccia, crops out along the southwestern border of Ridge Basin adjacent to the San Gabriel Fault. The formation accumulated as talus and in steep alluvial cones and fans that changed facies within a kilometer or so from a continuously rejuvenated fault scarp toward finer-grained sediments along the trough of the basin. All of the material constituting the breccia, exposed along the fault for about 40 km (25 mi), was derived from a limited source area, identified on the northwest. Over 11,000 m (36,000 ft) of the formation was deposition as the Ridge Basin depocenter moved by this source area. This source area was raised to erosion within a constraining bend of the ancient San Andreas-San Gabriel transform fault. The depositional system resulted in a shingled arrangement of stratal units within the basin so that younger units successively overlie older toward the northwest. The vertical stratigraphic thickness at any one place, however, is only about one-third of the total stratigraphic thickness measured in outcrop.

INTRODUCTION

The Violin Breccia is a sedimentary formation consisting of a rubble of blocks, boulders, and cobbles embedded in an earthy and sandy matrix, at most places massive but becoming bedded toward the northeast. It is exposed in a continuous band along the southwestern flank of Ridge Basin for 34 km (21 mi) adjacent to the San Gabriel Fault. Along its full extent it changes facies abruptly into shale and sandstone within 1,500 m (less than 5,000 ft) where tongues of the sedimentary breccia reach toward the trough of Ridge Basin. Stratigraphically the formation is over 11,000 m (36,000 ft) thick, including an estimate of the thickness in the subsurface to the southeast. An unknown additional thickness probably lies along the buried San Gabriel Fault to the northwest. The lowermost units of the formation are Upper Miocene in age (10 and 12 m. y. ago) and marine and the upper part nonmarine. The youngest beds are Pliocene in age, and were deposited about 3 m. y. ago. It therefore took between 7 and 9 m. y. to deposit the 11,000 m, or between 1,200 and 1,500 m per m. y. Its lithology and field occurrence implies continuous or intermittent activity on the San Gabriel Fault in order to rejuvenate its fault scarp.

The formation was named and a type section designated for exposures along the divide between Violin and Palomas Canyons northwest of Castaic (Crowell, 1954b). In this general region it is especially well exposed in Palomas Canyon, and may petroleum geologists had used the name Palomas Conglomerate for the unit previously. Unfortunately, however, Palomas was preoccupied as a stratigraphic name and so Violin was selected. Violin Canyon, according to local legend, was named when topographers found a broken violin within alluvial debris at the mouth of the canyon after a flood. The unit had been briefly referred to in the literature previously by Clements (1937, p. 213) and by Eaton (1939), who was first to recognize that the nature of the sedimentary breccia required continuous activity on the fault during the time of its deposition. The uppermost part of the Violin Breccia was described by Crowell (1950) as the Conglomerate - Breccia Member and Brown Conglomerate Member of the Hungry Valley Formation into which both graded laterally. As the result of additional investigations within Ridge Basin following the work in the Hungry Valley region, however, the continuity of these members with the belt of breccia to the southeast indicated that these members more properly belonged to a new formation, subsequently named the Violin Breccia (Crowell, 1954b) and included within the Ridge Basin Group. Before this redefinition was published, however, the significance of the unit in indicating right slip of up to 40 km (25 mi) was recognized and the Violin Breccia was referred to as the gneiss-bearing sedimentary breccia (Crowell, 1952a).

DESCRIPTION

Next to the San Gabriel Fault the Violin Breccia consists of massive rubble without bedding in which large blocks and boulders up to 2 m in diameter are embedded in an earthy or sandy sparse matrix (Figs. 1, 2) that make craggy outcrops (Fig. 3). Within a few tens of meters from the fault, however, intercalations of sandy mudstone, diamictite, and obscure sorting of the breccia outline faint bedding (Fig. 4). (See also, Link and Osborne, this volume, Fig. 6.) In moving away from the fault still farther, bedding becomes increasingly distinct and is characterized by irregular units up to as much as a meter in thickness with irregular and gradational contacts (Fig. 5). The coarse beds here are mainly diamictites and conglomerates showing a lower part with reversed grading, a central part consisting of a mosaic of angular and subangular stones, and a normally graded upper part. Both bottom and top contacts grade into thin but massive units of greenish and brown sandy mudstone. Only rarely are such mudstones intercalated with laminated and cross-bedded sand lenses and irregular channels with erosional bottoms. Midway along its outcrop belt a dioritic and anorthositic mass 300 by 120 m in outcrop area is interbedded within the breccia and abuts against the San Gabriel Fault on the southwest (Shepard, 1962). It is interpreted as a landslide mass squeezed out of the fault zone during the Miocene, or a mass that slid across the fault from high terrane on the southwestern side. Near the fault the Violin Breccia is broken by many joints, shears, and small faults.

The distal parts of the Violin Breccia are moderately well bedded but lenticular. Brown

in Crowell, J.C., and Link, M.H., eds., 1982, **Geologic History of Ridge Basin, Southern California,** Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 89-98.

Figure 1. Violin Breccia, Palomas Canyon, 2.5 km northwest of Castaic. Outcrop is about 400 m from the San Gabriel Fault. Note large block of augen gneiss embedded in rubbly matrix. Material here probably deposited in a marine environment inasmuch as it is laterally equivalent to beds of the marine Castaic Formation only 400 m along bedding to the northeast.

sandstone and mudstone lenses are intercalated within more persistent conglomerate-breccia beds. The lateral transition into shale beds characteristic of the trough of Ridge Basin is abrupt. At places, such as within the gorge of Piru Creek below Frenchman Flat, fine material has been removed from between clasts in conglomerates, presumably by current winnowing, so that stones rest against each other to make a clast supported mosaic. Some clasts in this vicinity are coated with an algal layer as much as 5 mm thick indicating that they lay near the shoreline of one of the lakes that intermittently occupied Ridge Basin. Elsewhere, stromatolitic beds up to 10 cm thick are intercalated in the section and record shoal water at the lake margin. Paleocurrent indicators, although uncommon with the Violin Breccia, show flow to the east and northeast (Link, this volume). Distal tongues of the conglomerate-breccia at some places display imbrication showing northeastward flow, and a few sandstone layers preserve both tabular and trough cross-bedding, suggesting the same direction of transport. Channels cut into the substrate and now filled with coarse debris and lobate termini of debris flows, observed in cross section, also suggest northeastward flow.

Upward-thickening and upward-coarsening sequences of conglomerate beds are conspicuous along the middle reaches of the Violin Breccia. The origin of these sequences, now under study by Dr. R. J. Steel, University of Bergen, Norway, may have resulted from sharp tectonic rejuvenation of the fault scarp followed by a flood of coarse debris that was carried down steep fans and prograded toward the axis of Ridge Basin. The repetitious and cyclic aspect of these progradations may reveal information on the timings and intensities of the tectonic activities responsible. As pointed out by Hollywood and Osborne (this volume) prograding tongues of Violin Breccia reaching into the thick shale units of Ridge Basin show that mud was deposited primarily within the basin trough when the San Gabriel Fault was most active.

These characteristics of the Violin Breccia lead to the interpretation that the material accumulated as a talus deposit along a steep escarpment. The talus quickly gave way downslope to bedded alluvium which in turn led onto gentler slopes. The older parts of the breccia extended to slopes that reached directly into the sea. Middle and younger parts extended onto alluvial flats and at times into lakes that occupied Ridge Basin intermittently. Where tongues of sandstone, mudstone, and conglomerate extend almost to the San Gabriel Fault in contrast to the usual massive rubble, it is inferred that a steep stream carrying alluvium crossed the escarpment at or near that place. In view of the presence of the San Gabriel Fault, it is also inferred that the steep escarpment was the fault scarp caused by activity on that fault zone. At a few places small outcrops of Violin Breccia lie unconformably upon basment rocks within the fault zone. These are interpreted as places where talus transgressed across the fault scarp as it retreated between times of active rejuvenation.

Clasts contained within the Violin Breccia are primarily gneissic and granitic in composition. Twenty counts are shown by means of pie diagrams in Figure 6 obtained from localities distributed along the band of outcrop. Many of the large stones provide samples not only of the rock type whence the stone came but display minor structures and textural details that aid significantly in characterizing their source. Sedimentary stones are very rare or absent. Moreover, there is little significant variation in the mix of rock types from bottom to top. These data indicate that the material was derived from a nearby basement source underlain by gneiss and granitic rocks. Violet quartz gneiss, indicative of granulite facies, and some gabbro and anorthosite and related rocks, help to characterize the source, as do the abundance of amphibolities, augen gneisses of different sorts, and other distinctive rock types (See also, petrography, Link, this volume). As discussed more fully below, all of the material must have been derived from the Alamo Mountain - Frazier Moutain region now bordering Ridge Basin on the northwest. Streams notching the San Gabriel fault scarp at places along its reach evidently were very steep and short inasmuch as they failed to extend headward into terrane underlain by pre-Miocene sedimentary strata that are still widely preserved to the west.

Figure 2. Violin Breccia, Piru Creek, 1 km downstream from confluence of Snowy Creek. Proximal sedimentary breccia consisting primarily of angular blocks of gneiss within a rubbly matrix.

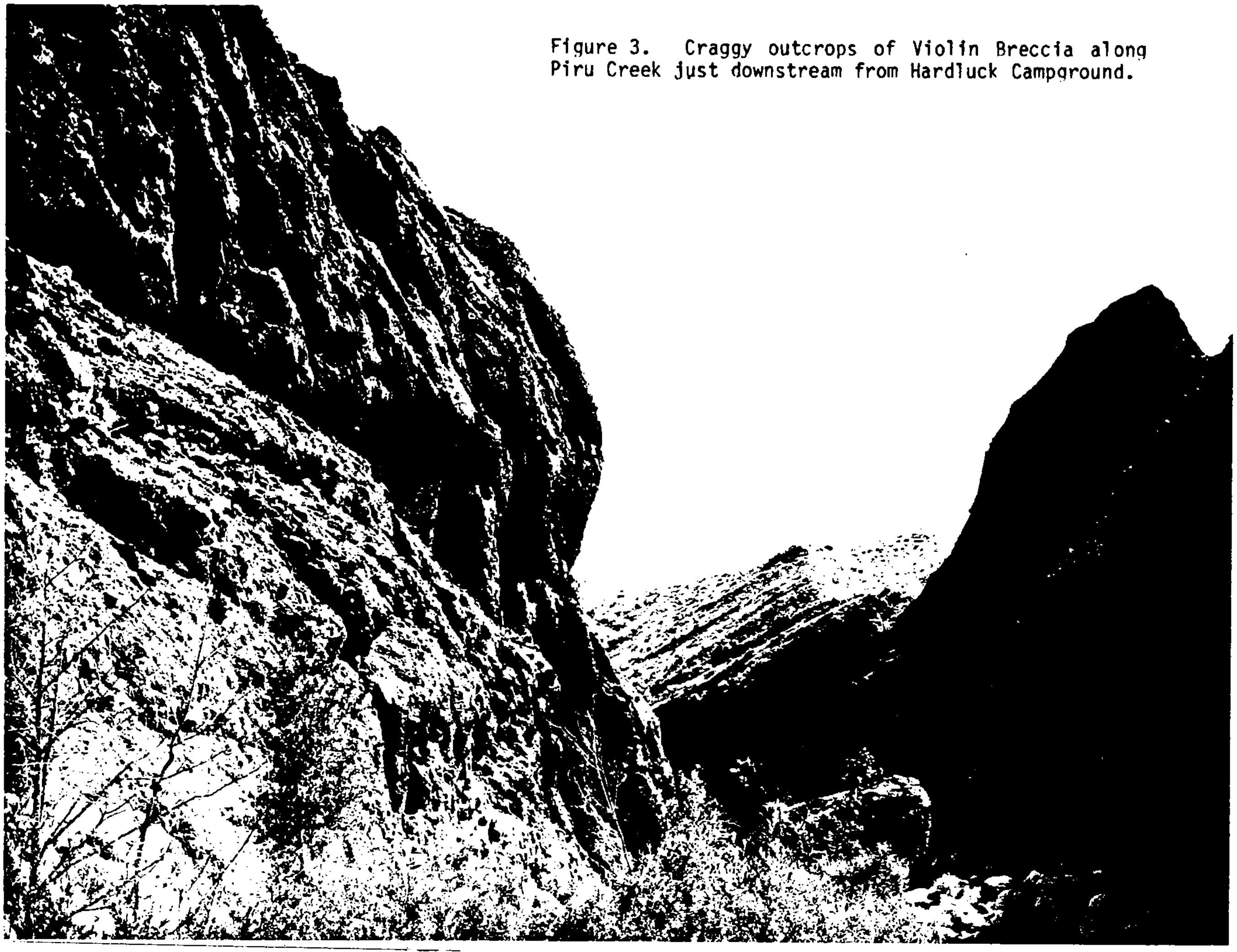


Figure 3. Craggy outcrops of Violin Breccia along Piru Creek just downstream from Hardluck Campground.

THE DEPOSITIONAL SYSTEM

The great thickness, distinctive provenance, and rapid facies change of the Violin Breccia laterally into finer sediments of Ridge Basin suggest that the formation was deposited as the result of an unusual depositional system. As explained in the tectonics discussion (Crowell, this volume) 13,400 m (44,000 ft) of strata were laid down within the trough of Ridge Basin by means of a converyor-belt mechanism within a strike-slip regime. The depocenter migrated relatively northwesterly through time beginning about 10 or 12 m. y. ago, and ending early in the Pliocene, about 4 m. y. ago. As it did so, strata were laid down so that younger units overlapped older toward the northwest. Much of the evidence for this manner of deposition within the splintered transform belt comes from a consideration of the system involved in the depositional of the Violin Breccia.

As described above, all of the material within the Violin Breccia was derived from a limited source area underlain by distinctive gneissic, granitic, and dioritic rocks in the vicinity of Frazier Mountain and Alamo Mountain, now situated at the northwestern margin of Ridge Basin. Figure 7 is a diagram in which this limited source region is reconstructed. Inasmuch as no sedimentary clasts have been recognized within the formation that could have come from the wide region to the west underlain by pre-upper Miocene beds, primarily Eocene, the source area for the sedimentary breccia did not include this western region. Streams extending from the San Gabriel fault scarp apparently did not reach westward to tap this region; instead, they were short and steep and drained the area shown within the hachures only. In addition, the central Transverse Ranges have been uplifted in latest Cenozoic time and as a consequence have been deeply eroded. It therefore seems reasonable to draw the figure as shown, with a more limited source region during the Miocene than the region with appropriate source rocks at the surface today. The limits of the source area on the northwest (Fig. 7, C) is questionable because displacements on the Big Pine Fault and related faults have not yet been documented as these displacements might influence the shape of the source region during the Miocene.

The oldest units of the Violin Breccia are found at the southeast in the vicinity of Castaic, both at the surface and in the subsurface (Fig. 7, A) (Stitt, this volume). Here they are marine, and intertongue with shale of the Castaic Formation toward the northeast. Across the San Gabriel fault zone toward the southwest, however, they are faulted against Modelo conglomerate, also marine, so that two coarse units lie "back to back" in a sedimentological missmatch (Crowell, this volume, Fig. 5; Crowell, et al., 1982, cross section Z-Z'). Tongues of the Violin Breccia here were derived as well from the source area in the Frazier - Alamo Mountain region but now displaced at least 40 km (25 mi) by right slip. In addition, the continuity of the sedimentary breccia between the source region and the many tongues in between A and B (Fig. 7) require continuous displacement on the San Gabriel Fault. This displacement involved both right slip and low-angle oblique slip. The latter was related to depression of the depocenter in Ridge Basin and the concomittant uplift of the source region.

The depositional system is somewhat analgous to the loading of a coal train, car by car, as the train moves beneath a hopper (Fig. 8). In this illustration a tectonic pulse raises the source area - the hopper - and depresses the depocenter - the coal car. This tectonic pulse is accompanied by lateral shifting, pictured as corresponding to the rattle and bang as the train engineer moves an empty car beneath the hopper. Figure 9 shows the analogue in map view. In this diagram, stratal units of the Violin Breccia are portrayed as semi-circular fans extending into Ridge Basin from the San Gabriel fault scarp. They lie as shingled units, like scales on a fish or shingles on a roof, with successively younger ones lying to the northwest. The great stratigraphic thickness of the Violin Breccia is measured by adding up the thickneses measured in outcrop where bedding can be ascertained. As explained in the tectonics discussion (Crowell, this volume) the reproducible stratigraphic so measured is not the same as that that would be penetrated by a well drilled orthogonally to the bedding. This vertical thickness is deemed to be only about one-third as much.

CONCLUSIONS

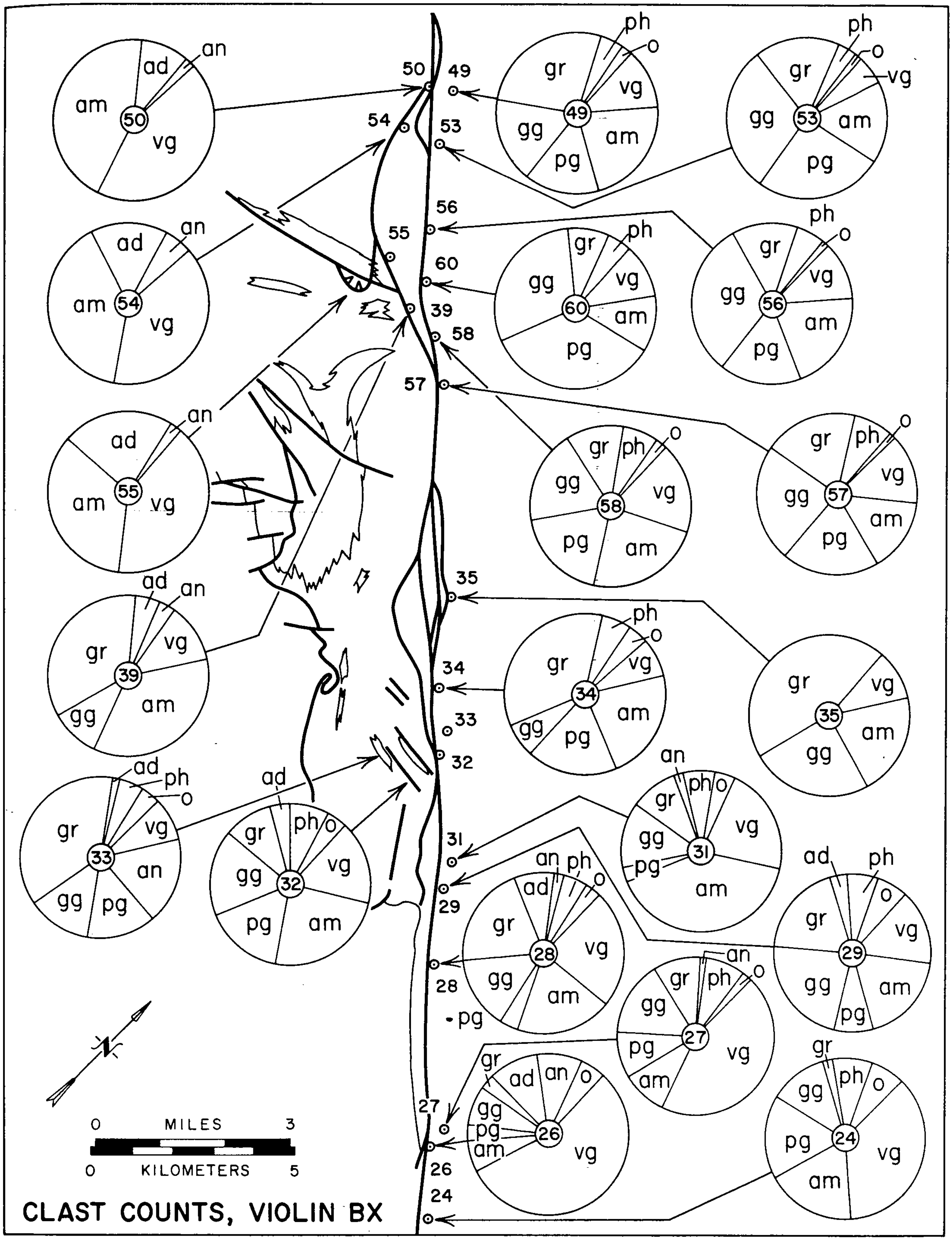
A depositional system involving a sigmoidal bend in the San Gabriel Fault and major right slip accounts satisfactorily for the rubbly lithology and great thickness of the Violin Breccia (Crowell, 1974a). The source area was squeezed and raised within the restraining bend of this major fault strand, part of the San Andreas transform system between the Pacific and North American lithospheric plates during late Miocene time. The fault scarp separating the source area from the depositional site

Figure 4. Bedded Violin Breccia with northern Ridge Basin and Liebre Mountain in the background, and old U. S. Highway 99 in valley. Adjacent to northern spur of Whitaker Peak Fire Road, Roger Hope in photograph.

Figure 5. Distal Violin Breccia. Roadcut where Hungry Valley Road first reaches Piru Creek.

within Ridge Basin on the northeast was continuously rejuvenated, so that steep talus deposits and alluvial cones were continually maintained. Although creep and small earthquakes may well have contributed to this scarp rejuvenation, major earthquakes may be documented by marked lobes of the sedimentary breccia that prograded down steep fans toward the trough of Ridge Basin. This depositional reconstruction is interpreted to account for the great thickness of the coarse debris strewn out in a band along the transform fault. Similar systems have probably operated within transform boundary zones elsewhere, and may account for the distribution of sedimentary breccias there as well. Uplift and deep disection of the central Transverse Ranges has allowed erosion to expose the Violin Breccia and neighboring rocks so that the system can be documented whereas in many other regions, critical evidence may be buried or has been eroded away.

Figure 6. Violin Breccia clast counts. Pie diagrams graphically show percentage of stones of different types, largely identified with a handlens only. Abbreviations: gr = granitic rocks, including quartz diorite, quartz monzonite, and granite primarily; gg = heterogeneous gray gneiss, variable in structure, and including migmatites; pg = augen gneiss, primarily with cuniform porphyroblasts but also including large ovoid porphyroblasts of aggregates of potash feldspar and quartz; vg = gneiss with distinctive violet-colored quartz; ph = phyllonite, similar to that exposed within the Alamo Mountain movement zone; am = amphibolite and very dark gneisses; ad = diorite and related rocks, some verging on gabbro; an = anorthosite, including diorite transitional into anorthosite; o = others. Sites of pebble counts shown with respect to the trace of the San Gabriel fault zone.



CLAST COUNTS, VIOLIN BX

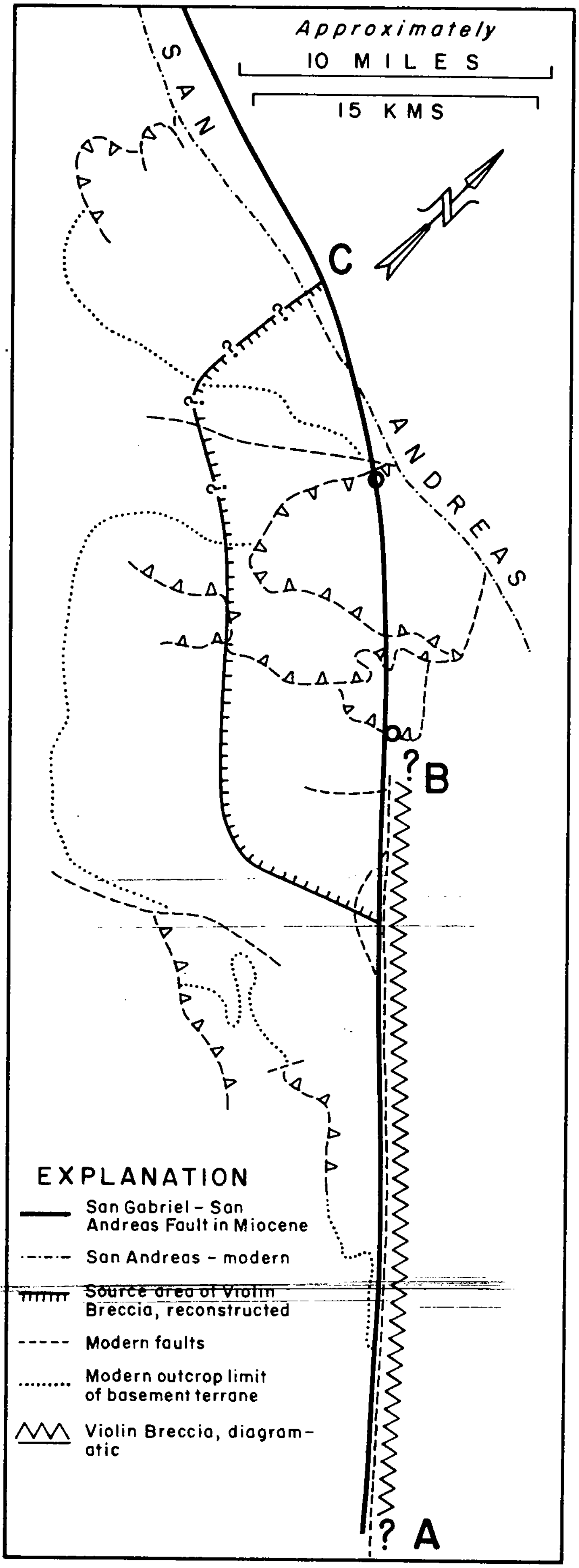


Figure 7. Diagrammatic map showing reconstruction of size of the source area for the Violin Breccia with respect to other structures in the region west and northwest of Ridge Basin. A = southernmost exposure of the Violin Breccia and the oldest beds; B = northwesternmost exposure and the youngest beds at the surface; C = conjectural position of the northwestern boundary of the source region.

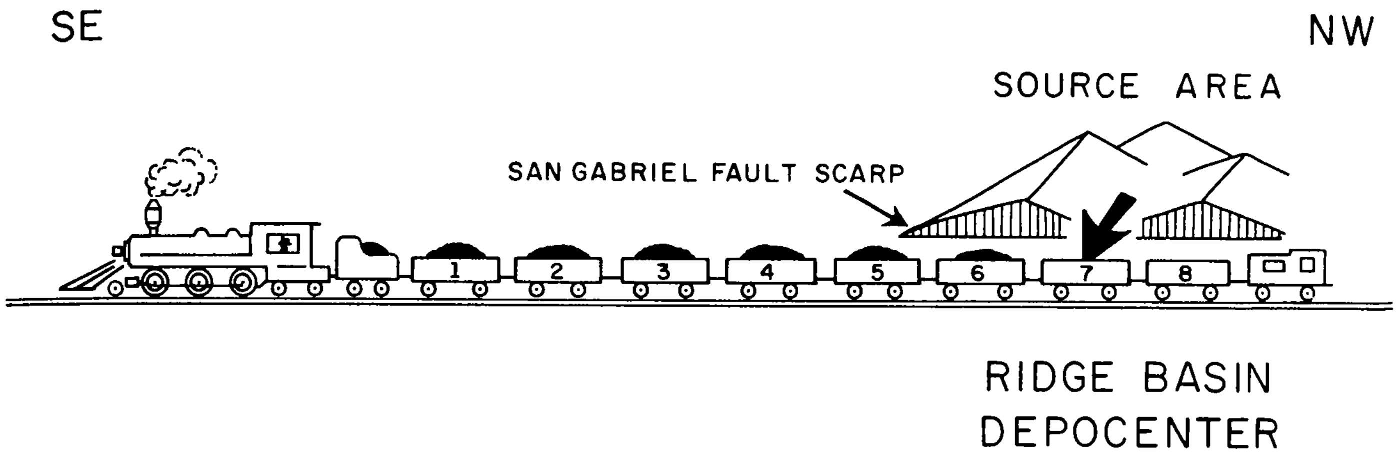


Figure 8. The Violin Breccia Coal Train. Diagram showing the analogy between the deposition of the Violin Breccia and the loading of a coal train opposite a hopper.

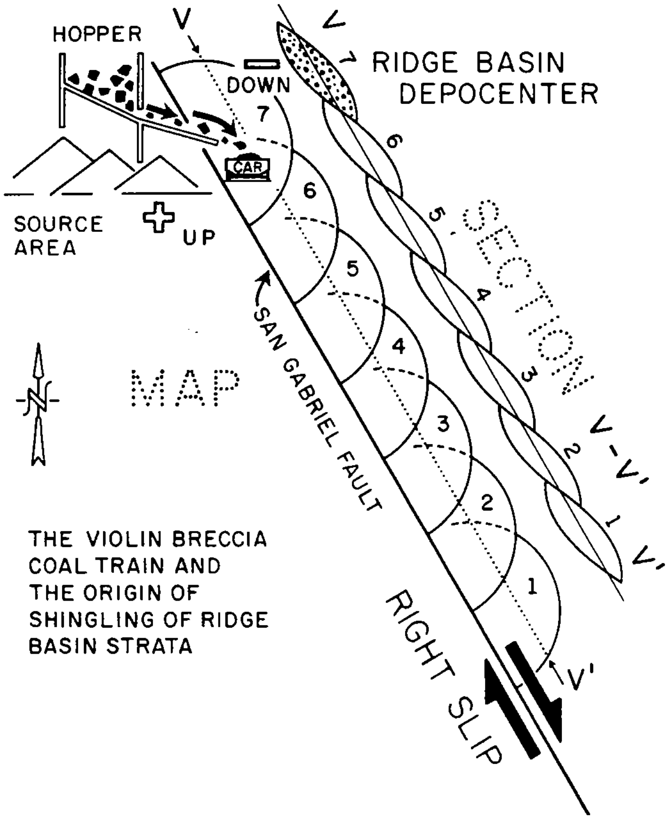


Figure 9. May view of stratal shingling resulting from the coal-train analogy. Refer to text.

Air view looking northwest across Frenchman Flat in foreground showing belt of Violin Breccia and trace of San Gabriel Fault through notch to left of round to the left of center. Frazier Mountain on skyline in distance. Photograph taken before construction of Pyramid Dam and impounding of Pyramid Lake.

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