Late Cenozoic Tectonism of the Sacramento Valley, California

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USER: This reprint of figure 5 should replace the present figure 5 of U.S. Geological Survey Professional Paper 1359. It was reprinted to correct the seismic reflection profile which was inadvertently printed backwards.

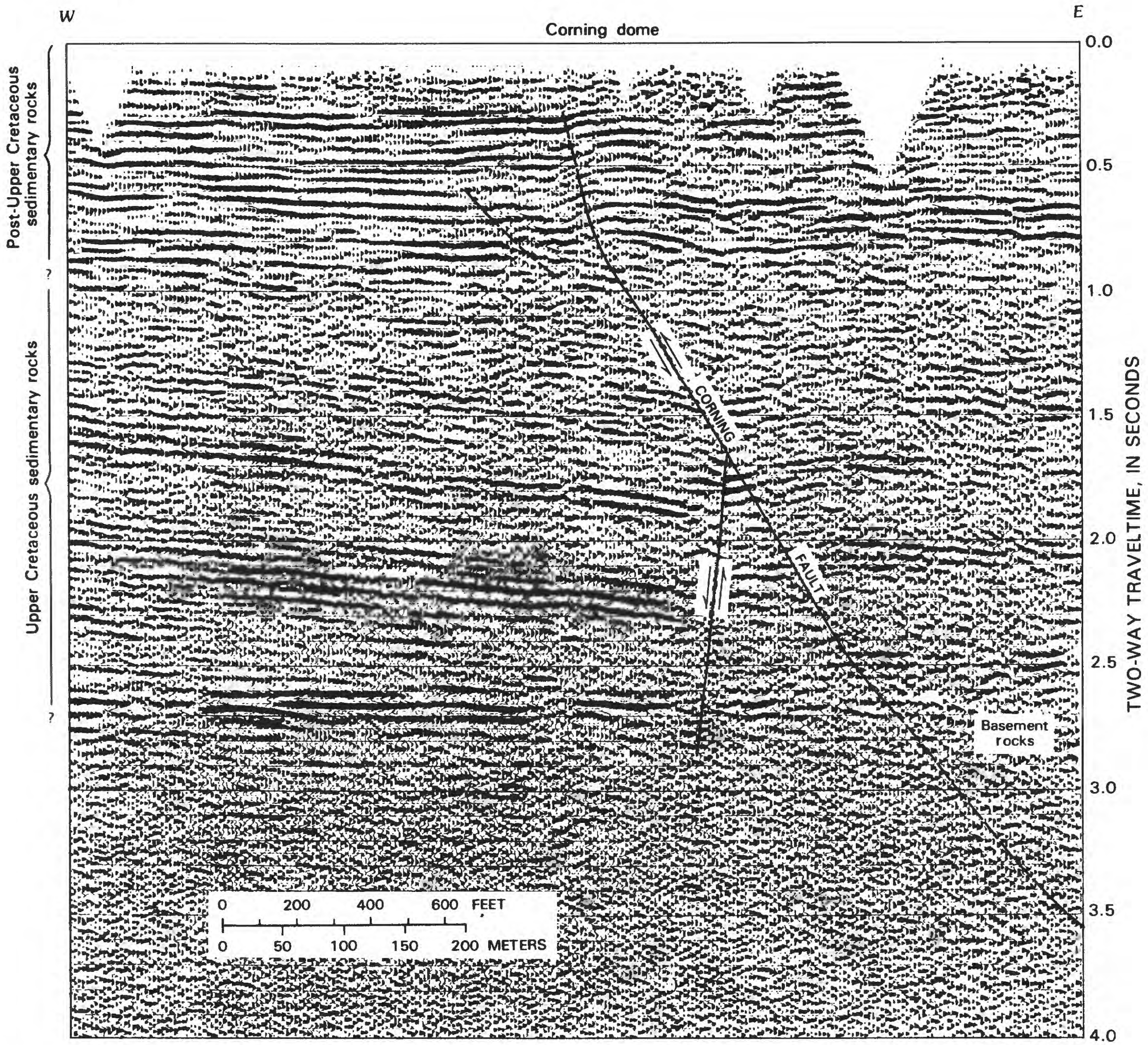

FIGURE 5.—Seismic reflection profile, from Seisdata Services Inc., showing structural and generalized stratigraphic relations across Corning fault between Corning and Red Bluff, Calif.

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By DAVID S. HARWOOD *and* EDWARD J. HELLEY

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LATE CENOZOIC TECTONISM OF THE SACRAMENTO VALLEY, CALIFORNIA

By DAVID S. HARWOOD and EDWARD J. HELLEY

ABSTRACT

Structure contours drawn on top of the Cretaceous rocks in the Sacramento Valley define a large number of diversely oriented folds and faults that are expressed in topographic, hydrologic, and geologic features at the land surface. Although many of the structures in the valley have a protracted history of movement, some dating back to the late Mesozoic, a remarkable number of these structures show late Cenozoic deformation that can be accurately determined from folding and faulting of widespread, dated Pliocene and Pleistocene volcanic units. These time-stratigraphic units are used to define structural domains of essentially contemporaneous late Cenozoic deformation that was characterized by east-west compressive stress. The oldest structural domain is located in the southeastern part of the valley, where east-side-up reverse movement on the Willows fault ceased prior to deposition of continentally derived sediments of late Miocene and early Pliocene age. In the middle Pliocene to early Pleistocene, east-west compressive deformation progressed northward through the valley so that the youngest late Cenozoic deformation is recorded in east-northeast-trending folds and faults in the Battle Creek domain, at the northernmost part of the valley. The northward progression of east-west compressive deformation appears to be related to the northward eclipse of eastward subduction of the Juan de Fuca plate before the northwestward migration of the Mendocino triple junction along the continental margin west of the valley.

Much of the east-west compressive stress that affected the valley in the late Cenozoic was accommodated by east-side-up reverse movement on the steeply east-dipping, northwest-trending Willows fault and the north-trending Corning fault that splays off from the main stem of the Willows fault north of Sutter Buttes. Significant strain release also occurred on the northwest-trending fault beneath the Chico monocline and on the east-northeast-trending Red Bluff, Battle Creek, and Bear Creek faults in the past 2.0 m.y. Southeast of Sutter Buttes, the Willows fault follows the boundary between dense, magnetic, presumably ophiolitic basement to the west and Sierran basement to the east. The Chico monocline follows the same basement boundary north of Sutter Buttes, but that structure is stepped eastward from the trace of the Willows fault. It seems reasonably certain that the southeastern extension of the Willows fault and the Chico monocline fault are middle and late Cenozoic structures, respectively, that owe their existence and orientation, in part, to earlier, Mesozoic tectonic juxtapositioning of significantly different basement terranes.

INTRODUCTION

The Oroville earthquake ($M = 5.6$, August 1, 1975) added a new dimension to geologic studies of the Sacramento Valley. That event, more than any of the infrequent earthquakes that previously occurred in the area, awakened the geologic and engineering communities to the potential of active faulting in the valley and in the adjacent Sierran foothills. Because the Oroville earthquake occurred near the Lake Oroville dam (fig. 1) at a time when other large dams were being constructed along the Sierran foothills, several geologic studies, including this one, were conducted to specifically evaluate the late Cenozoic structural history of the region (Woodward-Clyde Consultants, 1977; Harwood and others, 1981; Helley and others, 1981; Helley and Harwood, 1985).

For about a half century prior to the Oroville earthquake, most geologic investigations of the Sacramento Valley (fig. 1) focused on locating natural gas in the deeper parts of the valley fill and water resources in the near-surface deposits. These investigations produced a wealth of stratigraphic and structural data that was gathered primarily from numerous gas fields scattered throughout the valley (map A, pl. 1). Several geologic syntheses resulted from these studies, but they were designed primarily to facilitate further exploration for these valuable resources and did not emphasize late Cenozoic tectonism (Bryan, 1923; Repenning, 1960; Olmsted and Davis, 1961; Safonov, 1968; Hackel, 1966; Redwine, 1972; California Department of Water Resources, 1978).

Recent developments in the concept of plate tectonics have given new emphasis and scope to regional geologic analyses by enabling integration of studies in the valley with those in the Coast Ranges to the west and the Sierra Nevada to the east (fig. 2). Within this concept, the Sacramento Valley was interpreted to be a late Mesozoic forearc basin (Dickinson, 1970; Ingersoll, 1976, 1978a, 1978b; Dickinson and Seely, 1979) that formed contemporaneously with and between the accretionary trench deposits of the Franciscan Complex to the west (Ernst, 1965, 1970; Blake and Jones, 1974; Blake and others, 1974) and an eastern magmatic arc complex, the roots of which are exposed in the Sierra Nevada (Hamilton, 1969). Although debate about the extent, geographic origin, and nature of the material accreted to the Mesozoic continental margin continues (Coney and others, 1980), the basic three-component model of an arc-trench system remains fundamental (Dickinson and Seely, 1979). Although the Sacramento Valley may be the most completely studied ancient forearc basin in the world (Ingersoll and others, 1977), the emphasis and scope of published paleogeo-

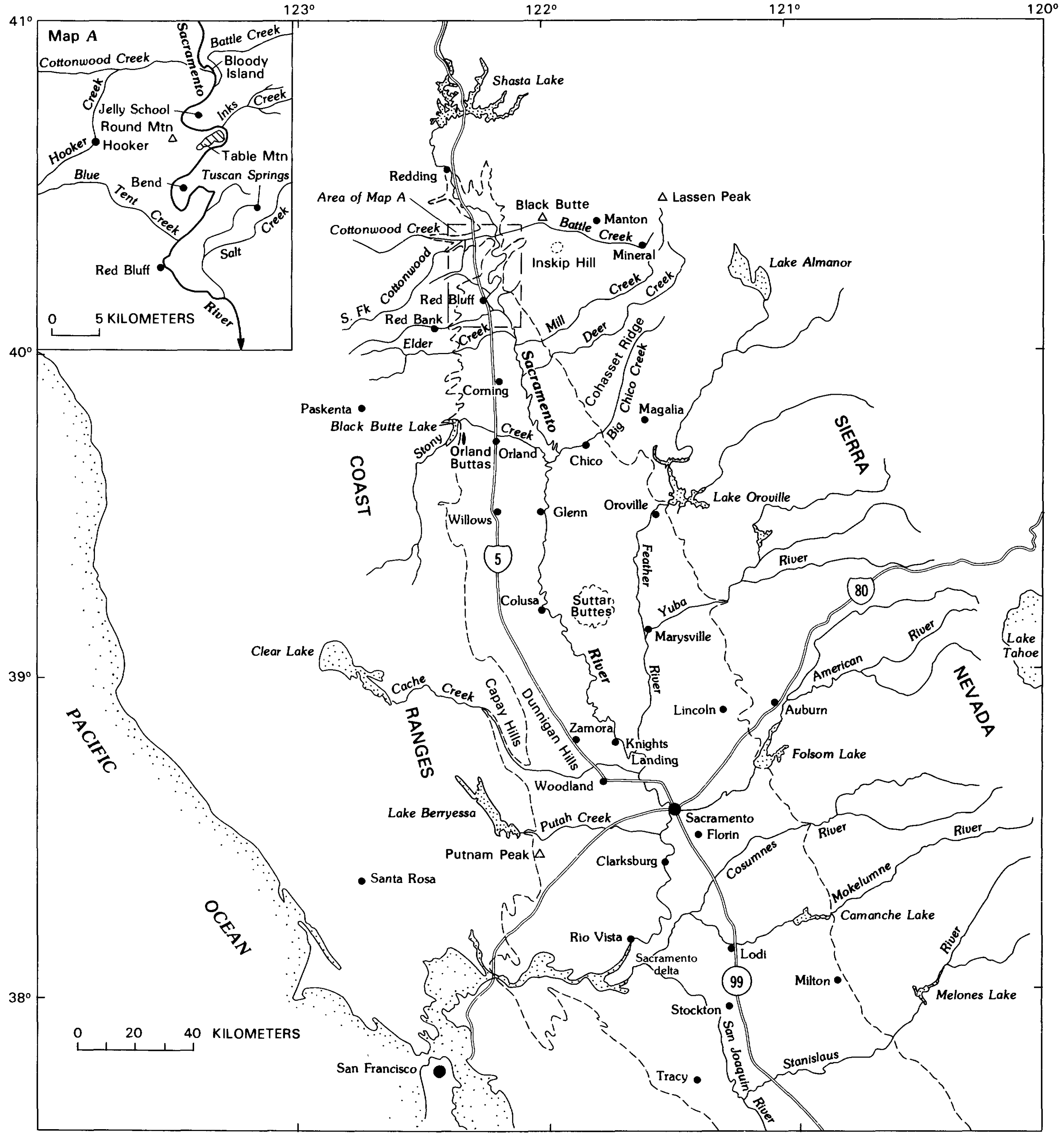


FIGURE 1.—Index map of northern California showing 500-foot topographic contour (dashed), which coincides approximately with boundary of Sacramento Valley.

graphic and paleotectonic studies generally precluded detailed analysis of late Cenozoic tectonism.

Structures discussed in this report are generally too young to provide new data for the late Mesozoic arc-trench model. In fact, a major objective of this report is to document that most of the folds and faults in the valley either formed in the late Cenozoic or are older structures that have undergone renewed deformation in the past few million years. In this respect, structures in the valley are more closely related to tectonic models of Cenozoic subduction, including subsequent lateral convergence along the San Andreas fault system and extension in the Basin and Range province to the east, than they are to the late Mesozoic tectonic models (Hamilton and Myers, 1966; Atwater, 1970; Christiansen and Lipman, 1972; Atwater and Molnar, 1973; Dickinson, 1979).

ACKNOWLEDGMENTS

The structure contour map presented in this report (pl. 1) integrates subsurface data from many summary reports of oil and gas operations published by the California Division of Oil and Gas. Credit is given to specific authors of gas field reports in the "Sources of Data" shown on plate 1, and we would like to express our appreciation to those authors, whose work contributed much of the basic data for our compilation. In addition to published reports, we analyzed well logs from a number of areas in the valley, and we acknowledge the courteous service provided by the staff of the Woodland office of the California Division of Oil and Gas.

Interpretation of the subsurface data was influenced to a great degree by exceptionally high-quality seismic-reflection profiles purchased from Seisdata Services Inc. We thank Seisdata Services Inc. for allowing us to reproduce a small segment of one of those profiles in this report.

We thank David Wagner of the California Division of Mines and Geology for sending us a preprint of his detailed map and discussion of the Capay Valley-Capay Hills area. Special thanks are given to the California Academy of Sciences for making available their collection of thin sections of basement rocks that were penetrated in wells in the Sacramento Valley.

We also acknowledge stimulating discussions with our colleagues J.A. Bartow, M.C. Blake, Jr., K.F. Fox, Jr., D.L. Jones, R.J. McLaughlin, T.H. Nilsen, A.M. Sarna-Wojcicki, and C.M. Wentworth. These discussions sharpened the focus of many points in this report, but responsibility for the analysis of subsurface structures and regional tectonism rests with Harwood, and the mapping of surficial deposits, which document the youthfulness of many structures, rests with Helley.

REGIONAL SETTING

The Sacramento Valley comprises the northern third of the Great Valley of California (fig. 2)—a broad, fertile lowland situated between mountainous terrains of the Coast Ranges to the west and the Sierra Nevada to the east. Marine sedimentary rocks, ranging in age from Late Jurassic to early Miocene, underlie the deeper parts of the Sacramento Valley. They are unconformably capped by a relatively thin cover of alluvial deposits and locally prominent volcanic rocks of early Miocene to Holocene age.

Marine sedimentary rocks exposed on the west flank of the Sacramento Valley record a nearly unbroken depositional sequence from Upper Jurassic through Upper Cretaceous (Lachenbruch, 1962; Ingersoll and others, 1977). On the east side of the valley, Upper Cretaceous sandstone and shale, described by Taft and others (1940) and Haggart and Ward (1984), rest unconformably on metamorphic and plutonic rocks of the Sierra Nevada basement and indicate a progressive eastward onlap of marine sedimentation during the late Mesozoic (Ingersoll and others, 1977). In a broad area south of Sutter Buttes, younger Upper Cretaceous sandstone and shale record a westward-prograding deltaic sequence and marine regression (Drummond and others, 1976; Garcia, 1981) during the Late Cretaceous, as the depositional basin was broadly uplifted and tilted to the south and west. Intermittent periods of uplift and subsidence, apparently caused by Paleogene subduction and possibly by lateral faulting to the west (Nilsen and Clarke, 1975; Nilsen and McKee, 1979), affected the depositional basin through the early Tertiary.

Four distinct submarine canyons developed during the uplift cycles, and they were subsequently filled with transgressive marine sequences during times of subsidence (fig. 2). Almgren (1978) has documented three ancient oscillatory tectonic-depositional cycles in the southern part of the Sacramento Valley. He concludes that the Martinez canyon was cut in the late early to middle Paleocene, the Meganos canyon was cut in early Oligocene and filled in late Oligocene and early Miocene, and the Markley canyon was cut in early Oligocene and filled in early Miocene time. Although the timing of the fourth submarine canyon is not as closely constrained by the sedimentary record as the other three, Redwine (1972) concluded that the Princeton canyon, located in the north-central part of the valley (fig. 2), was cut and filled between the late Paleocene and early Eocene, probably coincident with the formation of the Martinez canyon to the south (Almgren, 1978). Eocene marine deposits in the Princeton canyon apparently grade northward and eastward into alluvial deposits derived from the northern Sierra Nevada.

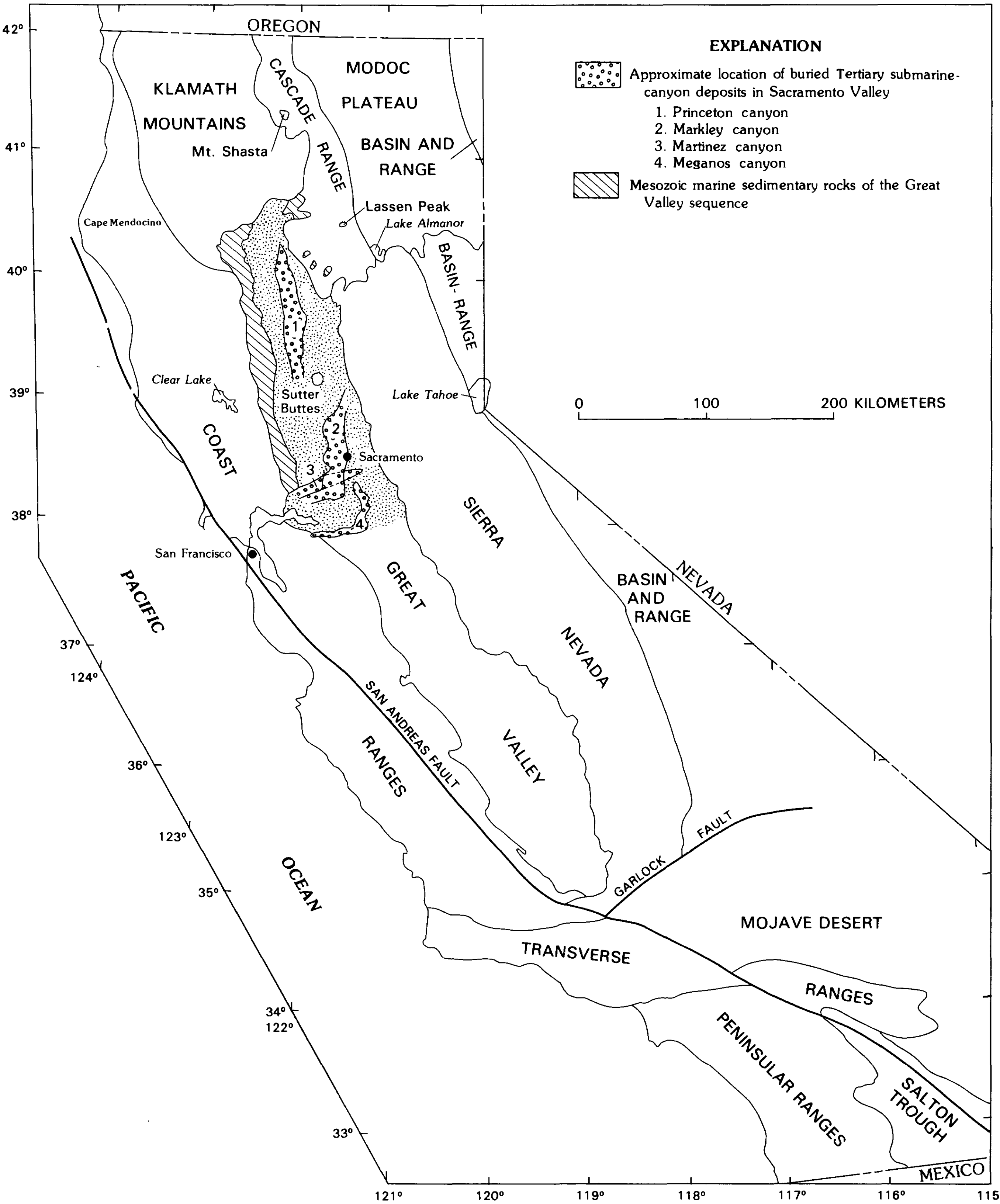


FIGURE 2.—Outline of geomorphic provinces in California and general location of buried Tertiary submarine canyon deposits in Sacramento Valley (stippled).

By the early Miocene, the northern part of the Sacramento Valley clearly was an emergent area, subjected to fluvial erosion and deposition. During this time, between 23.8 and 22.3 m.y. ago according to Dalrymple (1964), the Lovejoy Basalt was erupted from a vent or vents in the northern Sierra Nevada (Durrell, 1959) and flowed westward in a confined channel until it reached the Sacramento Valley near Oroville. Upon reaching the valley, the Lovejoy Basalt spread laterally, filling distributary alluvial channels (van den Berge, 1968) and flowed both westward, across the present axis of the valley now occupied by the Sacramento River at least to Orland Buttes, and southward at least to the vicinity of Putnam Peak (Weaver, 1949) (fig. 1), 260 km from the presumed source. Because of its distinctive characteristics, widespread distribution in both the surface and subsurface parts of the Sacramento Valley, and radiometrically determined age, the Lovejoy Basalt is an important marker unit (pl. 1) that places an age constraint on Cenozoic deformation in the valley.

After eruption of the Lovejoy Basalt, alluvial sediments of late middle Miocene to early Pliocene age accumulated in the central and southern part of the valley, capping marine sedimentary rocks deposited in the Princeton and Markley canyons (Redwine, 1972; Almgren, 1978). These alluvial deposits, however, are not present in surface exposures in the northern part of the valley, where Pliocene alluvium of the Tehama Formation in the northwest and coeval volcanic rocks of the Tuscan Formation in the northeast rest directly on the Lovejoy Basalt or on older rocks where the Lovejoy is missing.

A widespread felsic pumiceous tuff, the Nomlaki Tuff Member of the Tuscan and Tehama Formations, occurs at or near the base of the Pliocene rocks and serves as a second important time-stratigraphic marker in the valley, dated at 3.4 m.y. (Evernden and others, 1964). The Nomlaki, which was erupted from the vicinity of Lassen Peak, generally marks the beginning of volcanic activity that has occurred intermittently into historic time at the southern end of the Cascade Range (fig. 2). Pliocene and Pleistocene volcanic rocks, erupted from the Lassen area, flowed westward and interfingered with alluvial deposits of the ancestral Sacramento River system in the northern part of the valley (Anderson, 1933; Lydon, 1968; Harwood and others, 1981; Helley and others, 1981). Several of the volcanic units have been dated (Harwood and others, 1981), and they are important time indicators for the late Cenozoic depositional and tectonic evolution of the Sacramento Valley. Two informally designated volcanic units, the Rockland ash bed (referred to as the ash of Mount Maidu by Harwood and others, 1981) and the basalt of Deer Creek, dated at 0.45 and 1.09 m.y., respectively (Meyer and others, 1980; Harwood and others, 1981), are particularly significant because they bracket the age of the Red Bluff Formation. The Red Bluff forms a thin, widespread gravel pediment that unconformably caps the Pliocene rocks in the northern part of the valley. The fact that the Red Bluff Formation is faulted and folded clearly establishes the presence of late Cenozoic deformation, within the past million years, on most of the major features shown on the structure contour map (pl. 1).

A volcanic center, composed of andesitic fragmental deposits intruded by late-stage rhyolite domes, formed at Sutter Buttes in the central part of the valley between 2.4 and 1.4 m.y. ago (Williams, 1929; Williams and Curtis, 1977). This late Cenozoic volcanic center is located over a northwest-trending tectonic boundary that juxtaposes a basement of dense, magnetic, presumed oceanic crust (Cady, 1975; Griscom, 1973) on the west against metamorphic and plutonic rocks of the Sierran basement on the east. The boundary between the basement terranes is a Mesozoic tectonic feature that apparently has been reactivated by Cenozoic deformation (Harwood, 1984). Southeast of Sutter Buttes, the Willows fault approximately follows the contact between oceanic and Sierran basement terranes, and northeast of Sutter Buttes that basement boundary lies beneath the Chico monocline.

In the southeastern part of the valley, structure contours are drawn on the basement surface (pl. 1). In the rest of the valley, the datum is the top of the Cretaceous rocks. Contouring procedures are described on plate 1.

INTERPRETATION OF BASEMENT CONTOURS

The most pronounced deflection in basement contours is located beneath Sutter Buttes (pl. 1). This anomaly in basement topography was shown first by Smith (1964) and mentioned briefly by Williams and Curtis (1977, p. 6), who noted more than 395 m of local relief on the basement surface but offered no interpretation of the anomaly. Although some of the upwelling of the basement surface may be related to volcanic intrusion at Sutter Buttes, Harwood (1984) suggests that much of the basement uplift was related to east-side-up movement on the Willows fault, which passes just west and south of the Sutter Buttes.

Data from deep wells a few kilometers north of Sutter Buttes (pl. 1) indicate that the basement surface has a major deflection related to a northeast-trending fold or fault, or both. All of these possible structural configurations were investigated, and we conclude that the structure in the area is a northeast-trending fold, faulted along its southeast limb. The fault is interpreted to be a steeply dipping to vertical normal fault with southeast-side-down displacement. Both the fold and fault, however, appear to die out to the northeast.

In a zone southeast of Sutter Buttes that extends to the Stockton fault (pl. 1), variations in the slope of the basement surface are shown by northwest-trending zones of broadly spaced and relatively close-spaced contours. The basement surface may dip eastward locally in those areas marked by broadly spaced contours (see Jacobsen, 1981), but well data are too sparse to locate closed contours in this area, even if they would show at the chosen contour interval of 150 m.

From Sacramento south to the Stockton fault, the slope of the basement steepens significantly west of the −1,500-m contour. This change in basement slope lies near the eastern edge of the positive gravity and magnetic anomalies in the Great Valley that Cady (1975) interpreted to represent ophiolitic oceanic crust tectonically juxtaposed against Sierran basement rocks to the east. Reports of two wells south of Sacramento, however, show "granite" and "gneiss" at −1,954 m and −2,861 m, respectively, suggesting that the interpreted suture between oceanic and Sierran basements is not marked by the topographic break in the basement surface at the −1,500-m contour. It must be emphasized, however, that these identifications of basement rocks were made some time ago by the operators who drilled the wells, and it is possible that the "granite" may be a leucocratic rock associated with ophiolitic basement, such as plagiogranite, and the "gneiss" may be any dark-colored rock, even part of the ophiolitic basement. It seems reasonably certain, in any case, that the topographic break at the −1,500-m contour is controlled in some measure by a fault, possibly the extension of the Willows fault southeast of Sutter Buttes. It is interesting, and not altogether fruitless, to speculate that the Willows fault southeast of Sutter Buttes coincides with and represents the up-section propogation of the major fault between oceanic crust and the arc-massif basement of the Sierra Nevada. If such is the case, the Willows fault southeast of the Sutter Buttes may represent a structure inherited from the upper-slope discontinuity, a region of structural weakness over the boundary between forearc-basin and arc-massif basements, identified by Karig and Sharman (1975) in modern forearc basins.

From Sutter Buttes north to Chico, a number of wells penetrate basement rocks reported as diorite, gabbro, noritic gabbro, and serpentine. Not surprisingly, this part of the Sacramento Valley is characterized by large, positive gravity and magnetic anomalies that mark the northern part of the inferred Great Valley ophiolitic basement (Cady, 1975). Some thin sections of gabbro and diorite from wells in this area, contained in a collection of thin sections of basement rocks held by the California Academy of Sciences, reveal remarkably fresh coarse- and medium-grained clinopyroxene-plagioclase rocks that show pronounced cumulate textures. These textures, previously unrecognized, provide supporting evidence for the interpretation that the source rocks of the Great Valley anomaly are oceanic crust.

South and east of Sutter Buttes, several wells penetrate rocks reported as granite, granodiorite, metavolcanic rocks, and "gneiss" typical of rocks exposed in the Sierran foothills. The outline of granitic plutonic rocks, shown on plate 1, was determined from reported basement rocks, augmented by our interpretation of residual isostatic gravity data (Roberts and others, 1981). The boundary between the Sierran basement and mafic and ultramafic rocks of probable ophiolitic basement trends about N. 45° E. from Sutter Buttes to Oroville. It is clearly marked by significant changes in the gravity and magnetic patterns of the respective areas. The basement contours do not reflect this inferred lithologic change, and the northeast-trending fault north of Sutter Buttes is contained within the mafic-ultramafic basement terrane.

Sierran crystalline rocks are exposed in several deep canyons that cut through the Pliocene volcanic rocks of the Tuscan Formation east of Chico (Harwood and others, 1981). The mafic-ultramafic basement appears to abut against Sierran basement rocks along a major basement fault beneath the Chico monocline. This tectonic juxtaposition of basement terranes was proposed first by Griscom (1973), and the basement fault he proposed from geophysical data is confirmed by stratigraphic data. The basement fault beneath the Chico monocline shows east-side-up displacement with a minimum stratigraphic throw of 367 m on the basement surface. This fault represents late Cenozoic deformation superposed on a segment of the late Mesozoic boundary between the forearc-basin and arc-massif basements, and it is thus part of the upper-slope discontinuity (Karig and Sharman, 1975).

STRUCTURAL FEATURES IN THE VALLEY FILL

Structure contours drawn on top of the Cretaceous rocks outline a large number of diversely oriented folds and faults distributed throughout the valley. Some of these structures, such as the Corning domes, the Chico monocline, and the Dunnigan Hills anticline, have topographic expression and were recognized in the early ground-water studies of the valley by Kirk Bryan (1923). Most of the other structures do not have obvious surface expression, and they were discovered during exploration and development of the numerous small gas fields in the valley (pl. 1, map A). Although detailed reports have been published for many of the gas fields, few reports have synthesized the data from these studies into detailed structural analyses of the valley as a whole. Noteworthy exceptions to this generalization exist, however, and include the early stratigraphic and structural synthesis by Safonov (1968), the detailed study of the Princeton submarine channel system by Redwine (1972), and the continuing efforts of the California Division of Mines and

Geology to compile available data from the valley into small-scale maps of the entire state (Jennings, 1977).

The extent of many of the faults, particularly the complex pattern of the Willows fault and the relations between that fault system and many folds in the northern valley, has not been reported previously. Analyses of well logs, which are the basis for most reports of individual gas fields, do not provide sufficient data to interpret unequivocally the structure of each field. If recognizable stratigraphic units in the well logs are at different elevations in adjacent wells, it is generally impossible to determine if the units are offset by folding or faulting, or both, without additional data from surface exposures or seismic-reflection profiles. Our interpretation of the structural relations in the valley is based on data from well logs, augmented by detailed mapping of the surficial deposits by Helley, and exceedingly valuable seismic-reflection data obtained from Seisdata Services Inc.

WILLOWS FAULT SYSTEM

The main stem of the Willows fault was discovered in the subsurface rocks of the northern valley when it was penetrated by the Marathon Oil Company (formerly Ohio Oil Company) "Capital Company No. 1" well during development of the Willows-Beehive Bend gas field in the late 1950's (California Division of Oil and Gas, 1960; Alkire, 1962; 1968). From the discovery well, Redwine (1972) traced the Willows fault in the subsurface southeast to Sutter Buttes and suggested that it extended northwest of the discovery well, possibly connecting with the surface fault mapped west of the Orland Buttes (Anderson and Russell, 1939; Jennings and Strand, 1960). On the basis of the distribution of Quaternary units immediately south of Orland Buttes and of two seismic profiles north of the buttes, we have extended the Willows fault into northwestern Tehama County. At Orland Buttes, Upper Cretaceous rocks, the Lovejoy Basalt, and the Tehama Formation, on the up-thrown side of the Willows fault, are juxtaposed against the Tehama Formation on the down-thrown fault block to the west.

Redwine (1972) documented displacement on a 40-km segment of the main stem of the fault from east of Willows to Sutter Buttes and concluded that it dipped 74° or steeper to the east and showed reverse, east-side-up movement that decreased upward toward the surface. He found that the Princeton submarine channel was localized by movement on the fault and that vertical separation in the discovery well varied from about 488 m on top of the Cretaceous rocks to about 477 m on the top of the Eocene Capay Formation.

The pattern of deformation on the Willows fault and its effect on the thickness and facies patterns in the Capay Formation are clearly shown in figure 3. In this approximately east-trending cross section through the Marathon "Capital Company No. 1" well (No. 3, fig. 3), the thalweg of the Princeton channel, which is occupied by marine shale of the Capay, coincides with a syncline in the underlying Upper Cretaceous rocks. The west limb of the complementary anticline in the Upper Cretaceous rocks is broken by east-side-up movement on the Willows fault. The electric logs show that west of the fault the Capay is composed of shale, and east of the fault it is dominantly sandstone that interfingers with shale eastward toward the thalweg of the channel. Folding and faulting clearly postdated the deposition of Upper Cretaceous strata and preceded the deposition of the Capay and occurred between about 60 and 53 m.y. ago. During that time there was some erosion of the Upper Cretaceous rocks, particularly along the channel thalweg. The westward increase in sandstone and a complementary decrease in thickness of the Capay toward the fault suggest that the anticlinal welt of Upper Cretaceous rocks was a shoal area during Capay deposition. Because the shale-rich Capay west of the fault is about the same thickness as the sandstone-rich Capay east of the fault, there appears to have been little, if any, deformation on the Willows fault in this area during deposition of the Capay. The section shows about 475 m of post-Capay offset and possibly small offset on the base of the Tehama, but the amount of post-Pliocene offset is difficult to determine because there is no record for the upper part of well 2 (fig. 3).

CORNING FAULT

Data from a number of sources indicate that the Willows fault is far more extensive and complex than previously thought. The first clue that the Willows fault branched into a multistrand fault system was provided by an analysis of seismicity of the northern valley and Sierran foothills after the Oroville earthquake. Marks and Lindh (1978) located a number of small-magnitude earthquakes along a zone that originated near the Marathon "Capital Company No. 1" well in the Willows-Beehive Bend gas field and extended north about 30 km, rather than following the northwest trend of the Willows fault. These epicenters are shown relative to the trace of the Willows fault in figure 4. The trend of seismic events suggests that a north-trending fault splayed off from the main stem of the Willows fault and passed west of the Corning domes (pl. 1). In the Corning gas fields, analysis of well records by the Sacramento Petroleum Association (1962) showed an anticlinal fold in the area of the Corning domes, with about 121 m of maximum closure on the base of the Tehama Formation in the north dome and a steeply dipping southeast-trending fault located at the north end of south Corning dome, but it did not identify a fault west of them.

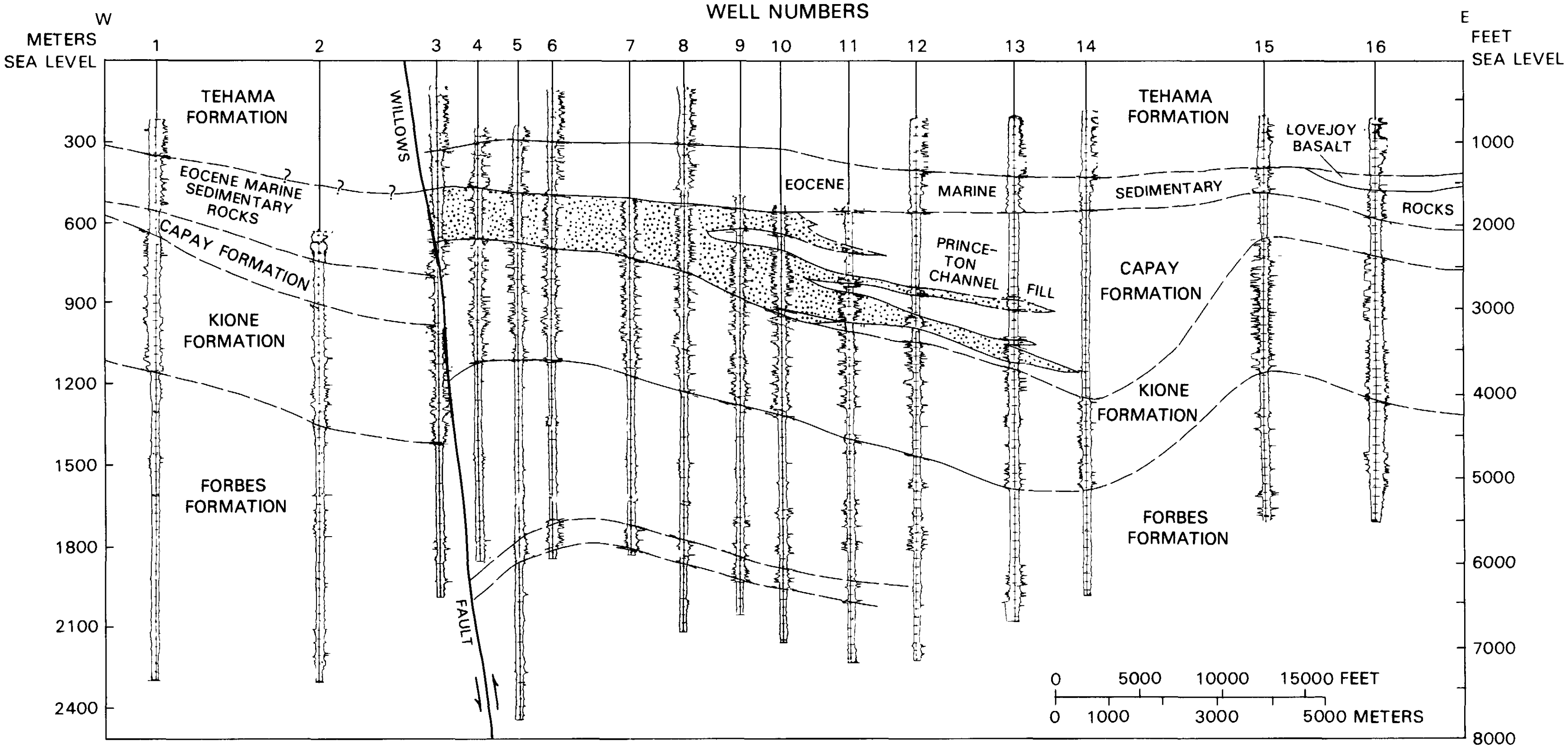


FIGURE 3.—Cross section showing electric logs and stratigraphic and structural relations across Willows-Beehive Bend gas field, Calif. Sandstone in Capay Formation indicated by stipple. Willows fault penetrated in well No. 3. Names of wells shown in appendix. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological Survey.

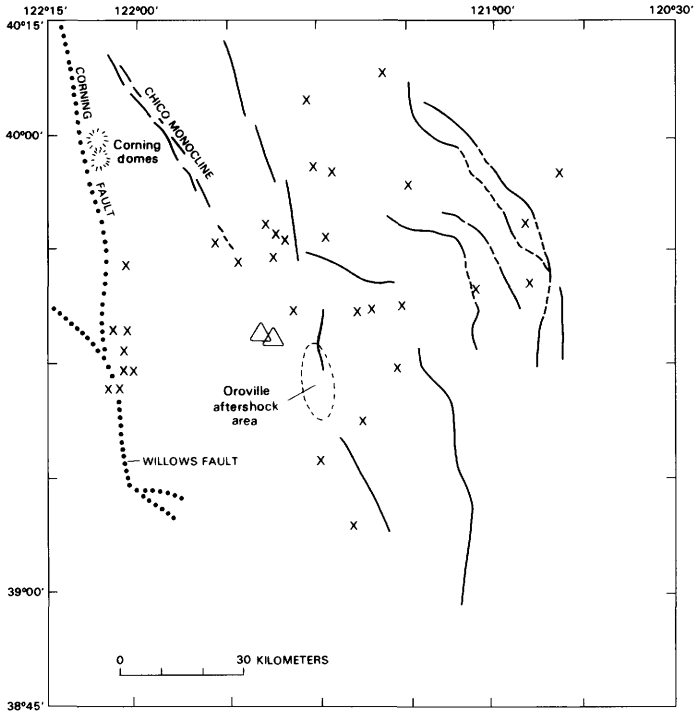


FIGURE 4.—Epicenters (X) of seismic events in area of Oroville, Calif., from June 1975 to August 1976 (modified from Marks and Lindh, 1978). Note north trend of epicenters extending from Willows fault approximately along trace of Corning fault. Unlabeled heavy lines indicate other faults. Triangles show epicenters of two unusually deep (40 km) earthquakes.

Seismic-reflection profiles in the area by Seisdata Services Inc. have identified a major north-trending, steeply east-dipping reverse fault that passes west of the Corning domes and the Greenwood anticline. We call this structure the Corning fault (pl. 1), and part of one of the profiles across it is shown in figure 5. The profile clearly shows that the shallow reflecting horizons in the eastern block are arched beneath the Corning dome by east-side-up drag on the Corning fault. It is also apparent in the profile that the vertical displacement increases with depth, indicating progressive deformation through time, similar to the deformation pattern on the main stem of the Willows fault in the Willows-Beehive Bend gas field (fig. 3). Curiously, however, the amount of vertical displacement of the Upper Cretaceous and younger strata shown on the profile is relatively small compared to the offset of the basement surface suggested by the few deep wells in the area shown on figure 6.

The deepest well east of the Corning fault is the Shell Oil Company "Victor Ranch No. 4" (sec. 7, T. 23 N., R. 2 W.), which bottomed in the Venado Sandstone of Late Cretaceous (Turonian) age at an elevation of −3,713 m (fig. 6). Lithologic information from this well helps interpret the seismic-reflection profile, and its total depth provides a minimum value for the basement elevation 8 km east of the Corning fault. Although local variations in the slope of the basement surface exist, the basement generally dips about 5° SW. between the Chico monocline and the Corning fault. The intersection of the basement surface and the Corning fault ranges in elevation from about −3,900 m in the north to about −4,600 m in the south. This south-southeast plunging line of intersection appears to lie significantly above the basement surface west of the Corning fault.

West of the Corning fault, only one well, the Shell Oil Company "Vilche No. 2" (sec. 5, T. 27 N., R. 4 W.), reached basement, which was serpentinite penetrated at an elevation of −5,978 m (fig. 6). Data from this well are not sufficient to determine the dip of the basement surface west of the Corning fault, but the dip immediately west of the fault may be the same as that just east of the fault, assuming that the basement surface was not chaotically tilted prior to deposition of the Great Valley sequence. If such is the case, the −6,000-m contour can be extrapolated south of the Shell Oil Company "Vilche No. 2" well parallel to the contours east of the Corning fault, as shown on figure 6. This extrapolation gives a vertical separation of about 1,500 m on the basement surface if one assumes no offset on the Red Bluff fault (fig. 6). Because the Red Bluff fault (discussed in a following section of this report) shows down-to-the-south displacement of the valley fill, similar basement offset on the Red Bluff fault would increase the basement separation on the Corning fault. Therefore, 1,500 m of vertical offset on the basement surface across the Corning fault is considered to be a minimum value.

The reason why the basement offset does not appear on the seismic reflection profile is not known. Perhaps the energy input was not sufficient to generate reflections from the deeper basement. Alternatively, the Venado Sandstone, which thickens westward, may have masked reflections from older rocks of the Great Valley sequence and the basement surface west of the Corning fault. Whatever the reason, geologic data strongly suggest that the basement was offset at least 1,500 m on the Corning fault prior to Turonian time and that pre-Turonian rocks of the Great Valley sequence, deposited on the downdropped block of the fault were unconformably overlapped by the Venado Sandstone. Post-Turonian displacement on the Corning fault is a few hundred meters.

The youngest deposits deformed by the Corning fault are gravels of the Pleistocene Red Bluff Formation, the age of which is between 0.45 and 1.09 m.y. (Harwood and others, 1981).

NORTHWEST EXTENSION OF WILLOWS FAULT

The location of the Willows fault system north and northwest of Orland Buttes is not clearly indicated by direct evidence in either the surface or subsurface rocks.

Wells are sparse in this area of the valley, particularly west of the probable trace of the main stem of the Willows fault. For that reason, the structure contours are very generalized and of little value in locating even major structures.

Our projection of the Willows fault into the Cold Fork, Elder Creek, and Paskenta faults (pl. 1) mapped by Jones and others (1969) is based primarily on the outcrop pattern of the Tehama Formation. North of Elder Creek, the Tehama Formation dips gently east, and the Nomlaki Tuff Member is at the base of the Tehama or is, at most, a few tens of meters above its base. South of Elder Creek, however, the Tehama dips more steeply eastward into the valley, and the Nomlaki is a few hundred meters above the base of the Tehama. This outcrop pattern of the Tehama Formation suggests that the underlying Great Valley sequence was topographically higher and projected farther east into the valley north of the Willows fault prior to deposition of the Pliocene strata. The position of the Nomlaki Tuff Member, relative to the base of the Tehama on opposite sides of the Willows fault, indicates that the Tehama filled a topographic low southwest of the fault

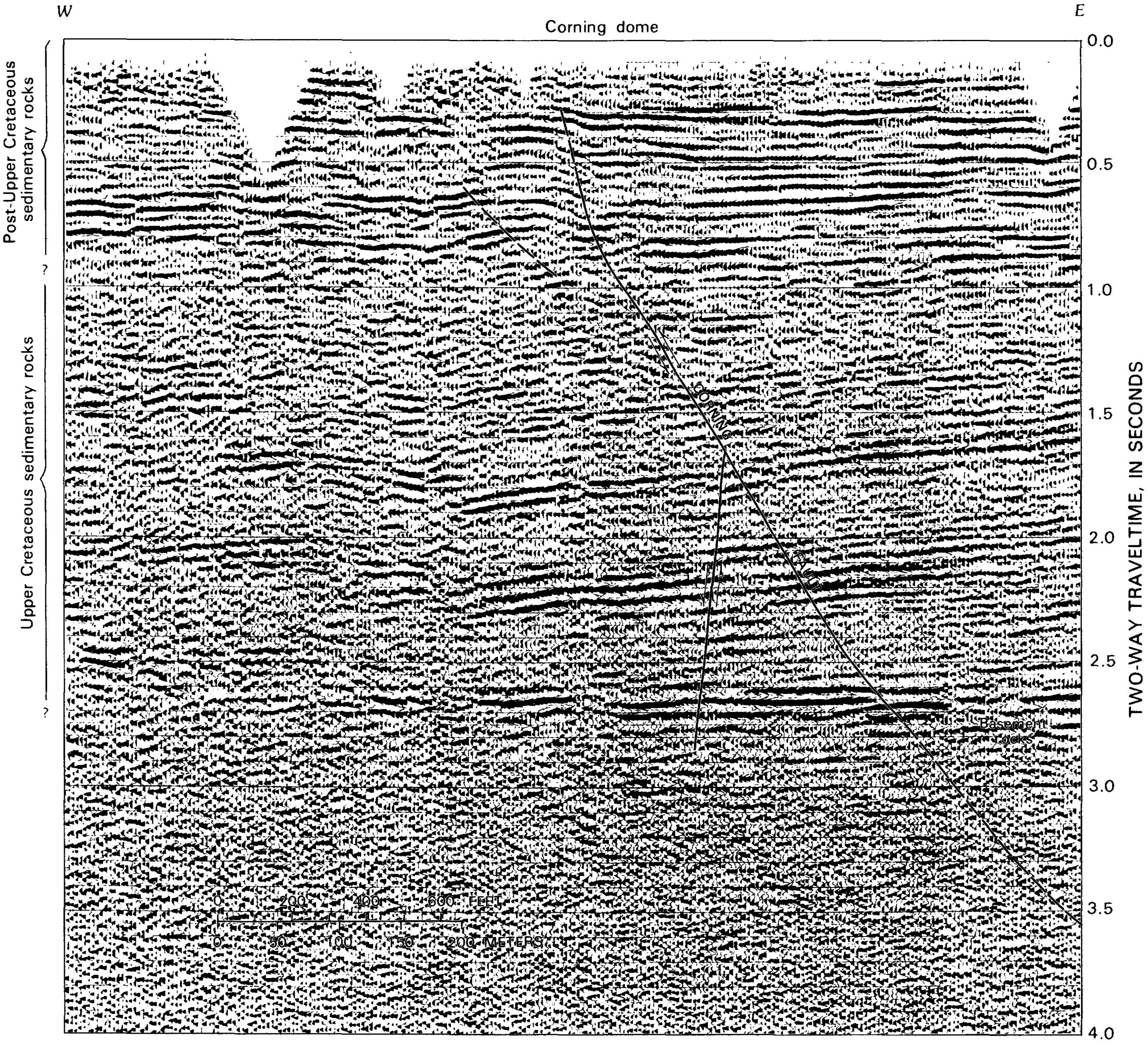


FIGURE 5.—Seismic-reflection profile, from Seisdata Services Inc., showing structural and generalized stratigraphic relations across Corning fault between Corning and Red Bluff, Calif.

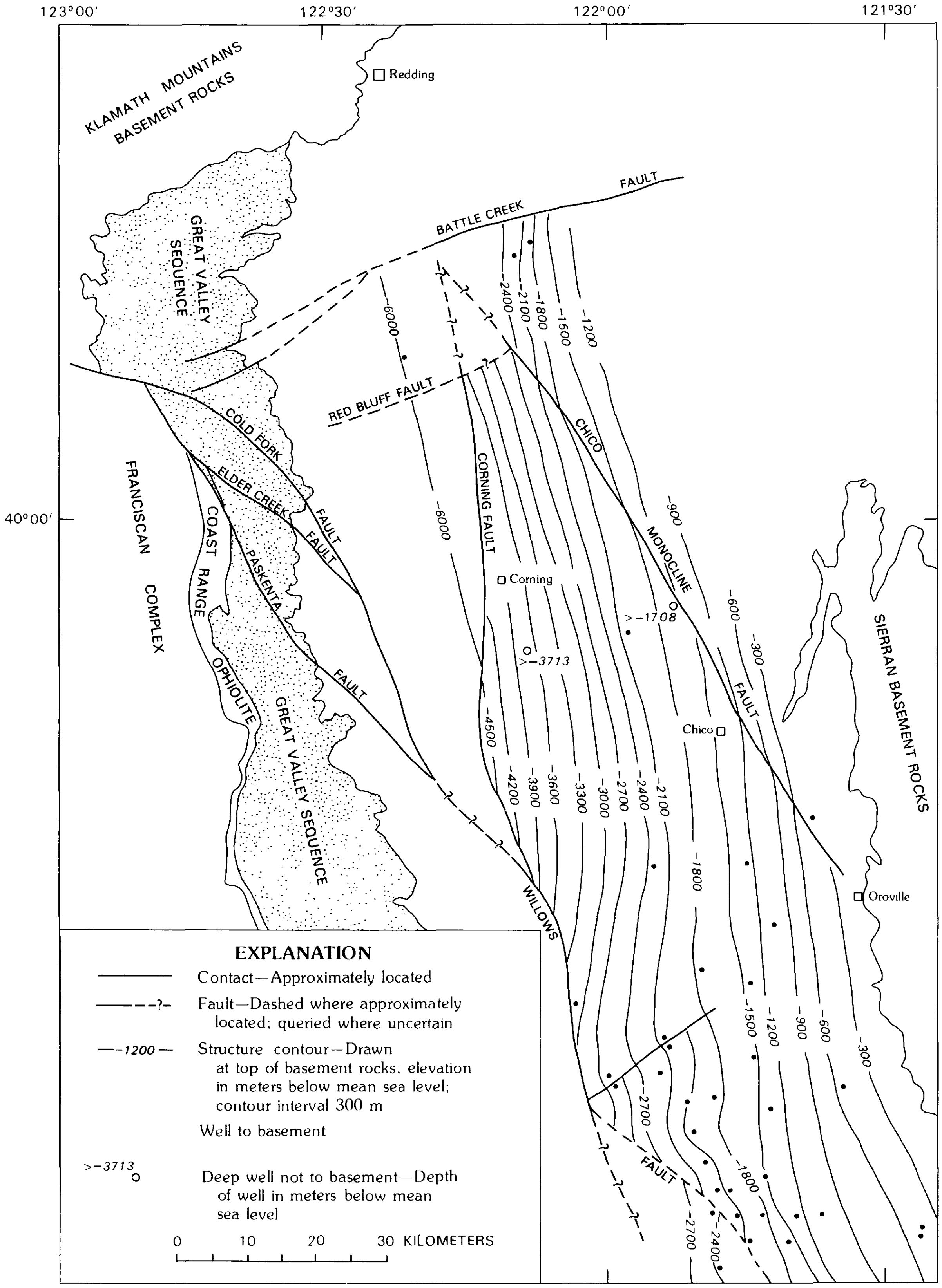


FIGURE 6.—Map of northern Sacramento Valley showing structure contours, drawn on top of basement rocks, used to estimate amount of vertical displacement of basement across Corning fault.

prior to eruption of the Nomlaki about 3.4 m.y. ago. We interpret the topographic low, reflected by the thicker basal part of the Tehama, to be the result of east-side-up movement on the Willows fault, prior to and possibly during, deposition of the lower part of the Tehama Formation. If this interpretation is correct, late Cenozoic movement on the Willows-Elder Creek fault system differs significantly, in style and amount of displacement, from the Cretaceous movement on the Elder Creek-Cold Fork-Paskenta fault system oulined by Jones and Irwin (1971).

Jones and Irwin (1971) inferred at least 96 km of left-lateral displacement of the Lower Cretaceous (Valanginian) strata along the combined Cold Fork-Elder Creek-Paskenta faults. They concluded that this deformation commenced shortly after deposition of the Valanginian strata and continued concurrently with deposition until at least the middle Late Cretaceous. Well-documented vertical displacement of Upper Cretaceous and overlying rocks on the Willows fault in the Willows-Beehive Bend gas field is not incompatible with left-lateral displacement on the Elder Creek fault system to the northwest, but it does indicate that the inferred lateral displacement was accompanied, or postdated by, a major component of east-side-up vertical movement. This vertical movement is consistent with the interpretation that the Elder Creek fault system represents tear faults in the upper plate of the Coast Range thrust (Jones and Irwin, 1971), along which the Klamath Mountains terrane moved upward and westward over the Coast Ranges province (fig. 2).

SOUTHEAST EXTENSION OF WILLOWS FAULT

Location of the Willows fault southeast of Sutter Buttes is not closely constrained by sparse well data along the eastern margin of the valley. No seismic-reflection profiles, that we are aware of, extend far enough east to cross the fault. However, wells in three widely spaced areas do extend far enough east from the deeper parts of the valley to cross the fault, and they provide the most direct evidence for the southeastern extension of the Willows fault. The first area we investigated extends southwest from near Marysville and passes through the Tisdale gas field (pl. 1, map A) just southeast of Sutter Buttes. A cross section (fig. 7) through wells in this area shows that the slope of the basement is significantly steeper east of the Willows fault (between wells 4 and 5, fig. 7), and we infer about 165 m of east-side-up displacement on the basement surface. Shale of the Eocene Capay Formation is recognizable in all of the electric logs in figure 7, and that unit shows about 100 m of displacement between wells 4 and 5. The base of continental deposits (nonmarine sedimentary rocks), which coincides generally with the base of fresh water in the valley, is difficult to identify in all the well logs of this section, but it does not appear to be offset between wells 3 and 5. The dip of the fault is not constrained in this section, and we have shown a steeply east-dipping reverse fault between wells 4 and 5, primarily because the pattern of offset in that area is so similar to that on the Willows fault to the northwest.

About 23 km southeast of the Tisdale gas field, a second line of wells extends from the Sacramento River near Knight's Landing northeastward toward Lincoln. Interest was directed initially to this line of wells because the outline of the Markley canyon, shown by Almgren (1978), is sharply deflected to the south in this area. A section through this line of wells is shown in figure 8. All of the sedimentary rocks in this part of the valley, except those in the Markley canyon, are offset up to the east between wells 3 and 4 (fig. 8). The amount of displacement appears to increase downward in the section from about 30 m at the base of the continental deposits to about 45 m in sandstone and shale sequences in the Upper Cretaceous rocks. Displacement of the basement is inferred to be about 150 m, but the depth to basement west of the Willows fault is not closely constrained by available well data. The most interesting aspect of the section is the marked asymmetry of the Markley canyon fill. Canyon development may have shifted progressively westward, to be localized immediately east of this part of the Willows fault, where the deepest part of the asymmetric canyon fill is located. This pattern of erosion and sedimentation suggests that late Oligocene and early Miocene movement on the Willows fault directly influenced the location and form of the Markley canyon.

The third line of wells we investigated extends from the Sacramento River near Clarksburg to the area east of Florin, where a series of low hills underlain by Pliocene continental deposits of the Laguna Formation rise above the surrounding Pleistocene alluvial deposits. A section through these wells is shown in figure 9. The electric log of the D. D. Feldman "Unit Plan No.1" well (sec. 8, T. 7 N., R. 6 E.); (No. 9, fig. 9) shows an obvious repeat in the Cretaceous rocks just above the basement. We interpret this repeat to represent east-side-up displacement on the Willows fault, which was intersected by the well at an elevation of – 1,277 m. If this interpretation is correct, the basement surface is offset about 120 m. A similar offset is shown by Upper Cretaceous rocks.

West of the Willows fault (fig. 9), a steeply east-dipping normal fault, previously unreported, offsets the Capay Formation and older rocks down to the east about 153 m. This normal fault does not appear to offset the base of post-Eocene and younger nonmarine rocks, nor are these rocks offset by the Willows fault to the east. Offset of the basement between wells 5 and 6 (fig. 9) appears to be east side up and suggests early reverse movement on the east-dipping fault.

The location of the Willows fault southeast of Florin is

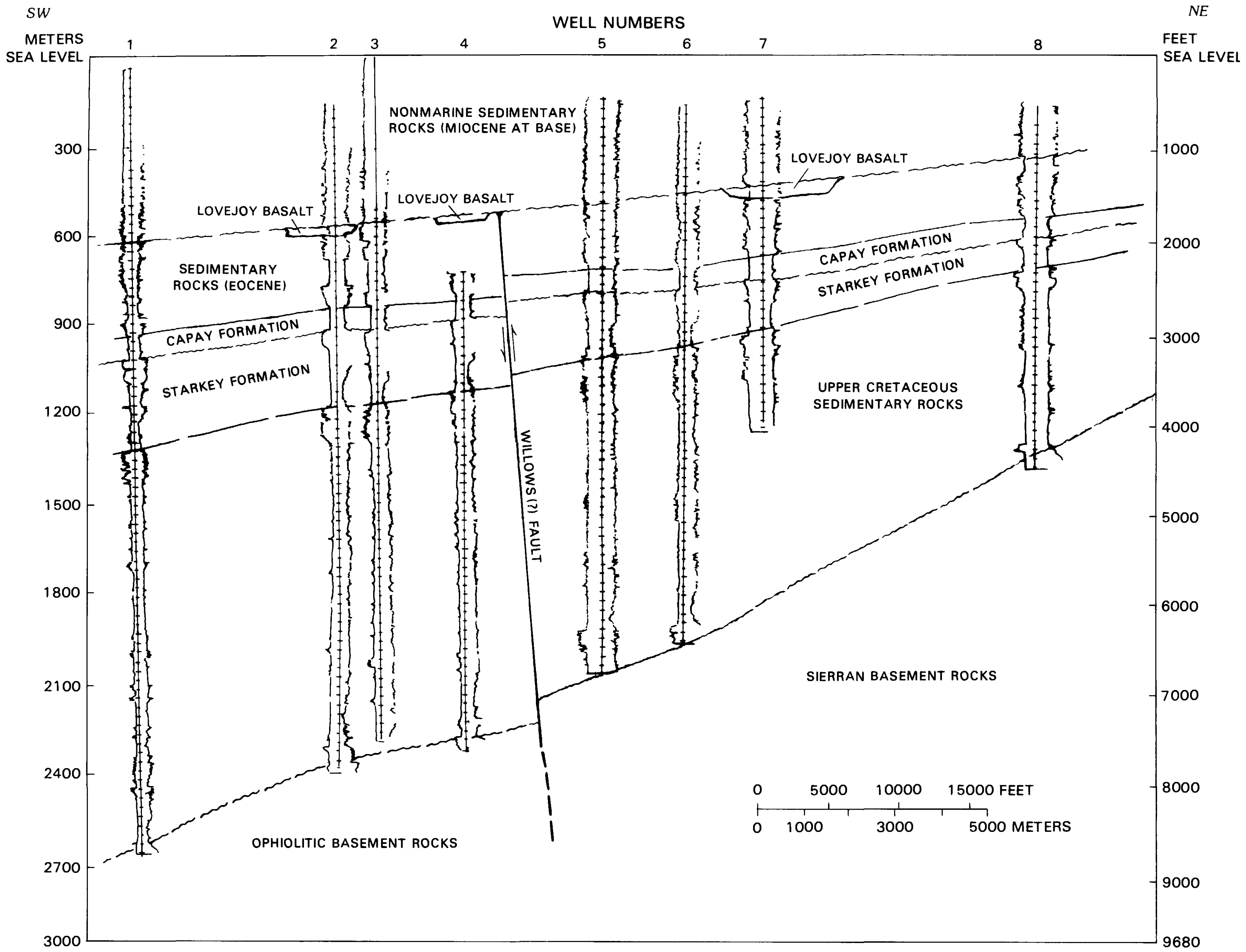


FIGURE 7.—Cross section showing electric logs and stratigraphic and structural relations across possible southeast extension of Willows fault near Tisdale gas field southeast of Sutter Buttes, Calif. Names of wells listed in appendix. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological Survey.

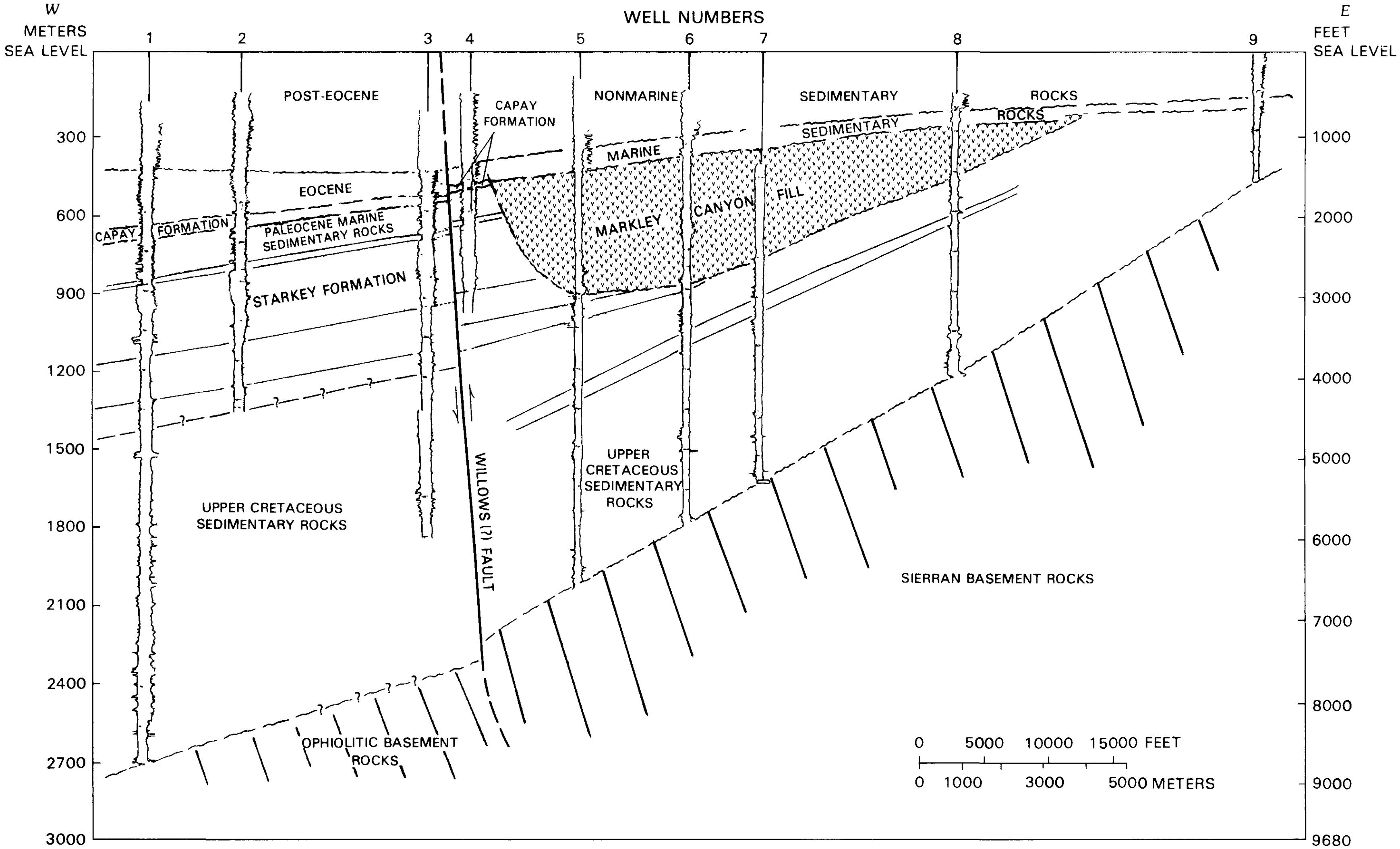


FIGURE 8.—Cross section showing electric logs and stratigraphic and structural relations across possible southeast extension of Willows fault near Catlett-Nicolaus gas fields, Calif. Names of wells listed in appendix. Note asymmetry of Markley canyon fill (patterned) and absence of offset in post-Eocene continental deposits. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological Survey.

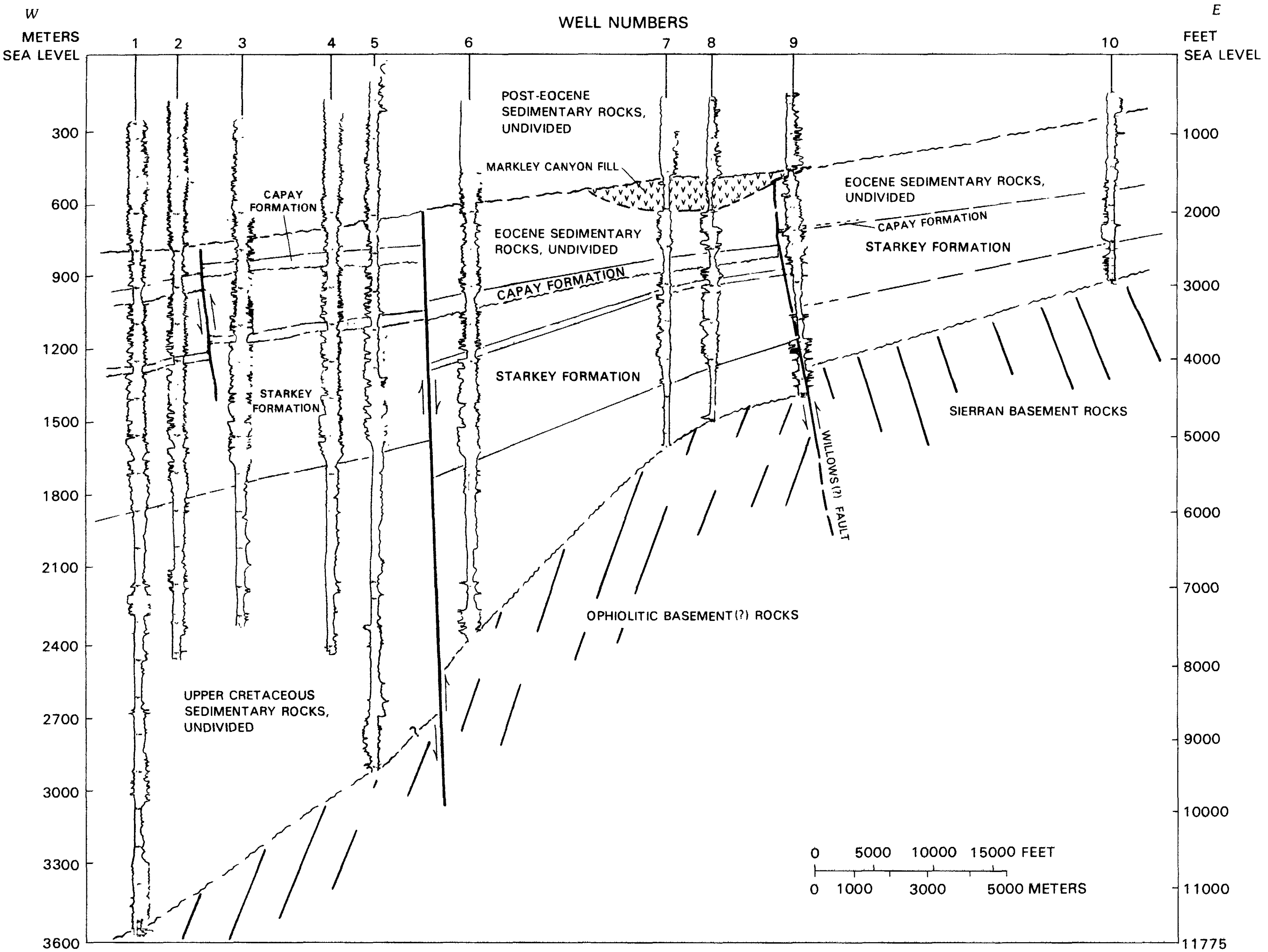


FIGURE 9.—Cross section showing electric logs and stratigraphic and structural relations across possible southeast extension of Willows fault southeast of Florin, Calif. Names of wells listed in appendix. Note repeated electric-log pattern near bottom of well 9 and apparent early east-side-up displacement of basement by unnamed fault between wells 5 and 6, followed by east-side-down offset of Cretaceous and early Tertiary rocks. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological

uncertain. A section through several wells east and west of the Lodi gas field, shown in figure 10, indicates a minor deflection in the Capay Formation between wells 6 and 7, but the underlying Cretaceous rocks and the basement surface are not significantly offset in that area. The general east-side-up flexure of the Capay between wells 6 and 7 is the type of deformation that would be expected adjacent to the Willows fault, but the absence of offset in underlying units argues against the existence of the fault. Deformation expressed as fault movement north of Lodi may be expressed as gentle warping to the south.

STRUCTURES NORTHEAST OF THE WILLOWS FAULT

The Willows fault diagonally transects the Sacramento Valley from northwest to southeast and divides the region into two late Cenozoic structural provinces. To the northeast, on the upthrown block of the Willows fault, the present axis of the Sacramento Valley is underlain by the Los Molinos and Glenn synclines (pl. 1). These narrow synclines, coupled with the Corning domes to the west and the Chico monocline to the east, have tightly controlled the course of the Sacramento River and influenced alluvial deposition during the late Quaternary. This tectonic control of alluvial deposition has profoundly influenced land use, particularly in the northeastern part of the valley, where a narrow strip of cultivated land, shown near the top of figure 11, flanks the river and lies directly over the trace of the Los Molinos syncline. South of the constricting influence of the Corning domes and the Chico monocline, late Quaternary alluvial deposits spread laterally across much of the valley over the Glenn syncline, the axial trace of which controls the present course of the Sacramento River as far south as Glenn, Calif. (pl. 1). From Glenn to Colusa, the course of the river lies along the trace of the Willows fault.

From the vicinity of Willows southeast to the Sacramento Delta, the downthrown southwest block of the Willows fault is capped by Holocene basin deposits that flooded over the broad Zamora syncline (pl. 1), burying older deposits, to rest unconformably on Upper Cretaceous rocks north of the Dunnigan Hills. The following part of the report discusses specific structural features on the northeast block of the Willows fault.

SUTTER BUTTES

Sutter Buttes (fig. 11) is a prominent set of hills composed of late Cenozoic volcanic rocks that rise about 635 m above the floor of the Sacramento Valley 15 km northwest of Marysville. According to Williams and Curtis (1977), volcanism and accompanying deformation of the surrounding sedimentary rocks occurred between 2.4 and 1.4 m.y. ago and consisted of two phases. During the early phase of magma injection, Upper Cretaceous and Tertiary rocks were arched into a dome 13 km across, broken by normal and high-angle reverse faults, and rapidly eroded so that the Tertiary rocks were stripped from the core of the dome before the explosive phase of volcanism. Explosive volcanism produced the rampart beds of tuff and tuff breccia that form the peripheral deposits of the buttes (fig. 11).

Extensive drilling for natural gas located an elongate dome in the subsurface about 10 km west of Sutter Buttes. Williams and Curtis (1977) referred to this structure as the buried Colusa buttes, but because it lacks surface expression, we refer to it as the buried Colusa dome (fig. 12). In the buried Colusa dome, several wells (fig. 12) penetrated volcanic rocks similar in composition and age to those in the Sutter Buttes. These data led Williams and Curtis (1977) to conclude that the buried Colusa dome formed solely by forceful magma injection contemporaneous with magmatism at Sutter Buttes. We suggest, however, that magmatism may have been localized by movement on the Willows fault and possibly by movement on the Mesozoic tectonic boundary between ophiolitic and Sierran basement terranes which passes beneath the eastern margin of Sutter Buttes (fig. 12).

The orientation and movement patterns of structures at Sutter Buttes and the buried Colusa dome suggest that deformation occurred in a regional east-west compressive stress field. The distribution of structural features at Sutter Buttes is distinctively asymmetrical. Beds in the Upper Cretaceous rocks on the south and southeast sides of the buttes dip steeply away from the core. Locally, they are overturned. Beds on the north and northwest sides of the buttes generally dip gently away from the core. As shown in figure 12, folds and faults are more abundant on the eastern half of the buttes and the wavelength of folds is tighter there than on the west. In addition to the small normal faults that surround the core of the dome, the eastern half of the dome is cut by a prominent set of steeply inward-dipping, arcuate high-angle reverse faults. This pattern of structural features indicates that the core of Sutter Buttes was thrust eastward as it moved upward.

The northward elongation of the buried Colusa dome strongly suggests that it too was formed in an east-west compressive stress regime. In fact, the amplitude, wavelength, and axial extent of the buried Colusa dome are remarkably similar to folds that occur just east of the Willows fault farther north (pl. 1; fig. 3). Although Redwine (1972) projected the Willows fault southeastward across the buried Colusa dome, our analysis of subsurface data suggests that a fault, possibly the Willows fault or a splay fault related to it, may lie just west of the buried Colusa dome. We conclude that the buried Colusa dome formed partly by east-side-up drag on a high-angle reverse fault along its western margin and partly by magma in-

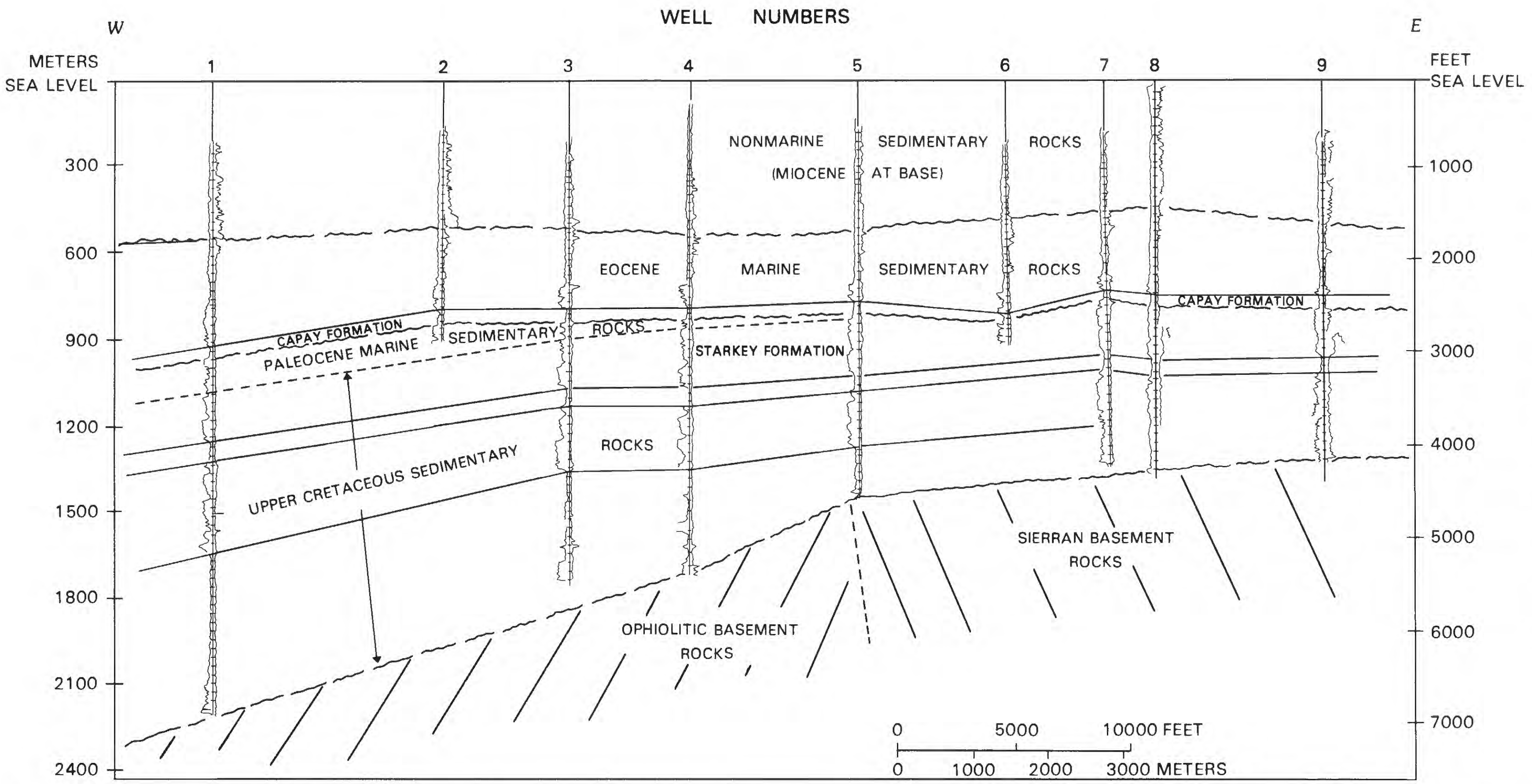


FIGURE 10.—Cross section showing electric logs and stratigraphic and structural relations across possible southeast extension of Willows fault (approximately follows trace of well 5) in area of the Lodi and Galt gas fields, Calif. Names of wells listed in appendix. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological Survey.

jection that was localized by movement on that fault.

The pattern of radial normal faults around the core of Sutter Buttes, flanked by reverse faulting to the east and west, assuming that the Willows fault was involved in the tectonism, is the type of deformation found by Withjack and Scheiner (1982) in their modelling experiments of doming in a compressive stress field. Although it is difficult to unequivocally separate magmatic doming from regional compressive deformation in the Sutter Buttes area, supporting evidence for the regional east-west compressive stress field is found in the Chico monocline about 40 km to the north.

FIGURE 11.—U.S. Air Force high-altitude oblique aerial photograph showing northern Sacramento Valley viewed to the north from vicinity of Sutter Buttes (center foreground), a late Pliocene and early Pleistocene volcanic center, approximately 16 km in diameter. Dark core of buttes composed primarily of andesite surrounded by lighter colored volcaniclastic deposits that form peripheral rampart beds of Williams and Curtis (1977). Narrow strip of cultivated land (dark) along Sacramento River (SR) between Corning domes (CD) and Chico monocline (CM) is surface expression of Los Molinos syncline (pl. 1). Oroville dam (OD) located near north end of Cleveland Hills faults where ground rupture occurred during August 1, 1975, Oroville earthquake.

CHICO MONOCLINE

The Chico monocline (fig. 13) is a northwest-trending, southwest-facing flexure that bounds the northeast side of the Sacramento Valley between Chico and Red Bluff. East of the monocline, Pliocene volcanic rocks of the Tuscan Formation dip less than 5° SW., but bedding steepens to 20° or more along the monoclinal flexure where the Tuscan dips beneath Quaternary deposits of the valley. The trace of the monocline is characterized by a complex surface pattern of anastomosing fault strands

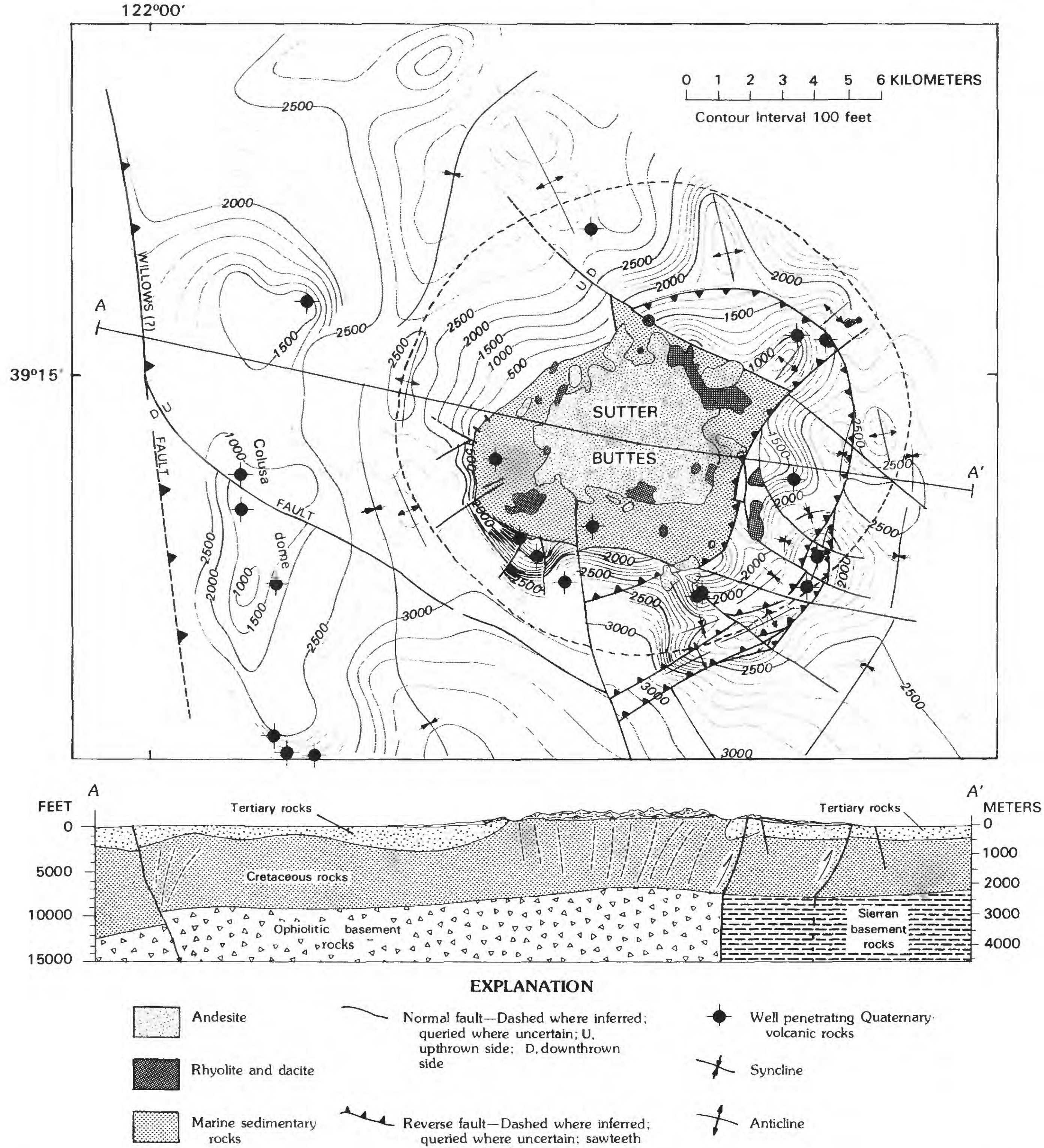


FIGURE 12.—Structure contour map of area around Sutter Buttes, modified from Williams and Curtis (1977, fig. 11) with data from Redwine (1972) and from our examination of electric logs in area of buried Colusa dome. Structure contours in feet; datum is top of the Kione sand of local usage (Williams and Curtis, 1977). Dashed circle on map marks approximate outer limit of rampart beds on valley floor; thin dashed lines on section represent schematic paths of magma injection.

FIGURE 13.—Composite of two U.S. Air Force high-altitude oblique aerial photographs showing pronounced linear trace of Chico monocline (CM, upper right to lower left), broad alluvial fan west of Chico monocline capped by Red Bluff Formation (RB, gray, uncultivated grazing land), and dissected topography of Tuscan Formation (TF) to east. DC, Deer Creek; CD, north Corning dome. Sacramento River (SR) follows trace of Los Molinos syncline (LMS) between north Corning dome and alluvial fan to east.

that show both west- and east-side-down displacements of small magnitude (Harwood and others, 1981).

Structure contours drawn on top of the Cretaceous rocks in the vicinity of the monocline indicate clearly that the Cretaceous strata are flexed and faulted by a major northwest-trending fault at depth beneath the surface trace of the monocline (pl. 1). The best control on the amount of displacement on the master fault beneath the Chico monocline comes from a line of wells that extends generally westward from the Exxon "C. C. Baccala No. 1" well (No. 4, fig. 14), which was drilled just west of the monocline about 18 km northwest of Chico. Although the Baccala well did not reach basement, it bottomed in Upper Cretaceous conglomerate at a depth of −1,700 m and provides a minimum elevation on the basement immediately west of the monocline when used in conjunction with the Pacific Western "Cana No. 1" well (No. 2, fig. 14) that reached serpentine at a depth of −2,100 m about 9 km to the west. Although no wells have been drilled east of the monocline, some control on the elevation and dip of the unconformities at the top and base of the Upper Cretaceous rocks is provided by exposures of these surfaces in several deep canyons to the east. When the trend of the basement surface at these exposures is projected westward, to pass beneath the bottom of the Baccala well, a minimum east-side-up offset of 365 m in the basement rocks is indicated on the fault beneath the monocline (fig. 14). The dip of this major fault cannot be determined directly from surface exposures or subsurface well data, but we show it as a steeply east-dipping reverse fault, because gravity and magnetic data (Griscom, 1973; Cady, 1975; Roberts and others, 1981) indicate that the relatively dense and magnetic basement beneath the northern part of valley extends eastward beneath the Sierran basement exposed east of the monocline. In the Sierran foothills southeast of Oroville, steep east dips are also commonly found in surface exposures of faults that may be the southeastward continuation of the fault beneath the monocline.

From available geophysical evidence, it appears that the fault beneath the monocline is a major tectonic boundary, with a long and complex tectonic history, along which the Sierran basement to the east was juxtaposed against highly magnetic, dense ophiolitic basement to the west (Cady, 1975; Griscom, 1973). These basement terranes were tectonically juxtaposed prior to deposition of the Upper Cretaceous strata. If this interpretation is correct, the fault beneath the Chico monocline may represent a part of the upper-slope discontinuity of Karig and Sharman (1975) that marks the boundary between basement of the late Mesozoic forearc basin to the west and that of the magmatic arc complex to the east. The Chico monocline, in that case, would be a structure inherited from the upper-slope discontinuity, possibly when that major tectonic break was reactivated by late Cenozoic subduction north of the migrating Mendocino triple junction.

The Chico monocline is clearly a late Cenozoic tectonic feature. It formed after deposition of the Ishi Tuff Member of the Tuscan Formation, about 2.6 m.y. ago and prior to the eruption of the olivine basalt of Deer Creek 1.09 m.y. ago (Harwood and others, 1981). Late Cenozoic displacement on the Chico monocline fault has been predominantly east side up, with an apparent component of left-lateral movement that may have contributed to the formation of the Salt Creek, Tuscan Springs and Sevenmile domes at the north end of the monocline and possibly influenced the northeast-trending Inks Creek fold system and Battle Creek fault zone farther northwest.

There is some indication that the Chico monocline fault may be active. The olivine basalt of Deer Creek is offset, but the amount of offset is small. Similarly, a few faults along the monocline show scarps about 1 m in height in the surface of the Tuscan Formation (Harwood and others, 1981). These observations indicate that movement has taken place on the monocline fault system within the past million years. In their study of regional seismicity after the Oroville earthquake, Marks and Lindh (1978)

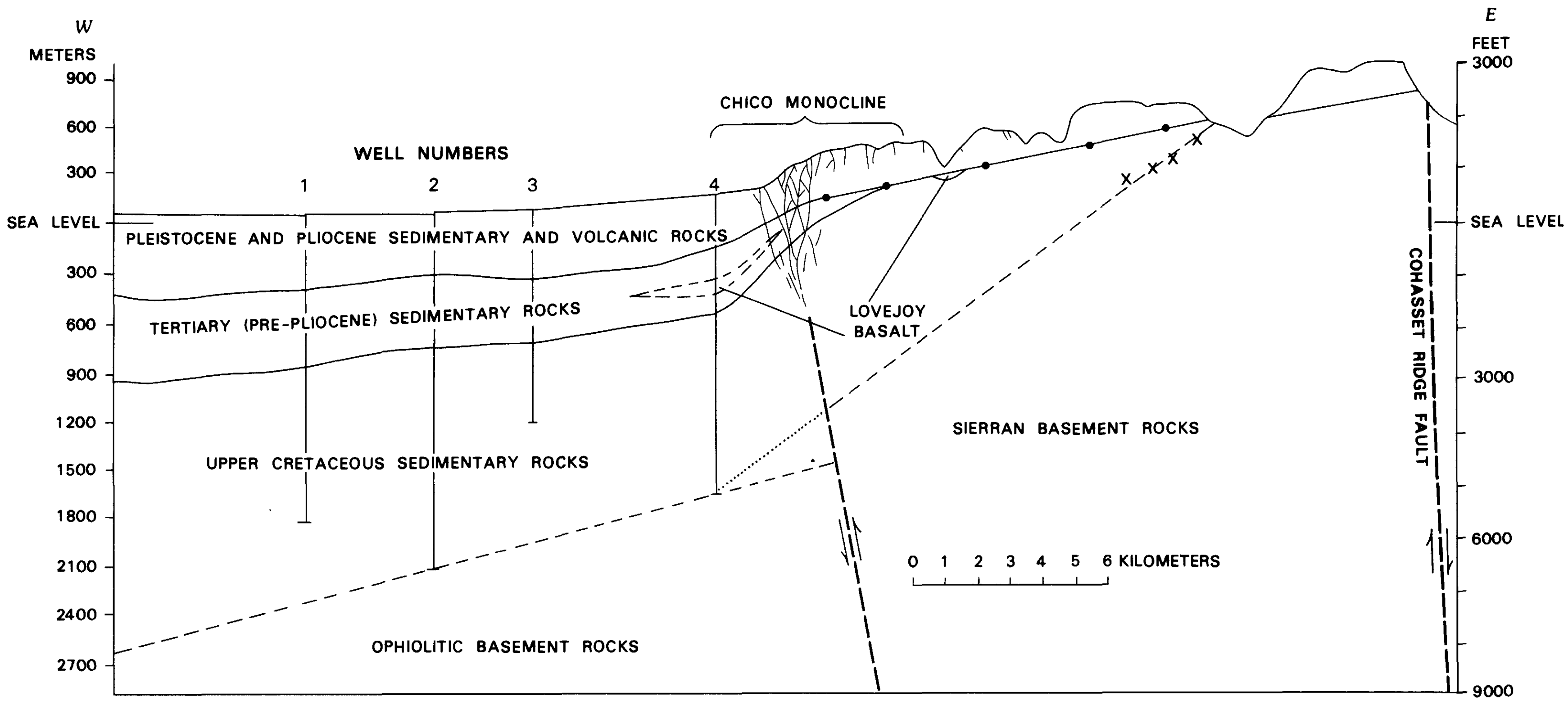


FIGURE 14.—Cross section across Chico monocline about half way between Chico and Red Bluff, Calif., showing offset of basement terranes. Dots indicate unconformity at base of Tuscan Formation, as projected into line of section; x's indicate unconformity at base of Upper Cretaceous rocks, as projected into line of section. Names of wells listed in appendix.

identified two events west of the Oroville aftershock zone at a depth of 40 km, significantly below the main cluster of aftershocks associated with the Oroville earthquake. We interpret those deep events to have occurred on the

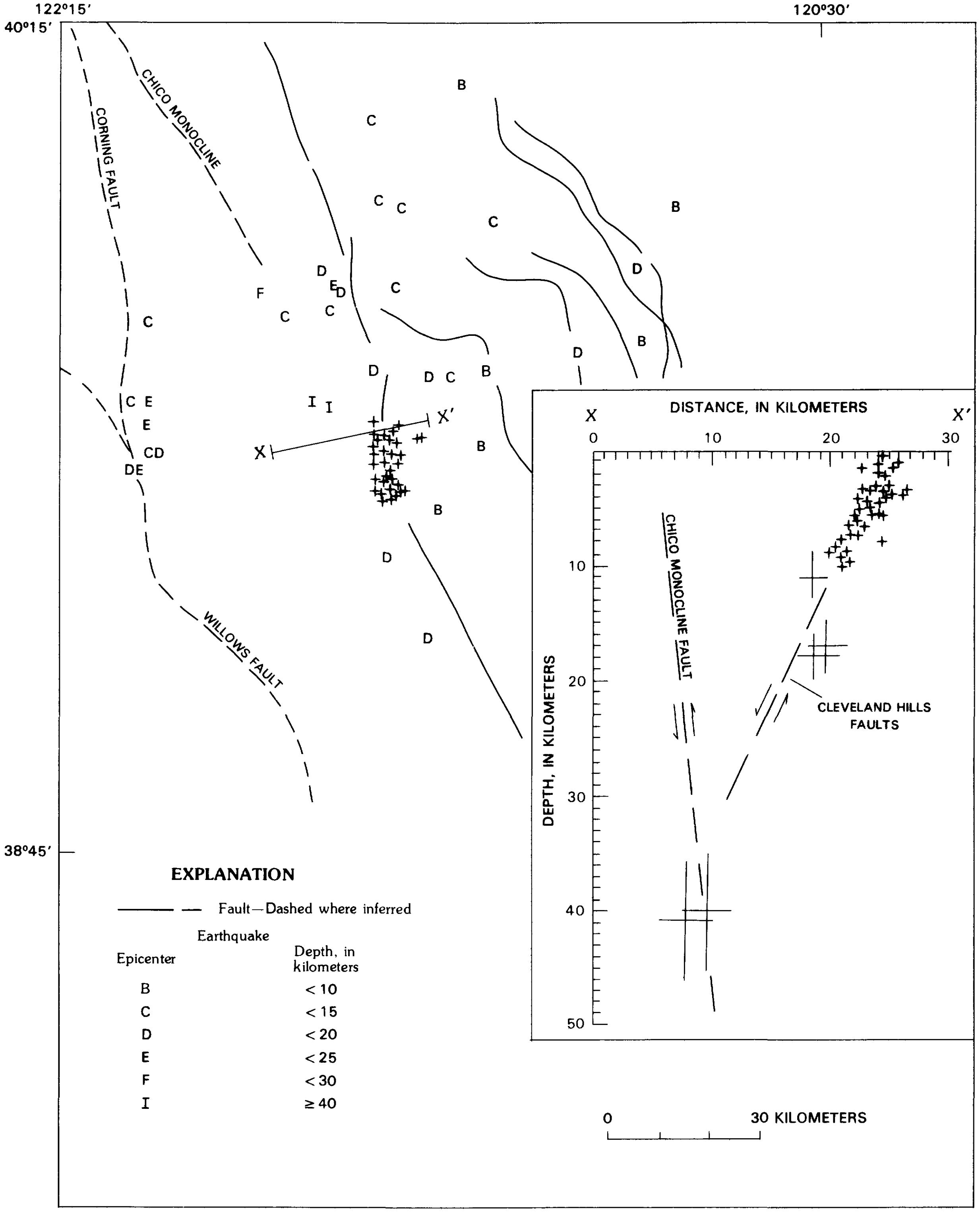


FIGURE 15.—Map and section showing locations of Oroville earthquake aftershocks (crosses on section) in relation to Cleveland Hills faults and Chico monocline (modified from Marks and Lindh, 1978). Earthquakes labeled I on map (largest crosses on section) at a depth of about 40 km are interpreted to have occurred on Chico monocline fault.

basement fault beneath the Chico monocline and not on the Cleveland Hills faults, where most of the aftershock events were located (fig.15). The steeply east-dipping normal fault, shown east of the Chico monocline on figure 15, is one of several short, north- and north-northwest-trending normal faults that extend along the northern foothills of the Sierra Nevada (pl. 1). This normal fault is probably part of the fault system on which the August 1, 1975, Oroville earthquake occurred.

BATTLE CREEK FAULT ZONE

The Battle Creek fault zone strikes east-northeast across the Sacramento Valley between Red Bluff and Redding, nearly at right angles to the trend of the Chico monocline. East of the Sacramento River, the Battle Creek fault zone is marked by a pronounced south-facing escarpment, shown in figure 16, that extends from the river northeastward toward Lassen Peak for a distance of 32 km. Fault strands within this part of the fault zone dip steeply southeast and show predominantly south-side-down, normal fault movement (Harwood and others, 1980; Helley and others, 1981). Vertical displacement on the fault zone increases from about 45 m just east of the Sacramento River to 330 m at Black Butte (fig. 1) and to about 440 m north of Manton. A small component of right-lateral strike-slip movement is suggested by fractures on some of the fault strands, but the exact amount of lateral displacement is not known.

West of the Sacramento River, the valleys of Cottonwood Creek and its south fork are probably controlled in part by the Battle Creek fault system, but modern stream activity and agricultural practices obscure any young traces of the faults east of the South Fork of Cottonwood Creek (pl. 1). At the South Fork of Cottonwood Creek and along Red Bank Creek, late Quaternary terraces show evidence of young faulting (Helley and others, 1981). To the west-southwest, the faults in the terrace deposits merge into previously mapped faults in the Red Bluff and Tehama Formations, and, on strike, they appear to merge with tear faults in the upper plate of the Coast Range thrust mapped by Bailey and Jones (1973).

FIGURE 16.—Photograph of Battle Creek fault scarp (BC) capped by basaltic cinder cone of Black Butte. View to northeast along Highway 36 and across rolling grassy plains underlain by volcanic fanglomerate correlated with the Red Bluff Formation.

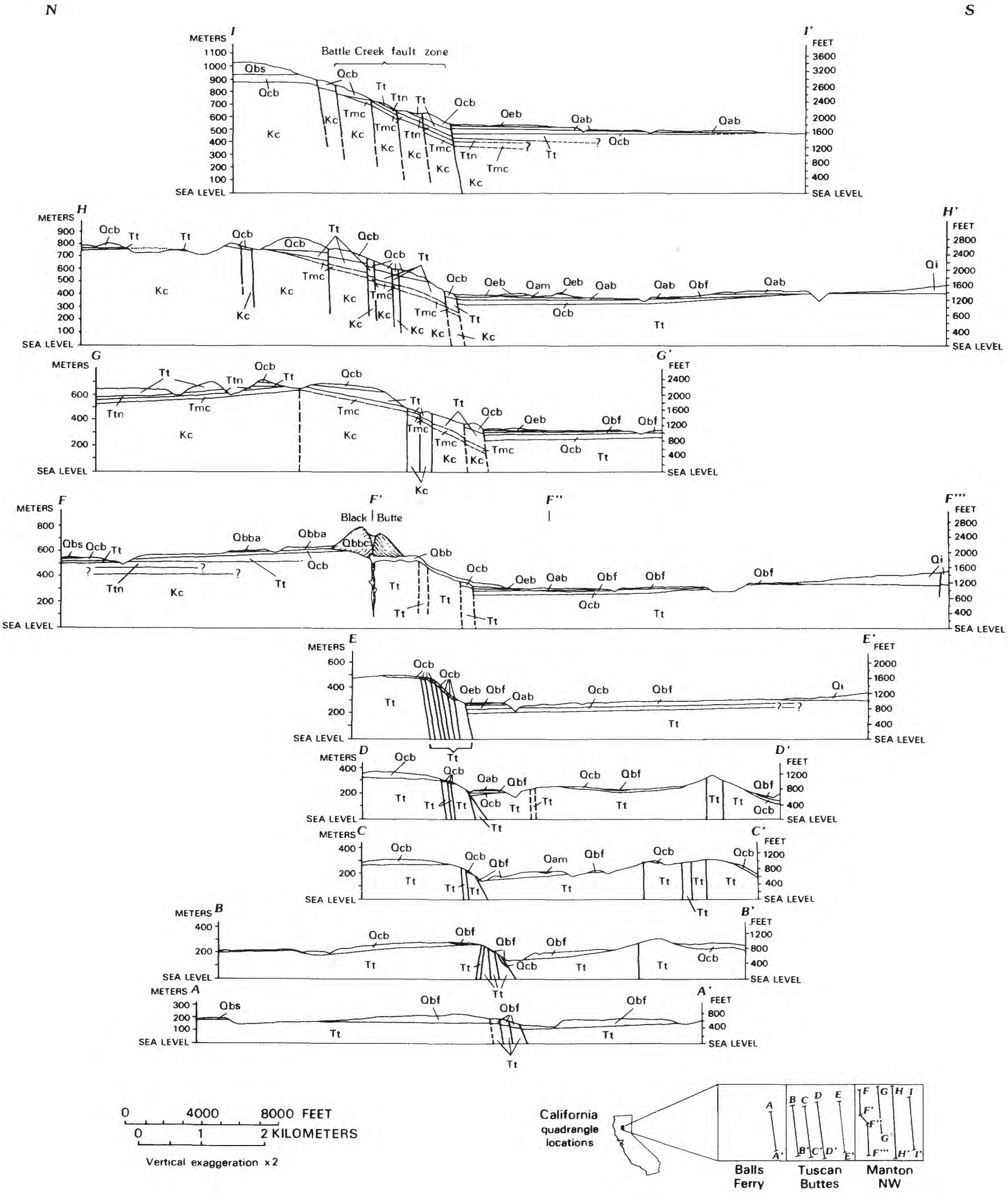
The orientation and distribution of faults in the Battle Creek fault zone east of the Sacramento River are shown in figure 17. Section *A-A′* is located 2.4 km east of the Sacramento River and shows relatively little offset of the fanglomerate that caps the Tuscan Formation. As the fault zone extends eastward, offset on the various Pliocene and Pleistocene units and the height of the fault scarp increases. From near the Sacramento River east to Black Butte (sections *A-A′* through *F-F′*, fig. 17), the fault strands are closely grouped along the fault scarp, but east of Black Butte (sections *G-G′* through *I-I′*, fig. 17) the fault strands are more widely spaced. In map view, the eastern fault strands veer northwestward and appear to die out a short distance north of the escarpment (Harwood and others, 1980).

Several clusters of small-magnitude earthquakes were reported along the eastern third of the Battle Creek fault zone, generally east of Black Butte (Bolt, 1979). Since 1980 the California seismic net of the U.S. Geological Survey has recorded several small-magnitude earthquakes (J. Eaton, oral commun., 1982) that lie approximately along the Battle Creek fault zone and extend west nearly to the Sacramento River.

Coarse volcanic fanglomerate, correlated here with the Red Bluff Formation by soil stratigraphy and geomorphic surface, is offset by the Battle Creek fault zone east of the Sacramento River and dates the major movement as younger than 1.09 m.y. The Rockland ash bed (referred to as the ash of Mount Maidu by Helley and others, 1981) appears to have been partly channeled by the Battle Creek fault scarp, suggesting that the faulting occurred prior to about 0.45 m.y. ago, the age of the Rockland ash bed (Meyer and others, 1980).

INKS CREEK FOLD SYSTEM

A set of northeast-trending folds, referred to as the Inks Creek fold system by Helley and others (1981), deforms the rocks south of the Battle Creek fault zone and structurally controls the major loops in the Sacramento River at Jelly School and the nearby Table Mountain (pl. 1; figs. 1, 18). The axial trace of the major syncline in the fold set passes through Table Mountain and extends northeastward, nearly coincident with Inks Creek. The complementary anticline to the north arches the volcanic fanglomerate, which is correlated with the Red Bluff Formation, and the underlying basalt of Coleman Forebay (Helley and others, 1981), exposing strata of the uppermost part of the Tuscan Formation northeast of Jelly


N
S
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METERS
FEET
SEA LEVEL
Battle Creek fault zone
Qbs
Qcb
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Qeb
Qab
H
H'
Qam
Qbf
Qi
G
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F
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Black Butte
Qbba
Qbbc
Qbb
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D
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C
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0 4000 8000 FEET
0 1 2 KILOMETERS
Vertical exaggeration x2
California quadrangle locations
Balls Ferry
Tuscan Buttes
Manton NW

School. Axial surfaces of the folds are vertical, and fold axes plunge 20°–35° SW. Axial traces of the folds change strike to the northeast so that the anticlinal trace merges with the trend of the Battle Creek fault zone and apparently dies out along the fault to the east-northeast.

West of the Sacramento River, the Inks Creek fold system is expressed as a broad area of uplift known as the Hooker dome (fig. 18). This structure, which has profoundly influenced drainage patterns in the area, particularly along Hooker Creek and Blue Tent Creek (fig. 18), is underlain by the Tehama Formation capped by a few scattered erosional remnants of the Red Bluff Formation.

The Inks Creek fold system is clearly reflected in the structure contours drawn on top of the Cretaceous rocks, which show a closure of about 485 m over the anticline (pl. 1). Although only a few wells in this area reach basement, the elevation of the basement in two wells drilled on the anticline of the Inks Creek fold system is higher than the total depth of some wells to the west that bottomed in Upper Cretaceous rocks. These data indicate that the basement was deformed, along with the cover rocks, during formation of the Inks Creek fold system. They also strongly suggest that the folds shown by the structure contours on plate 1 involve the basement of the valley rather than deformation on a decollement or a series of detachment thrusts in the cover rocks.

The fact that the anticlinal trace of the Inks Creek fold system merges with the Battle Creek fault zone suggests a genetic relation between these major structures. If this interpretation is correct, evidence for the age of the Inks Creek fold system provides data for the age of movement on the Battle Creek fault zone. At Round Mountain, about 3 km southwest of Jelly School, the Rockland ash bed rests unconformably on the eroded western contact of the Red Bluff Formation at an elevation of 160 m. We assume that the ash was deposited in an ancestral channel of the Sacramento River, which apparently formed a large westward bend at that time extending from Bloody Island on the north to the vicinity of Bend on the south (fig. 18). Uplift and erosion of the Red Bluff Formation, which was deposited between 0.45 and 1.09 m.y. ago, had begun

EXPLANATION

Qi	BASALTIC ROCKS OF INSKIP HILL VOLCANIC CENTER (QUATERNARY)--Undifferentiated basalt of Inskip Hill
	BASALTIC ROCKS OF BLACK BUTTE VOLCANIC CENTER (QUATERNARY)--Consist of:
Qbbc	Cinder-cone deposits
Qbba	Cinder-blanket deposits
Qbb	Basalt flow of Black Butte
Qbs	BASALT OF SHINGLETOWN RIDGE (QUATERNARY)
Qeb	BASALT OF EAGLE CANYON (QUATERNARY)
Qab	HYPERSTHENE ANDESITE OF BROKEOFF MOUNTAIN (QUATERNARY)
Qam	ASH OF MOUNT MAIDU (QUATERNARY)--Equivalent to the Rockland ash bed of Sarna-Wojcicki and others(1985)
Qbf	ALLUVIAL FAN DEPOSITS OF BATTLE CREEK (QUATERNARY)
Qcb	BASALT OF COLEMAN FOREBAY (QUATERNARY)
Tt	TUSCAN FORMATION, UNDIVIDED (PLIOCENE)--Locally divided into:
Ttn	Nomlaki Tuff Member
Tmc	MONTGOMERY CREEK FORMATION (EOCENE)
Kc	CHICO FORMATION (UPPER CRETACEOUS)

FIGURE 17.—Geologic cross sections showing statigraphic and structural relations along Battle Creek fault zone. Note eastward increase in height of fault scarp and amount of offset of geologic units from section A-A′ to section I-I′. Geologic units from Helley and others (1981).

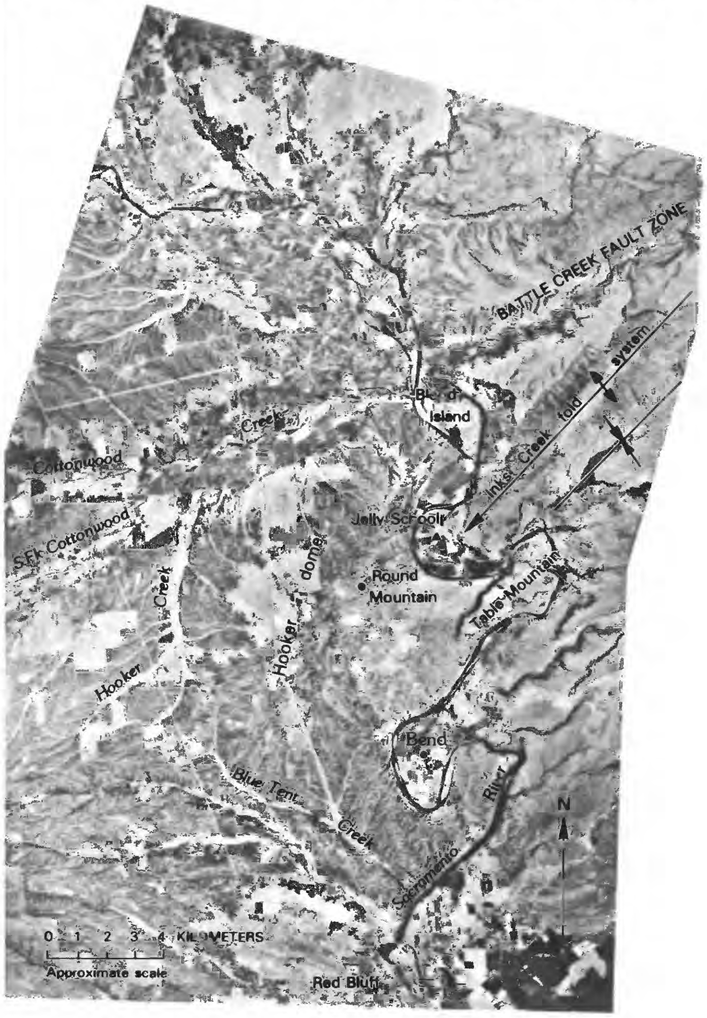


FIGURE 18.—Composite of two U.S. Air Force high-altitude vertical aerial photographs showing topographic expression of Inks Creek fold system (axial traces), Hooker dome, and Battle Creek fault zone.

prior to deposition of the ash. In contrast to the occurrence of the ash at Round Mountain, alluvial deposits of the lower part of the Riverbank Formation are found at an elevation of 120 m in this region, and they flank the present sinuous course of the Sacramento River through the structurally controlled loops of the river at Jelly School and Table Mountain. Clearly, the Sacramento River was forced into its tortuous course around the Inks Creek fold system by early Riverbank time. We conclude, therefore, that the Inks Creek fold system and at least some of the displacement on the Battle Creek fault zone developed in the timespan between 0.45 m.y., the age of the ash, and about 0.4 m.y. ago, the age of the lower part of the Riverbank Formation (Marchand and Allwardt, 1981).

SALT CREEK, TUSCAN SPRINGS, AND SEVENMILE DOMES

Small areas of Upper Cretaceous marine rocks are exposed beneath the Pliocene Tuscan Formation in Salt Creek, at Tuscan Springs, and in Sevenmile Creek between the north end of the Chico monocline and the Inks Creek fold system (pl. 1). In each area, beds in the Tuscan dip outward, away from the Upper Cretaceous rocks, and define three small domes, shown on plate 1 as the Salt Creek, Tuscan Springs, and Sevenmile domes.

The long axes of the domes trend north-northeast and plunge 15°–25° NE. and SW. Cold sulfurous springs are present in the Salt Creek and Tuscan Springs domes, and olivine basalt rests unconformably on steeply northwest-dipping Upper Cretaceous rocks in the core of Sevenmile dome. The olivine basalt was extruded from a north-trending fissure after the Tuscan Formation was domed and erosion had exposed the Upper Cretaceous rocks.

The age of the olivine basalt is unknown, but it probably is a few tens of thousands of years old, at most, and is roughly coeval with similar basalt flows and cinder cones found at Inskip Hill (pl. 1) and Black Butte along the Battle Creek fault zone. The domes probably formed during the same phase of late Quaternary deformation about 0.4–0.45 m.y. ago that produced the Inks Creek fold system and the Battle Creek fault zone, and they are thus younger than the Chico monocline.

RED BLUFF FAULT

The Red Bluff fault is a subsurface structure that extends southwest of Red Bluff for at least 25 km. West of Red Bluff, there is no surface feature associated unequivocally with the fault, and its location on plate 1 is taken from Jennings' (1977) geologic map of California. Apparently Jennings included the fault on that map on the basis of seismic-reflection data obtained from private industry (Oliver and Griscom, 1980). We have not seen that data or any other seismic profiles that might cross the fault. Griscom (1973) inferred the existence of a major northeast-trending fault in the area from magnetic data.

East of Red Bluff, structure contours on top of the Cretaceous rocks outline a major anticline, the broad crest of which lies beneath the Salt Creek, Tuscan Springs, and Sevenmile domes. The axial trace of this anticline roughly coincides with the east-northeast projection of the Red Bluff fault, but there is no indication that the Tuscan Formation is faulted at the surface along the N. 70° E. trend. The southeast limb of the anticline may be faulted at depth, but surface and subsurface data are too sparse to either prove or refute that structural possibility. Between Tuscan Springs and the Humble "Cone Ranch No. 1" well (sec. 20, T. 27 N., R. 2 W.) to the south, the top of the Cretaceous sequence drops about 1,433 m over a distance of less than 8 km. The lower elevation is due to warping and faulting on the Chico monocline fault, as well as to folding and possible faulting along the northeast projection of the Red Bluff fault.

From our reinterpretation of a north-south section along the Sacramento River near Red Bluff given by Redwine (1972, section *F-F′*), the base of the Tehama Formation could be offset, down to the south, by as much as 141 m across the Red Bluff fault just east of Red Bluff. Because of the lack of northeast-trending surface faulting in the Tuscan to the east, however, it seems unlikely that all of that differential elevation is due to fault displacement; much of it may be related to folding over a fault at depth.

FAULTS SOUTH OF INSKIP HILL

The northwest-trending arcuate faults that are shown south of Inskip Hill (pl. 1) are a gross simplification of the faults and fractures found in that area. In detail, the Tuscan Formation is laced with a dense network of short anastomosing faults along which thin lines of shrubs and trees grow (fig. 19). The amount of offset on a single fault is small, generally a few meters or less, and there appears to be no significant cumulative offset in the Tuscan Formation across the fractured area.

The arcuate pattern of faults is concave toward the area of Mineral, to the east, where Wilson (1961) identified a major volcanic center that he called Mount Maidu. It seems reasonably certain that the fracture pattern is related to volcanic-tectonic activity in the Lassen area, but whether it is associated with Mount Maidu or some other volcanic center is uncertain.

STRUCTURES SOUTHWEST OF THE WILLOWS FAULT

With the exception of the buried Colusa dome, the subsurface structure of the valley southeast of the Willows fault to the Sacramento Delta is dominated by the Zamora syncline. South of buried Colusa dome, the Zamora syncline is asymmetrical and has a broad, gently west-dipping east limb that extends to the Sierran foothills and a steeply dipping west limb that is warped by southeast-plunging folds. North of the buried Colusa dome, the axial trace of the Zamora syncline intersects the Willows fault, and the top of the Upper Cretaceous rocks defines a generally east-dipping homocline.The following sections discuss structural features west of the axial trace of the Zamora syncline.

CAPAY VALLEY-CAPAY HILLS AREA

Cache Creek drains east from Clear Lake and enters the Sacramento Valley west of Woodland, through the tectonically controlled depression of Capay Valley. On the west, Capay Valley is flanked by the eastern slope of the Coast Ranges that contain a remnant of Eocene marine sandstone and shale of the Capay Formation (Redwine, 1972) resting unconformably on Upper Cretaceous marine rocks (Kirby, 1943a). Beds in the Capay and underlying rocks dip 25°-55° E. The Capay Hills, previously known as the Rumsey Hills (Kirby, 1943b), lie immediately east of Capay Valley and are underlain by the same Upper Cretaceous rocks exposed to the west. Here, however, the Cretaceous rocks lie in the core of a faulted southeast-plunging anticline, shown by structure contours on plate 1. They are unconformably overlain by nonmarine sandstone and shale of the Pliocene Tehama Formation. The Eocene Capay Formation is not present in the Capay Hills or in the Capay Valley, where scattered subsurface and surface data indicate that the Tehama rests unconformably on Upper Cretaceous rocks.

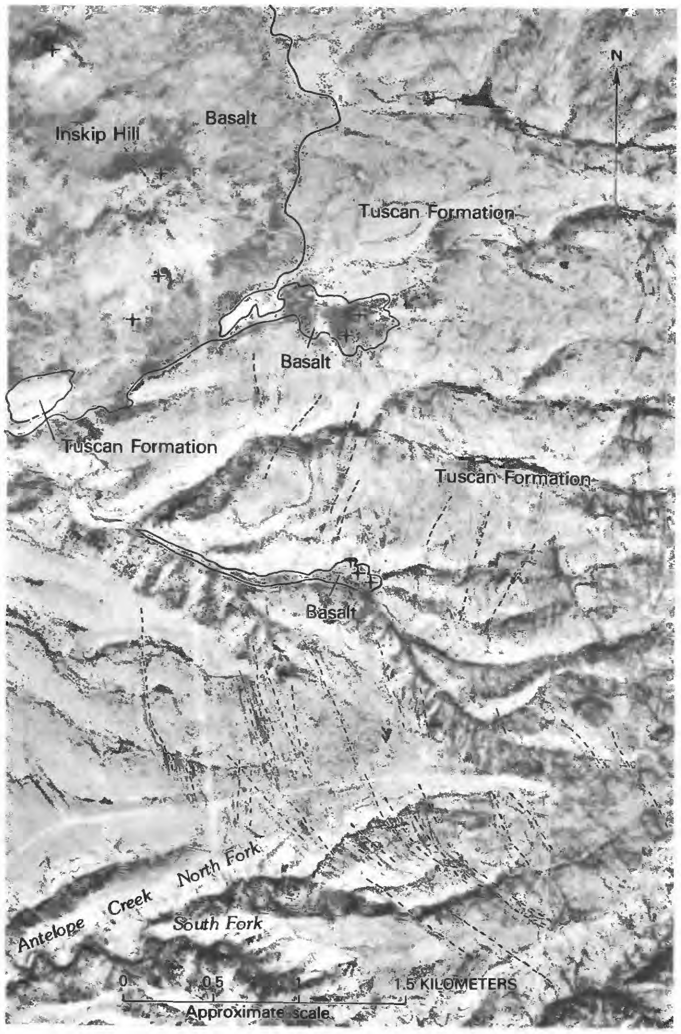


FIGURE 19.—Low-altitude vertical aerial photograph (U.S. Geological Survey) showing arcuate fault pattern (dashes) and cinder cones (crosses) south of Inskip Hill. Gray incised plains are underlain by the Tuscan Formation; thin dark lines are concentrations of brush and trees growing along fault traces.

On the floor of Capay Valley, the Tehama and older rocks have been eroded, and old stream channels have been filled with a variety of Quaternary alluvial deposits. Alluvial deposits of the upper part of the Modesto Formation indicate that Cache Creek entered the Sacramento Valley through a channel cut in the Tehama Formation near the west boundary of Capay Valley in late Wisconsin time. During the past 12,000 years the course of Cache Creek has shifted eastward so that the creek now exits Capay Valley through a sharp gorge cut into the relatively resistant Upper Cretaceous rocks at the southern end of Capay Hills. The shift in the course of Cache Creek and the surface and subsurface data that show faulting in the Tehama Formation provide strong evidence for late Cenozoic tectonism in Capay Valley.

Along the west flank of the Capay Hills, Kirby (1943b) mapped a northwest-trending, east-dipping thrust, the Sweitzer fault, and a lower ancilliary thrust, the Eisner fault; both faults place Upper Cretaceous rocks in contact with the Tehama. Wagner and Saucedo (1984) have reinterpreted the Sweitzer and Eisner faults as west-dipping normal faults. Their interpretation of the Sweitzer fault is supported by subsurface data at the south end of Capay Hills, where the contact between the Forbes and Guinda Formations of Kirby (1943a) is downfaulted about 50 m on the west (fig. 20, wells 3 and 4). Compared to other faults in the area, the Sweitzer fault appears to have played a relatively minor part in the structural evolution of the Capay Valley and Capay Hills.

West of the Sweitzer fault, the contact between the Forbes and Guinda Formations was reported at an elevation of about 219 m beneath the Capay Valley (fig. 20, well 2). That contact is displaced about 610 m up to the west along the northwest-trending, steeply east-dipping East Valley fault (fig. 20). Along the west margin of Capay Valley, the contact between Upper Cretaceous rocks and the Tehama Formation is down faulted about

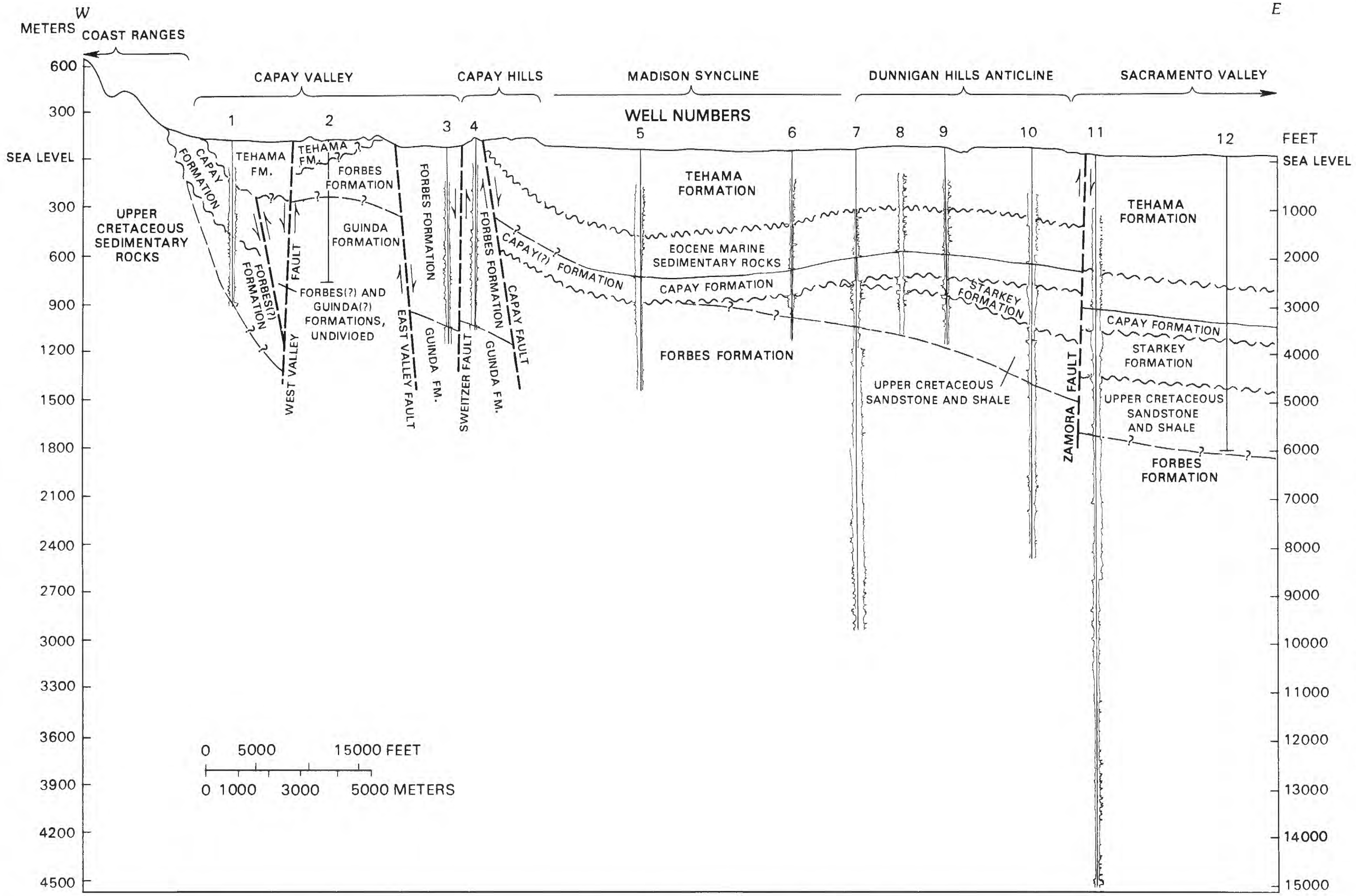


FIGURE 20.—Cross section showing electric logs and stratigraphic and structural relations in Capay Valley, Capay Hills, and Dunnigan Hills. Names of wells listed in appendix. Stratigraphic nomenclature used may not necessarily conform to that adopted by U.S. Geological Survey.

150 m to the west by the near-vertical West Valley fault (fig. 20). At the south end of Capay Valley, Upper Cretaceous rocks are thrust over the Eocene Capay Formation along a thrust or high-angle reverse fault that is exposed to the south along the east flank of the Coast Ranges (pl. 1). The location of this thrust or high-angle reverse fault is unknown east of the West Valley fault.

East of Capay Hills, the Capay Formation unconformably overlies east-dipping Upper Cretaceous rocks beneath the Madison syncline and the Dunnigan Hills anticline (pl. 1; fig. 20). The Capay Formation and overlying Eocene marine rocks are not exposed in the Capay Hills where the Pliocene Tehama Formation rests unconformably on Upper Cretaceous rocks. Because the Capay Formation is present west of Capay Valley, however, we assume that the Capay Formation once extended over the area of the Capay Hills but that it was eroded from the area after west-side-up movement on the Capay fault (fig. 20) and the other high-angle reverse faults to the west under Capay Valley. In this interpretation, the minimum displacement on the Capay fault would be about 700 m. Major west-side-up reverse movement occurred on the Capay fault, the East Valley fault, and the unnamed thrust or high-angle reverse fault to the west in post-Eocene time and before deposition of the Tehama Formation, which contains the 3.4-m.y. old Putah Tuff Member near its base (Sarna-Wojcicki, 1976). West-side-down movement occurred on the West Valley fault subsequent to deposition of the Tehama Formation, and east-side-down, post-Tehama displacement also may have occurred on the East Valley fault and the Capay fault.

DUNNIGAN HILLS ANTICLINE AND THE ZAMORA FAULT

Near Woodland, Bryan (1923) recognized accordant summit elevations capped by Red Bluff gravel on a series of northwest-trending dissected uplands that he called the Hungry Hollow Hills, but which are now known as the Dunnigan Hills. The northeast flank of the upland is bounded by a linear escarpment that Bryan called the Hungry Hollow fault, but which is referred to here as the Zamora fault (pl. 1; fig. 20). He recognized down-to-the-east displacement of at least 121 m at the north end of the fault scarp and offset of about 60 m near Cache Creek to the south (Bryan, 1923, p. 79).

Bryan's early work, combined with the topographic relief in the area, made the Dunnigan Hills a prime target for early seismic-reflection studies that resulted in the discovery of the Dunnigan Hills gas field in 1946 (Rofe, 1962). The gas-producing structure is a doubly plunging, northwest-trending anticline along which various Upper Cretaceous sandstone beds are unconformably capped by the Eocene Capay Formation (fig. 20). This major structure has topographic relief. Red Bluff gravel wraps around the northwest-plunging nose of the fold and occurs in scattered patches along the east flank, on the crest line, and at the southeast-plunging nose of the fold. Oat Creek (pl. 1) and Bird Creek to the southeast (not shown on pl. 1) are antecedent to the fold and change from southeast to northeast courses approximately at its axial trace.

Data from a recent well drilled near Zamora by the U.S. Geological Survey provide new information on the amount of late Cenozoic deformation in the area. A conspicuous volcanic ash bed, penetrated at a depth of −137 m (Page and Bertoldi, 1983), has been correlated tentatively by mineralogy and chemical composition of the glass with the Rockland ash bed by C. E. Meyer and A. M. Sarna-Wojcicki (oral commun., 1982). Because the ash occurs directly above Red Bluff gravels elsewhere in the valley (Harwood and others, 1981; Helley and others, 1981), it is assumed that the ash overlies the Red Bluff in the Zamora well and that there is a minimum of 220 m of vertical displacement of the Red Bluff Formation. This vertical separation is the result of folding on the Dunnigan Hills anticline and displacement on the Zamora fault (fig. 20).

MIDLAND FAULT

The Midland fault is a major subsurface structure that was discovered in the Sacramento Delta during development of the Rio Vista gas field between 1936 and 1943 (Frame, 1944). Through data from extensive drilling, the fault was extended about 25 km north of the Rio Vista field to the Maine Prairie gas field (Arleth, 1968). North of that field, the location of the Midland fault is uncertain. Redwine (1972) proposed that the Sweitzer fault, mapped by Kirby (1943b) in the Capay Hills, was the northwest continuation of the Midland fault. Although Jennings (1977) showed the Midland-Sweitzer fault connection, suitably queried, on his geologic map of California, that interpretation is no longer considered correct (C. Jennings, oral commun., 1981). The location of the Midland fault south of the Rio Vista gas field is uncertain.

In the Rio Vista gas field, the Midland fault is actually a north-trending, steeply west-dipping to vertical fault zone that offsets Paleocene and Eocene rocks down to the west in a series of fault blocks as shown in figure 21. Early in the development of the Rio Vista gas field, it became apparent that movement on the Midland fault had controlled local patterns of Tertiary sedimentation. Frame (1944) noted that the Capay Formation was significantly thicker west of the Midland fault, and he suggested that maximum movement occurred on the fault during or at the close of Capay deposition. Although the Capay Formation shows the greatest amount of syntectonic thickening across the fault, all of the Paleocene and Eocene units show some differential thickening west of the fault (Burroughs and others, 1968). In his detailed analysis of depositional and tectonic cycles in the southern Sacramento Valley, Almgren (1978) demonstrated approximately 610 m of episodic movement on the fault between the early Paleocene and early Oligocene. There is no indication that early Oligocene or younger deposits are offset by the Midland fault; therefore, it appears that the Midland fault was an active structural feature on the early Tertiary continental margin but that it has not been active in the late Cenozoic.

By combining data presented by Burroughs and others (1968) and Almgren (1978), it is possible to calculate approximate displacement rates on the Midland fault during the early Tertiary. Calculation of the displacement rates, shown in figure 22, assumes that displacement occurred on a single fault strand during the time deduced for various units by Almgren (1978) and that initial offset began in the Paleocene. Between 60 and 53 m.y. ago the slip rate was 3.8 cm/1,000 yr. The rate increased to a maximum of 7.2 cm/1,000 yr between 53 and 50 m.y., and it decreased significantly to 0.75 cm/1,000 yr between 49 and 42 m.y.

THORNTON ANTICLINE

The Thornton anticline, or the Thornton arch of Silcox (1968), is an east- to southeast-trending, west-plunging fold that is transected in part by closely spaced northwest-trending faults that generally show west-side-down displacements of less than 20 m. The fold extends from the east-central part of the Sacramento Delta to the vicinity of Lodi and provides structural control for several gas fields along its trace. Silcox (1968) provides the most detailed information on the structural development of the

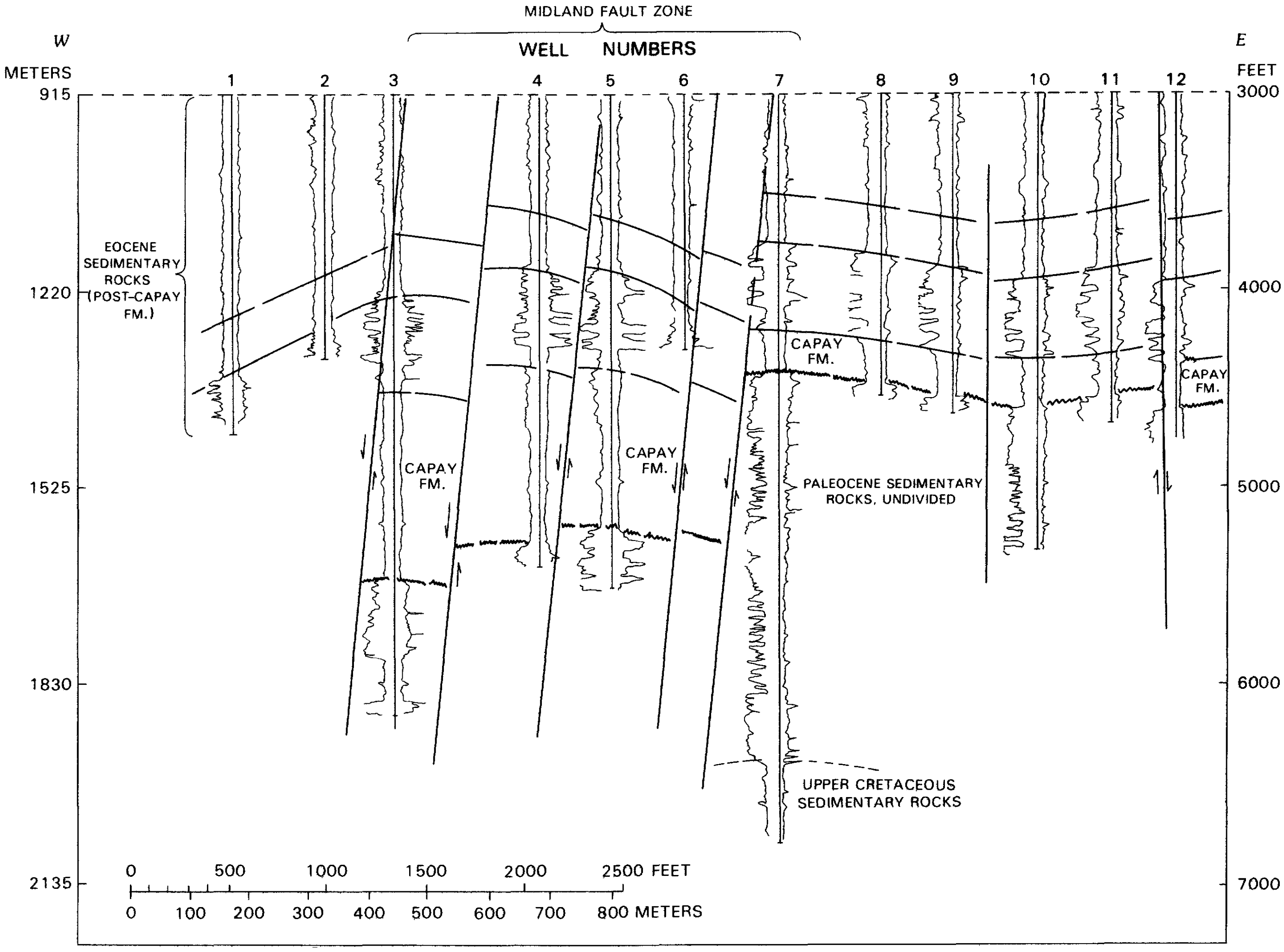


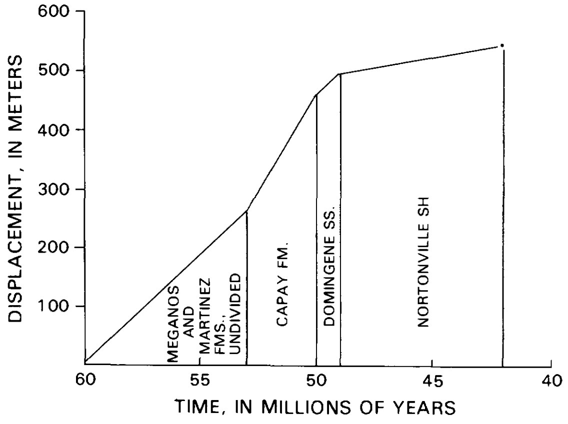
FIGURE 21.—Cross section showing electric logs and stratigraphic and structural relations across Midland fault zone in Rio Vista gas field (modified from Burroughs and others, 1968). Note significant thickening of the Eocene Capay Formation west of easternmost strand of Midland fault zone. Names of wells given by Burroughs and others (1968).

Thornton anticline in his description of the Thornton and Walnut Grove gas fields.

According to Silcox (1968) initial development of the Thornton anticline occurred prior to deposition of the basal Eocene rocks and may have been contemporaneous with Paleocene displacement on the Midland fault to the west. Maximum flexure of the structure apparently occurred in the late Eocene when lower and middle Eocene rocks were eroded and subsequently buried unconformably beneath deposits of the Markley submarine canyon. Miocene continental rocks, which were deposited unconformably on Markley canyon fill, do not show any significant folding over the Thornton anticline, nor are they offset by the northwest-trending faults that cut earlier deposits along the fold. Therefore, the structures are pre-Miocene.

STOCKTON FAULT

The Stockton fault is a northeast-trending, southeast-dipping subsurface fault that extends across the southern part of the Sacramento Valley between Tracy (south of pl. 1) on the west and the vicinity of Milton on the east. The fault lies north of a cross-valley, northeast-trending structural high in the basement, known as the Stockton arch, that apparently was discovered during some of the earliest seismic-reflection profiling done in the valley in the mid-1930's. Wells drilled near the Stockton fault during that time found only minor amounts of gas, and the relations between displacement on the Stockton fault and folding in the Upper Cretaceous marine rocks were not determined until development of the Lathrop gas field during the late 1940's. Teitsworth (1968), in his analysis of the Lathrop gas field, presents the most detailed study of the complex tectonic history of the Stockton fault. His work coupled with our analysis of wells to basement north of the Stockton fault, and Bartow's (1983) analysis of wells south of the fault provide the data for this discussion.



UNIT	DIFFERENTIAL THICKNESS (m)	PDSTFORMATION OFFSET (m)	TOTAL OFFSET (m)	TIME (m.y. ago)	RATE (cm/10³ yr)
Nortonville Sh	15	38	53	(49-42)	0.75
Domingene Ss.	26	23	49	(50-49)	4.9
Capay Fm.	215	0	215	(53-50)	7.2
Meganos and Martinez Fms , undivided	260	5	265	(60-53)	3.8

FIGURE 22.—Graph showing amounts and rates of offset on Midland fault zone, assuming all displacement occurred on a single fault strand and that displacement began in the Paleocene. Amount of offset and syntectonic thickening of units taken from figure 21 (modified from Burroughs and others, 1968).

Near the city of Stockton, the −3,000-m basement contour is offset about 5 km in a left-lateral sense and depressed about 300 m down to the south, but the vertical component of movement is not closely controlled in that area. Eastward, near the −2,500-m contour, the basement surface is offset left-laterally about 6 km and depressed nearly 500 m south of the fault. The vertical component of displacement there is reasonably well controlled by wells north and south of the fault. Both the left-lateral and vertical components of displacement decrease eastward along the Stockton fault and appear to die out completely between the −500-m and 0 basement contours. No trace of the Stockton fault has been mapped in the exposed metamorphic rocks of the Sierran foothills between the north end of Gopher Ridge and the Mokelumne River (pl. 1) (Wagner and others, 1981).

Teitsworth's (1968) study of the Lathrop gas field (pl. 1, map A), which is located about 9 km southeast of Stockton, corroborates our analysis of the basement offset on the Stockton fault and adds valuable data about the amount and timing of deformation recorded in the Upper Cretaceous rocks. Teitsworth concluded that the Upper Cretaceous rocks were thicker south of the Stockton fault, indicating south-side-down vertical movement during early phases of deposition, and that they were subsequently offset about 3 km left-laterally during Late Cretaceous time. Deposition of Upper Cretaceous marine shale followed earlier strike-slip displacement on the Stockton fault, and these marine shale units also show left-lateral displacement. Following deposition and strike-slip displacement of the uppermost Upper Cretaceous unit in the area, but prior to deposition of Miocene continental strata that unconformably cap the Upper Cretaceous section, the Stockton fault experienced about 1,000 m of reverse, south-side-up movement (Teitsworth, 1968). During this period of reverse movement, west- and west-northwest-trending anticlines formed in the Upper Cretaceous rocks south of the fault, providing closure that trapped the gas in the Lathrop and East Stockton gas fields (pl. 1, map A). The trend of these folds and the northwest-trending anticlines in the Roberts Island and McDonald Island gas fields (pl. 1, map A) north of the Stockton fault indicate northeast-southwest compression in the area during the period of reverse movement on the Stockton fault. Teitsworth (1968) concluded that reverse movement probably occurred after the middle Eocene but before deposition of Miocene continental deposits. However, he shows a west-northwest-trending fold in contours drawn on the base of the Miocene rocks south of the Stockton fault, from which we conclude that the northeast-southwest compressive stress operated in the area into the middle Tertiary.

FAULTS IN THE SIERRAN FOOTHILLS

Clark (1960) combined data from a number of sources with his own field observations to produce the first regional structural synthesis of the Sierran foothills. He defined the Foothill fault system as a group of northwest-trending, steeply east-dipping to vertical faults that tectonically separate distinctive belts of Paleozoic and Mesozoic rocks for more than 320 km along strike in the western foothills of the Sierra Nevada. He recognized that the component faults in the system were actually complex fault zones marked by multiple fault strands and intensely sheared, cataclastic, and crumpled rocks, which were associated geographically, in many areas, with major bodies of sheared serpentinite. On the basis of limited paleontologic and radiometric data, he concluded that the major tectonic activity of the Foothill fault system occurred in the Late Jurassic.

Although displacement of Tertiary volcanic units was recognized locally along segments of the Foothill fault system after Clark's study (Bateman and Warhaftig, 1966), few investigations were made into the late Cenozoic deformational history of the fault system before the Oroville earthquake. That seismic event produced ground rupture on the Cleveland Hills faults (Clark and others, 1976) and opened the question of potential late Cenozoic deformation within the whole Foothill fault system.

In their earthquake evaluation studies of the Auburn Dam area, Woodward-Clyde Consultants (1977) made a regional analysis of potential late Cenozoic deformation along many linear features related to the Foothill fault system, and they trenched those linear features that had direct bearing on their Auburn Dam study. That study provided most of the data for our analysis of those strands of the fault system in the border zone between the Sierran foothills and the Sacramento Valley.

COHASSET RIDGE FAULT

Cohasset Ridge is a prominent south- to southwest-trending interfluve located northeast of Chico, between the major drainages of Big Chico Creek and Deer Creek. The ridge is capped by at least two essentially contemporaneous olivine basalt flows that appear to have originated in the Butte Mountain area (about 12 km east of pl. 1), on the southwest slope of a Pliocene volcano named Mount Yana by Lydon (1968). The olivine basalts flowed southwestward in channels eroded in the underlying Tuscan Formation, but subsequent erosion has inverted the topography so that the basalt now stands as the highest unit on the ridge. The olivine basalt of Cohasset Ridge represents the final phase of volcanism that reached the Sacramento Valley from Mount Yana. The upper basalt flow has been dated at 2.41 m.y. (Harwood and others, 1981).

East of Brushy Mountain (pl. 1), the olivine basalt of Cohasset Ridge is abruptly truncated on the east by a steeply east-dipping fault that strikes about N. 40° W., roughly parallel to the Chico monocline fault. Aune (cited in Woodward-Clyde Consultants, 1977) first recognized this fault, which he called the Cohasset Ridge fault, and concluded that the eastern equivalents of the olivine basalt of Cohasset Ridge were downfaulted to the east about 30 m along the normal fault. The Cohasset Ridge fault can be traced north of Deer Creek through an intensely fractured zone in the Tuscan Formation to the vicinity of Mill Creek, where it becomes obscured by a complex pattern of west- and northwest-trending arcuate faults (pl. 1; fig. 19). Although the area south of Deer Creek is heavily forested and the slopes are covered by a veneer of colluvium, the trace of the Cohasset Ridge fault is defined by a prominent topographic linear feature, observed by Rich and Steele (1974), that extends nearly to Magalia, where it apparently is intersected by the Magalia fault.

The Cohasset Ridge fault is a northwest-trending, steeply east-dipping fault that has experienced at least 30 m of east-side-down normal movement in the past 2.41 m.y.

MAGALIA FAULT

About 9 km north of Magalia, (pl. 1), the Cohasset Ridge fault is intersected by the N. 20° W.-trending Magalia fault. Extensive mine workings, which followed Tertiary auriferous gravel deposits at the base of the Tertiary volcanic sequence in the area, provide detailed data on the nature and local extent of the Magalia fault (Gassaway, 1899; Logan, 1930; Woodward-Clyde Consultants, 1977).

Although the Magalia fault is shown as a single fault on the map, detailed mine maps indicate that the fault is actually a complex fault zone consisting of numerous fault strands that have different orientations and amounts of displacement. In the Dix mine at the north end of the mining district, the gold-bearing gravels are offset 28 m down to the east. Southeastward along the trace of the Magalia fault at the Black Diamond mine, the gravel deposits are offset 68 m down to the east. Farther southeast at the Magalia mine, the channel deposits are offset down to the west from 2 to 11 m along three reverse faults that trend N. 60° W. and dip 58° NE.

From the mining records it appears that the various fault strands in the Magalia fault zone have experienced episodic and different movement through the Tertiary, with both normal east-side-down and reverse east-side-up displacement recorded. The relative ages of this disparate movement pattern along the Magalia fault are unknown, but the youngest movement appears to post-date movement on the Cohasset Ridge fault.

CLEVELAND HILLS FAULTS

The Cleveland Hills faults coincide with surface ruptures that occurred during the Oroville earthquake. During that seismic event, the ground surface failed in a 3.8-km long en echelon pattern of north- and north-northwest-trending normal faults that showed at least 55 mm of horizontal separation across the surface ruptures and as much as 180 mm of west-side-down vertical separation (Clark and others, 1976). Aftershocks of the Oroville earthquake defined a zone of seismic activity, assumed to coincide with the controlling fault, that dipped about 60° W. and trended nearly due north (fig. 15; also Bufe and others, 1976).

Trenches, dug across the zone of surface rupture by the California Department of Water Resources and logged by Akers and McQuilkin (1975), showed a gouge zone about 2 m wide in the bedrock below the surface ruptures that was flanked by 2-4 m of intensely fractured rock transitional into the country rock of foliated greenstone. The gouge zone contains anastomosing shear zones of deformed gouge that provide evidence of repeated earlier faulting along the zone of the ground failure. Earlier deformation in the area is also indicated by 30-60 m of apparent west-side-down offset of an erosion surface of probable Pleistocene age across the Cleveland Hills faults (Aune, 1975).

About 5.7 km northwest of the Cleveland Hills faults (pl. 1), Creely (1965) mapped a north-trending, steeply east-dipping fault that offset the base of Pleistocene or older gravel deposits about 2 m down to the west. Clark and others (1976) reported a second fault in the area that showed minor east-side-down offset of the gravel deposit.

FAULTS SOUTHEAST OF OROVILLE

In their earthquake evaluation of the Auburn Dam area, Woodward-Clyde Consultants (1977) identified a number of prominent linear features in the Sierran foothills southeast of the Cleveland Hills faults. With the exception of the northeast- and north-trending Highway 49 and Hancock Creek lineament zones (pl. 1), the linear features they investigated have northwest trends that coincide closely with the projected trace of the Chico monocline fault. The linear features were trenched extensively and logged in detail (Woodward-Clyde Consultants, 1977), and trench data were integrated with surficial geologic investigations in each area. Evidence of movement in the past 10 m.y. (Woodward-Clyde Consultants' definition of the "late Cenozoic") was detected in the trenches across the northwest-trending linear features, which are shown on plate 1 as the Swains Ravine, Spenceville, and Maidu fault zones and the Dewitt fault.

Deformation in the bedrock exposed in the trenches is remarkably similar to that shown in trenches across the Cleveland Hills faults. In most areas, bedrock beneath the surface lineaments contained one or more zones of slickensided clay-rich gouge a few meters wide or less. The gouge commonly contained anastomosing shear zones composed of polydeformed gouge and thin clay seams that indicated multiple phases of deformation. Rock adjacent to the gouge zones commonly was highly fractured, bleached, or manganese-stained and injected by thin quartz veins. In some trenches, a paleosol rested unconformably on bedrock and was offset locally along the faults marked by the gouge zones. That evidence led Woodward-Clyde Consultants (1977) to conclude that at least some of the faulting had occurred in the past 100,000 years. Based on the evidence presented by Woodward-Clyde Consultants (1977) and our observations in many of the trenches that they excavated during that study, we agree with their conclusions.

REGIONAL STRUCTURAL ANALYSIS

STRUCTURAL DOMAINS IN THE SACRAMENTO VALLEY

As we have shown in preceding sections of this report, late Cenozoic structural features in the valley have diverse orientations and displacement histories. In spite of this diversity, large areas of the valley are characterized by coeval structures that appear to have formed in a similar stress regime. To analyze the late Cenozoic structural evolution of the Sacramento Valley, we have grouped coeval structures that have similar displacement patterns into structural domains (fig. 23).

Delineating the various structural domains obviously depends on correctly identifying and correlating late Cenozoic deposits throughout the valley. Many of the late Cenozoic alluvial deposits are similar in outcrop characteristics, but fortunately they contain or are bracketed by distinctive tephra beds or volcanic flows that allow widespread correlation of alluvial units and provide absolute age control for the alluvial and tectonic histories of the region. In this regard, we have used the following time-stratigraphic units to define the late Cenozoic structural domains of the valley:

Unit	*Age/Age range (m.y.)*	*Reference*
Rockland ash bed	0.4-0.45	Sarna-Wojcicki and others (1985); Meyer and others (1980)
Red Bluff Formation	0.5-1.0	Harwood and others (1981)
Olivine basalt of Deer Creek	1.09	Harwood and others (1981)
Ishi Tuff Member of the Tuscan Formation	2.6	Harwood and others (1981; unpublished data, 1985)
Volcanic rocks of Sutter Buttes	1.4-2.4	Williams and Curtis (1977)
Nomlaki Tuff Member of the Tuscan and Tehama Formations	3.4	Evernden and others (1964)
Putah Tuff Member of the Tehama Formation	3.4	Sarna-Wojcicki (1976)

With the exception of the Dunnigan Hills, the structural domains shown in figure 23 become progressively younger northward through the valley. Late Cenozoic deformation in the Dunnigan Hills domain apparently overlapped that in the Corning domain, but the respective styles of deformation and the inferred stress regimes discussed in a later section differ significantly. This point emphasizes the fact that the structural domains include tectonic features that are homogeneous in time and style of deformation as far as we can determine. To a certain extent, the ages of the structural domains reflect the distribution of the time-stratigraphic units used to identify the major late Cenozoic deformation in a domain. It may be possible to subdivide some domains as more information on the distribution of the younger time-stratigraphic units becomes available. Furthermore, it is important to note that ascribing an age to the late Cenozoic deformation in a domain does not imply that all deformation necessarily began or ended within that time. Many late Cenozoic structural features in the valley formed over earlier structures in the upper Mesozoic or Tertiary rocks, and thus they owe their existence, in part, to tectonic heredity. In some areas, historic low-magnitude earthquakes can be identified with specific structural features such as the Battle Creek fault zone (Bolt, 1979) and Corning fault (Marks and Lindh, 1978), indicating that strain is currently being released on these structures.

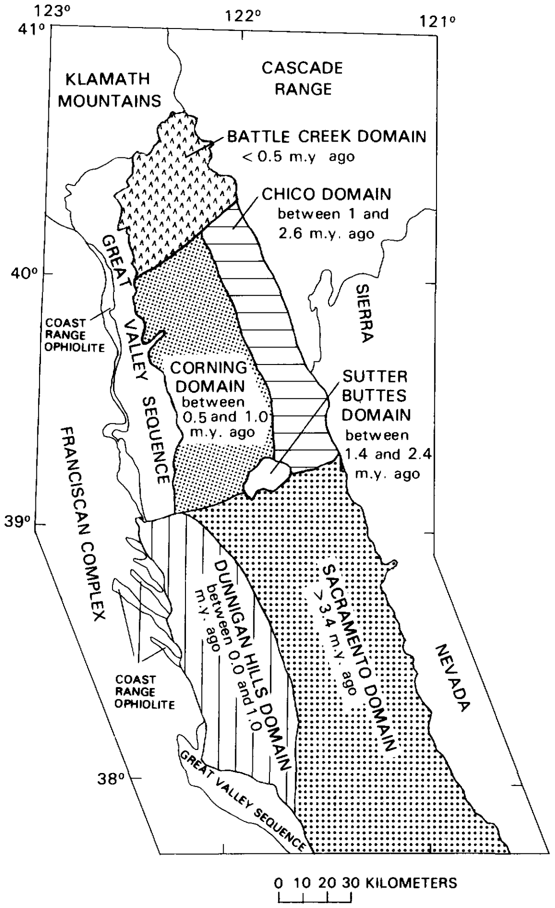


FIGURE 23.—Structural domains in Sacramento Valley. Numbers under domain names give duration or age of deformation that produced late Cenozoic structures in that domain.

SACRAMENTO DOMAIN

The Sacramento domain includes the Stockton fault, the Midland fault, the Thornton anticline, and the possible southeast extension of the Willows fault and other unnamed northwest-trending faults southeast of Sutter Buttes. None of these structures appear to offset or deform late Cenozoic alluvial deposits that are younger than the Nomlaki Tuff Member (figs. 7-10; Silcox, 1968; Teitsworth, 1968).

The minimum age of deformation is difficult to determine. If the Markley submarine canyon was localized, in part, by movement on the southeast extension of the Willows fault, as suggested by figures 8 and 9, then major deformation occurred in the domain in the middle Tertiary and could be as young as early Miocene, which is the upper age of the Markley canyon fill (Almgren, 1978). However, the Pliocene Laguna Formation, which contains the Nomlaki Tuff Member near its base just east of Marysville (Bussaca, 1982), does not appear to be offset in this domain; therefore, we conclude that deformation is older than 3.4 m.y. Upper Cretaceous rocks indicate syndepositional offset on the Stockton fault (Teitsworth, 1968) so the maximum age of deformation is Late Cretaceous.

The Willows fault north of Sutter Buttes (fig. 3) and the faults in Capay Valley (fig. 20) also show protracted deformation, ranging in age from Late Cretaceous to the middle Tertiary, that undoubtedly was contemporaneous with deformation in the Sacramento domain.

SUTTER BUTTES DOMAIN

The complex pattern of folds and faults at Sutter Buttes and the buried Colusa dome was interpreted by Williams

(1929) and Williams and Curtis (1977) to have formed by forceful intrusion of magma between 2.4 and 1.4 m.y. ago. Harwood (1984; this report) suggested that magmatism was localized at Sutter Buttes and the buried Colusa dome by movement on the Willows fault and possibly on the Mesozoic tectonic boundary between oceanic and Sierran basement terranes beneath Sutter Buttes. In this interpretation, deformation in the Sutter Buttes domain occurred through the combined effects of forceful intrusion of magma and east-west compression in a regional stress field.

Deformation in the Sutter Buttes domain may have overlapped in time with deformation in the Chico domain. However, the style of tectonism at Sutter Buttes is unique, and the area must be considered as a separate domain in a regional analysis.

Although Williams and Curtis (1977) suggested that Sutter Buttes formed in roughly half a million years, based on volume and morphology of volcanic deposits, their K/Ar ages for the volcanic complex range from about 1.4 to 2.4 m.y., and we have used that age range as the age of the structural domain.

CHICO DOMAIN

The Chico monocline is the dominant structure in the Chico domain (fig. 23). The monoclinal flexure and the controlling fault at depth (fig. 14) formed after the eruption of the Ishi Tuff Member of the Tuscan Formation, about 2.6 m.y. ago (Harwood, unpub. data, 1984), but before the eruption of the olivine basalt of Deer Creek, about 1.0 m.y. ago (Harwood and others, 1981). Formation of the monocline resulted directly from uplift of the northern Sierra Nevada and rupture along the controlling fault beneath the monocline. On the basis of geophysical evidence (Griscom, 1973; Cady, 1975), it seems reasonably certain that the controlling fault was a major late Mesozoic structure, separating oceanic from magmatic-arc crustal blocks, that was reactivated in the late Cenozoic.

Late Cenozoic tectonic uplift markedly increased erosion in streams antecedent to the monocline from Deer Creek northwest to Salt Creek just east of Red Bluff. Due to their tectonically steepened gradients, these streams deposited coarse, bouldery alluvial fanglomerate composed of volcanic material derived from the Tuscan Formation, extending from the monoclinal arch westward into the valley. The olivine basalt of Deer Creek filled the ancestral channel of Deer Creek east of the monocline and cascaded over the coarse fanglomerate, forming westward-overturned folds at the base of the flow west of the monoclinal flexure.

In Salt Creek and along nearby Hogback Road just east of Red Bluff, the fanglomerate (unit QTog of Harwood and others, 1981) rests unconformably on a partially stripped soil that developed on the Tuscan Formation. The soil indicates a significant period of nondeposition and weathering prior to flexing of the monocline and prior to rapid deposition of the alluvial fanglomerate. A period of erosion also followed deposition of the fanglomerate. The Red Bluff Formation was deposited unconformably on both the Tuscan and fanglomerate in Salt Creek. These stratigraphic relations, coupled with the fact that the Ishi Tuff Member is flexed, eroded, and unconformably overlain by the olivine basalt of Deer Creek along the monocline, establish deformation in the Chico domain between 1.0 and 2.6 m.y. This is clearly before deformation in the Corning domain to the west, where the Red Bluff is significantly folded and faulted.

The orientation of the horizontal compressive stress vector is not closely constrained in the Chico domain. Slickensides on some faults along the monocline flexure have a rake of 35° NW., suggesting some left-lateral oblique slip on the northwest-trending fault system. Whether the slickensides developed at the time the monocline formed or subsequently is unknown, but if they do record strain released during flexing of the monocline, the principal horizontal stress may have been oriented approximately east-west.

CORNING DOMAIN

West of the Chico monocline in an area extending to the Willows fault, late Cenozoic structures range in orientation from northwest to approximately north. The main stem of the Willows fault and the fault along the flank of South Corning dome nearly parallel the trend of the Chico monocline. On the other hand, the Corning fault, the Corning domes, and the Los Molinos and Glenn synclines trend more northerly than the monocline and converge on that structure near Red Bluff. All of these structures, within the Corning domain (fig. 23), deform the Red Bluff Formation and thus postdate the flexing of the Chico monocline. However, the subparallel orientation of structures in the Corning domain northeast of the Willows fault and of those in the Chico domain suggests that all may have formed in a common stress field.

The maximum horizontal compressive stress probably was oriented approximately east-west, normal to the trace of the Corning domes. The Corning fault might represent one shear diagonal of the strain ellipse, and the northwest-trending faults would indicate the other shear diagonal. In this model, strain release would have migrated westward with time, beginning along the Chico monocline and later moving to the Corning domes and Corning fault. Clearly, some offset may have occurred on the Corning fault and the northwest part of the Willows fault during formation of the monocline, and the small amount of offset in the olivine basalt of Deer Creek over the monocline may have occurred during the upwarping of the Red Bluff

Formation over the Corning domes.

The area of the Corning domain southwest of Sutter Buttes and the Willows fault is largely capped by Holocene basin deposits that lap westward onto Upper Cretaceous rocks along the western margin of the valley. These deposits could obscure many structures that might be recorded in the Red Bluff Formation in that area, indicating that the downthrown side of the Willows fault has been a structural basin throughout most of the late Quaternary.

DUNNIGAN HILLS DOMAIN

The Dunnigan Hills domain (fig. 23) includes the Zamora fault, the Dunnigan Hills anticline, the Madison syncline, and the tectonically complex area of Capay Valley and the Capay Hills. The Red Bluff Formation is folded by the Dunnigan Hills anticline and offset by the Zamora fault along the east margin of the Dunnigan Hills (fig. 20). Therefore, the late Cenozoic structures in that area are at least as young as the folds and faults in the vicinity of Corning domes, and they could be younger. Because there is no clear evidence for the minimum age of structures in the domain, we assume that late Cenozoic deformation ranges from 1.0 m.y. to the present.

Although the late Cenozoic deformation may overlap in time with that in the Corning and Battle Creek domains, the orientation and movement pattern of structures are significantly different. The northwest-trending axial surface of the Dunnigan Hills anticline and normal displacement on the northwest-trending Zamora fault suggest that these structures may have formed in a stress field in which the maximum horizontal compressive stress was oriented approximately north-south and the least horizontal compressive stress (maximum extension) was oriented approximately east-west. Because this is the present orientation of the stress field associated with the San Andreas fault in this area (Zoback and Zoback, 1980), late Cenozoic structures in the Dunnigan Hills domain may be related to right-slip wrench tectonism associated with movement on the San Andreas fault system (Allen, 1981).

BATTLE CREEK DOMAIN

The Battle Creek domain (fig. 23) is the youngest structural domain and occurs at the northern end of the valley. It is characterized by northeast-trending folds and faults oriented nearly perpendicular to the north- and northwest-trending structures of other domains to the south.

Late Cenozoic displacement on the northeast-trending, steeply south-dipping faults is invariably down to the south, with a suggestion of minor right-lateral slip on the Battle Creek fault system (Helley and others, 1981). The amount of horizontal separation on the Battle Creek fault is unknown, but it probably is on the order of a few tens of meters to a few hundred meters at the most. The Bear Creek and Red Bluff faults are not exposed well enough to establish or refute the occurrence of late Cenozoic horizontal slip on those structures. The amount of vertical displacement increases eastward along the Battle Creek fault zone, with about half of the vertical separation of units caused by broad anticlinal folding and half by fault displacement (fig. 17). Vertical separation of late Cenozoic units on the Bear Creek and Red Bluff faults is also accomplished partly by folding.

These observations, plus the fact that the Inks Creek fold system appears to be genetically related to folding and faulting on the Battle Creek fault zone, indicate that the structures in the Battle Creek domain formed either by northwest-southeast horizontal compressive stress or by right-lateral shear stress. Strain was released initially by folding, followed by rupture and predominantly normal fault displacement. Initial movement on the Battle Creek fault zone postdated deposition of fanglomerate, which is correlated with the Red Bluff Formation, and the fault scarp restricted movement of the proximal ash-flow part of the Rockland ash bed. Deformation younger than 0.45 m.y. is indicated by distribution of the early Riverbank-age alluvial deposits north of the Battle Creek fault zone and in the tectonically controlled loops of the Sacramento River around the Inks Creek fold system. The late Cenozoic structures in the Battle Creek domain formed in the past half million years.

Little surface evidence is available to document deformation before the late Cenozoic in the Battle Creek domain. A detailed analysis of paleomagnetic directions and lithofacies in the Upper Cretaceous marine rocks and Eocene continental rocks, which locally overlie the Cretaceous rocks, might provide data to resolve this question, but we did not pursue these studies. In the light of recent paleomagnetic studies (Simpson and Cox, 1980; Beck and Plumley, 1980; Magill and Cox, 1981), which suggest a two-phase, clockwise rotation of the Oregon Coast Ranges and possibly of the Klamath Mountains in Eocene and post-Oligocene times, we believe the Battle Creek domain may be a zone of decoupling between the Klamath Mountains and Oregon Coast Ranges provinces to the north and the Sacramento Valley-Sierra Nevada provinces to the south.

VALLEY STRUCTURES RELATED TO REGIONAL STRESS PATTERNS

With the exception of structures in the Dunnigan Hills domain and recent movement on the Cleveland Hills faults, late Cenozoic folds and faults in the Sacramento Valley appear to have formed in an east-west compressive stress regime. The north- and northwest-trending

Willows, Corning, and Chico monocline faults dip steeply east and show reverse displacements. The east-northeast-trending Red Bluff, Battle Creek, and Bear Creek faults dip steeply south and have normal offsets. The axial traces of many folds are parallel to the trends of adjacent faults, and the folding appears to be related to drag on those faults. The amount of lateral displacement on faults in the valley is difficult to determine, but the rake of slickensides and the orientation of ancilliary fractures suggest minor left-lateral movement on the northwest-trending Chico monocline fault and minor right-lateral slip on the east-northeast-trending Battle Creek fault zone. These kinematic patterns are consistent with a regional stress field in which the maximum horizontal component of compressive stress is oriented about N. 75° E. (Harwood, 1984).

The late Cenozoic kinematic pattern and inferred east-west compressive stress regime in the Sacramento Valley appear to be anomalous with respect to contemporary tectonism in adjacent regions. In the California Coast Ranges to the west and in the northern Basin and Range province to the east, north-trending faults generally show normal and right-lateral displacements, east-trending faults show reverse and left-lateral movement, and northwest- and northeast-trending faults show dominantly right-lateral and left-lateral displacement, respectively (Zoback and Zoback, 1980; Hill, 1982). From these kinematic patterns, as well as earthquake focal-plane solutions, and in-situ stress measurements, Zoback and Zoback (1980) and Hill (1982) inferred that the maximum horizontal compressive stress was oriented approximately north-south and the least horizontal compressive stress (maximum extension) was oriented approximately east-west. Hill (1982) related the contemporary stress patterns to movement between the North American and Pacific plates and visualized the western part of the North American plate as a continuous broad zone of deformation following the model first proposed by Atwater (1970).

East-west compressive deformation in the Sacramento Valley suggests that the late Cenozoic stress field was not homogeneous between the continental margin and the northern Basin and Range province. Instead, the Sacramento Valley apparently acted as an independent block where relatively small-scale compressive strain was periodically released in response to large-scale right-lateral transform tectonism in the San Andreas fault zone to the west and major east-west crustal extension in the northern Basin and Range province to the east.

Furthermore, the well-dated deformation patterns in the Sacramento Valley indicate that the compressive strain was not released randomly but, rather, that the late Cenozoic structural features formed in a sequential pattern that is progressively younger to the north. Northward progression of the compressive deformation implies a northward-migrating stress regime or a migrating energy source sufficient to initiate deformation in a regional compressive stress field. The interaction of lithospheric plates along the continental margin appears to provide a reasonable mechanism for generating the sequential compressive strain release observed in the valley.

Successive positions of the Mendocino transform, extrapolated backward in time to 4.0 m.y. at a rate of 5.5 cm/yr (Atwater and Molnar, 1973), are shown relative to the late Cenozoic structural domains in the Sacramento Valley in figure 24. Each domain is characterized by a particular set of late Cenozoic structural features that formed during a definite period of time. The time of deformation in the structural domains is linked to respective positions of the Mendocino transform by corresponding patterns on figure 24. Compressive deformation in the Sacramento domain occurred prior to eruption of the Nomlaki and Putah Tuff Members of the Tehama Formation about 3.4 m.y. ago, but we are unable to correlate movement on specific structures in that domain with relative plate motions shown by successive positions of the Mendocino transform. Beginning with magmatism and deformation in the Sutter Buttes domain about 2.4 m.y. ago, however, reasonable correlation exists between the position of the Mendocino transform and progressive east-west compressive deformation in the Sacramento Valley. This correlation clearly reflects the availability of geologic information that dates deformation in the northern part of the valley, but it may also reflect a greater degree of tectonism in the valley over the past 2.4 m.y. that was related to increased tectonism ahead of the advancing Mendocino triple junction (fig. 24).

According to Silver (1971), an episode of major bending and fracturing occurred in the Juan de Fuca plate (fig. 24) between 2.5 and 0.5 m.y. ago. This deformation apparently increased the north-south compressive stress ahead of the advancing Mendocino triple junction and caused underthrusting of oceanic crust in the Juan de Fuca plate beneath the Mendocino transform. Dickinson and Snyder (1979a) suggested that this type of instability at the Mendocino triple junction would increase the east-west extensional stress in the vicinity of the triple junction and possibly cause eastward migration of right-slip strain release along the San Andreas fault zone. Internal deformation of the Juan de Fuca plate also may have changed the rate and relative motion of the subducted part of that plate relative to the North American plate sufficiently to initiate magmatism at Sutter Buttes and at the Pliocene volcanic centers of Mount Yana and Mount Maidu (fig. 24) (Lydon, 1968) about 2.4 m.y. ago. Subsequent relative movement of the North American plate southwestward from its position above the subducted Juan de Fuca slab, over the eastern projection of the Mendocino transform and onto the area of the slab window of Dickinson and Snyder (1979b), may have initiated progressive deformation in the Sacramento Valley north of the Sutter Buttes domain.

According to Dickinson and Snyder (1979b), an expanding, triangular-shaped region, which they called the slab window, has developed east of the lengthening San Andreas transform since the rise-trench encounter about 20 m.y. ago. As the slab window expanded, upwelling asthenosphere came into contact with the base of the North American crust producing an expanding region of Neogene extensional tectonism. The northern boundary of the slab window, which is the eastern projection of the Mendocino transform, migrated northward as a consequence of the northward migration of the Mendocino triple junction and relative motion between the North American and Juan de Fuca plates. Successive positions of the northern boundary of the slab window beneath the Sacramento Valley for the past 2.5 m.y. are shown in figure 25.

About 2.5 m.y. ago, the northern boundary of the slab window was located approximately at the latitude of Sutter Buttes (fig. 25*A*). As the North American plate moved southwestward relative to the Juan de Fuca plate, the base of the valley crust near Sutter Buttes moved off the subducted ocean crust of the Juan de Fuca plate and encountered asthenophere in the slab window. This change in thermal regimes at the base of the valley crust possibly reactivated the Willows fault and the ancient tectonic boundary between ophiolitic basement rocks and the Sierran basement rocks, allowing magma to rise and differentiate below Sutter Buttes and the buried Colusa dome. Subsequent southwestward relative movement of the North American plate exposed an increasing area of the Sacramento Valley and northern Sierra Nevada crustal blocks to the higher thermal regime of the slab window. Basaltic volcanism erupted in the Tahoe-Almanor graben (fig. 25*A*) north of Lake Tahoe soon after magmatism began at Sutter Buttes (Dalrymple, 1964). Ac-

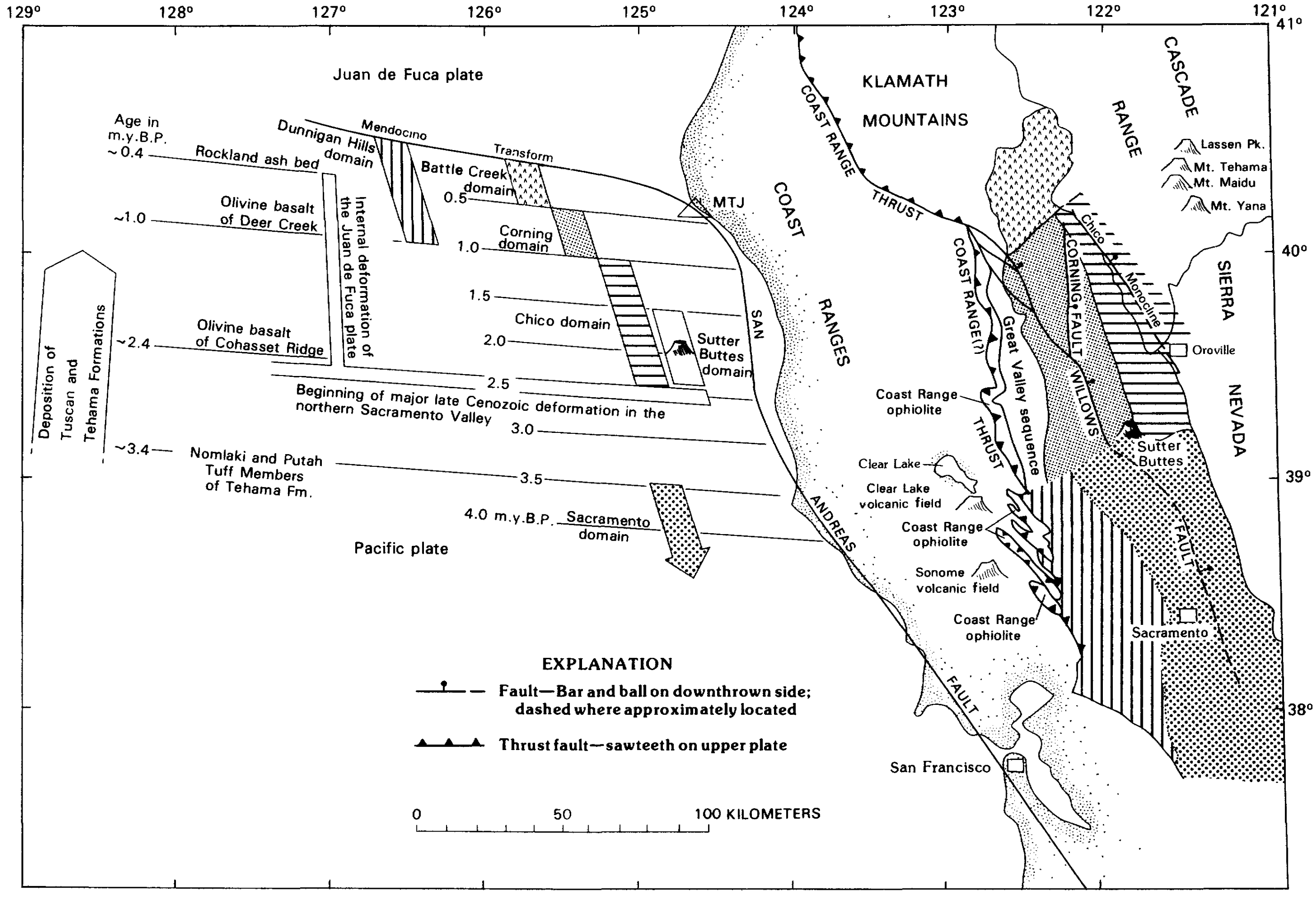
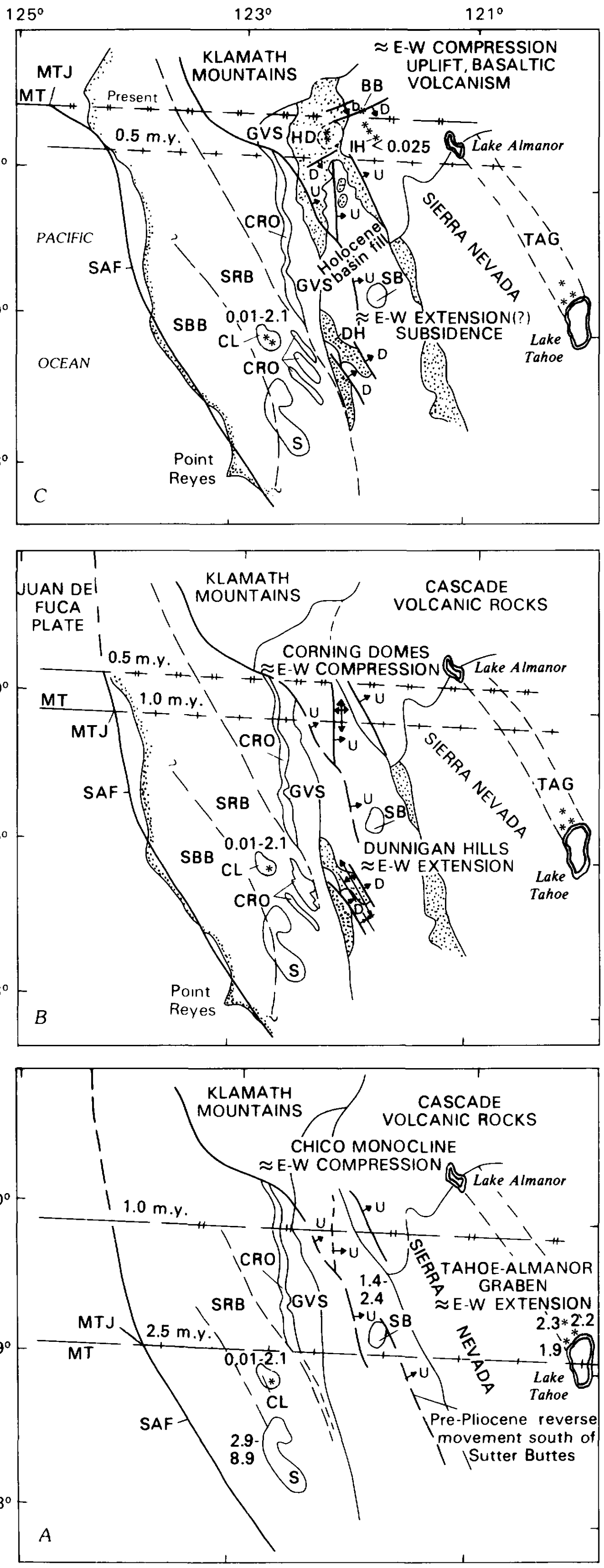


FIGURE 24.—Map of northwestern California relating late Cenozoic structural features and structural domains (patterned) to inferred position of the Mendocino triple junction (MTJ) during time structures formed. Position of Mendocino transform shown in 0.5-m.y. increments during past 4.0 m.y. Dated volcanic units used to bracket age of deformation in Sacramento Valley shown at left.

celerated uplift of the Sierra Nevada, which began about 3.0 m.y. ago in the central part of the range (Huber, 1981; Slemmons and others, 1979), eventually caused rupture between 2.5 and 1.0 m.y. ago along the Mesozoic fault that separates valley and Sierran blocks beneath the Chico monocline. Volcanism at Clear Lake (fig. 25), which began about 2.1 m.y. ago (Donnelly-Nolan and others, 1981), lagged behind passage of the northern boundary of the slab window and has continued nearly to the present.

Between 1.0 and 0.5 m.y. ago (fig. 25*B*), exposure of the base of the valley crust to the slab window may have reactivated the basement fault in the vicinity of the Corning domes, resulting in east-west compressional tectonism. Wrench tectonism and related east-west extension apparently started to affect the southwestern part of the valley after creating a zone of high shear strain in the Santa Rosa block (Fox, 1983), located in the eastern part of the Coast Ranges province. This is the earliest evidence of east-west extensional tectonism found in the valley and suggests a delay time of about 2.5 m.y. between passage of the slab-window boundary and production of recognizable extensional structures.

In the past 0.5 m.y., the northern boundary of the slab window has passed beneath about half of the Battle Creek domain (figs. 24, 25*C*). The morphology of the valley in the Battle Creek domain, therefore, may reflect recent,



EXPLANATION

▒	Pliocene and Pleistocene alluvium
*	Volcanic center at—
BB	Black Butte
IH	Inskip Hill
HD	Hooker dome
CL	Clear Lake
SB	Sutter Buttes
S	Sonoma volcanic field
TAG	Tahoe-Almanor graben
MT	Mendocino transform
MTJ	Mendocino triple junction
SAF	San Andreas fault
SBB	Sebastopol block
SRB	Santa Rosa block
CRO	Coast Range ophiolite
GVS	Great Valley sequence
DH	Dunnigan Hills
1.4-2.4	Numbers indicate age or time of formation of feature, in millions of years. ≈ means approximate
— —	Contact—Dashed where inferred
	Fault—Dashed where inferred; arrow indicates dip
↑D —	Normal—D, hanging wall
↑U —	Reverse—U, hanging wall
↔	Doubly plunging anticline
—+—	Syncline

FIGURE 25.—Sketch maps of northwestern California showing inferred positions of northern boundary (barbed dashed lines) of slab window of Dickinson and Snyder (1979b) at about 2.5 and 1.0 m.y. (*A*), 1.0 and 0.5 m.y. (*B*) and 0.5 m.y. and present (*C*).

and possibly on-going, structural changes in the deeper parts of the valley fill and basement caused in part by changes in the thermal regime beneath the thick valley crust. This hypothesis was investigated by Helley and Jaworowski (1983), who used the distribution of the Red Bluff pediment as a surface indicator of deeper valley structures. The elevation of the floor rises abruptly north of Red Bluff, and the dissected uplands expose the Pliocene Tehama Formation that is unconformably capped by remnants of the Red Bluff Formation. Drainage patterns reflect local structures, such as Hooker dome (fig. 18), and major creeks locally follow the dominant northeast structural trends. The gradient of the Sacramento River steepens significantly across the Battle Creek domain, and its incised meanders are structurally controlled. Many of these geomorphic features are preserved, in somewhat more subdued form, around Corning domes and elsewhere in the valley south of Red Bluff (Helley and Jaworowski, 1985).

The northeast-trending Inks Creek fold system and the folding associated with the Battle Creek and Red Bluff faults suggest a northwest-southeast compressive stress system. If such is the case, however, normal displacement on the Bear Creek, Battle Creek, and Red Bluff faults are anomalous. The structural features in the Battle Creek domain may reflect a stress regime in which the maximum horizontal compressive stress (east-west) and the least horizontal compressive stress (north-south) vectors are more nearly equal. This inferred change in the stress pattern may be caused by oblique subduction of the Juan de Fuca plate and the possible buttressing effect of the Klamath Mountains on deformation in the northernmost part of the Sacramento Valley.

CONCLUSIONS

Kinematic patterns of late Cenozoic structural features in the Sacramento Valley differ significantly from those in the Coast Ranges province to the west and the northern Basin and Range province to the east. Certainly for the past 2.5 m.y., and probably much longer, deformation in the valley has occurred in a regional stress field in which the maximum horizontal component of compressive stress was oriented approximately east-west and the minimum component of compressive stress (maximum extension) was oriented approximately north-south. Within this stress regime, strain has been released primarily by reverse movement on north- and northwest-trending high-angle faults and associated folding in the sedimentary rocks of the valley fill. This style of deformation contrasts sharply with the pattern of northwest-trending en echelon folds and pull-apart basins, and east-west oriented thrust faults associated with large-scale right-lateral displacements on the San Andreas fault system (Blake and others, 1978; Graham, 1978; Harding, 1976; McLaughlin, 1981; Page, 1981) and the pattern of pervasive normal faulting, widespread volcanism, and large-scale east-west extension in the Basin and Range province (Christiansen and Lipman, 1972; Lipman and others, 1972; Zoback and others, 1981). During the late Cenozoic, the Sacramento Valley appears to have acted as an independent block on which relatively small-scale compressive deformation was imposed by eastward-directed subduction that was followed by large-scale transform tectonism along the continental margin and major east-west crustal extension in the Basin and Range. Early Tertiary deformation on the Willows fault and in the area of Capay Valley may be related to eastward tectonic wedging of the Franciscan Complex described by Wentworth and others (1984).

Tectonic heredity appears to have played a significant role in late Cenozoic deformation in the Sacramento Valley. Many of the faults that break or drape the upper Cenozoic deposits are located over pre-existing structures within and at the boundaries of basement blocks that were tectonically juxtaposed and broken during the Mesozoic and early Tertiary. The tectonic boundary between ophiolitic and Sierran basement rocks apparently has been reactivated beneath the Chico monocline and the possible southeast extension of the Willows fault. The Corning fault occurs over a Mesozoic fault within the ophiolitic basement terrane.

Within the past 2.5 m.y., at least, late Cenozoic east-west compressive deformation has progressed northward so that structures observed in the Battle Creek domain are a million years younger than those around Sutter Buttes. The northward progress of late Cenozoic deformation correlates with extrapolated positions of the Mendocino transform. Deformation may have been initiated by thermal changes in the valley basement as the North American plate moved relatively southwestward off the subducted oceanic crust of the Juan de Fuca plate and over asthenosphere in the slab window proposed by Dickinson and Snyder (1979b). During this sequential deformation, the relative magnitude of the horizontal compressive stress vectors appears to have changed. The kinematic pattern of structures south of Red Bluff suggests that east-west compression was dominant ($S_{\mathrm{maxEW}} >> S_{\mathrm{minNS}}$) and resulted in high-angle reverse faulting. However, the horizontal compressive stresses appear to have been more nearly equal ($S_{\mathrm{maxEW}} \cong S_{\mathrm{minNS}}$) in the Battle Creek domain causing northeast-trending folds and normal displacements on east-northeast-trending faults.

Northwest-trending folds and normal displacement on the Zamora fault suggest that east-west extension may have affected the southwestern part of the valley during the past million years, when east-west compressive deformation was occurring in the Corning domain. Focal-plane solutions constrained by geologic data in the Clear

Lake area indicate that the maximum compressive stress in the Coast Ranges is presently oriented approximately north-south (Bufe and others, 1981). These data, coupled with the observation that right-lateral tectonism is more intense in the Coast Ranges adjacent to the Dunnigan Hills domain than it is to the west or north (Fox, 1983), suggest that late Cenozoic deformation in the Dunnigan Hills domain was related to wrench tectonism on the San Andreas fault system. If such is the case, the beginning of right-lateral wrench tectonism in the valley appears to have lagged behind the northward migration of the Mendocino triple junction by about 2.5 m.y. A similar delay was noted between the migration of the triple junction and the build-up of maximum heat flow in the Coast Ranges by Lachenbruch and Sass (1980).

The well-dated and diverse deformation patterns in the Sacramento Valley indicate that late Cenozoic tectonism evolved through the region in response to major crustal movements outside the valley's physiographic boundaries. East-west compressive tectonism may have been imposed on the valley by eastward subduction of the Juan de Fuca plate, and consequent tectonic wedging of the Franciscan Complex coupled with a major component of westward stress due to east-west extension in the Basin and Range province.

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APPENDIX—*Wells with electric well-log data shown in cross sections*

Well no.	Operator	Well name	Sec.-T.-R.
	Wells on figure 3		
1.	Texaco	Cross No. 1	27-20N.-3W.
2.	Franco Western Oil Co.	Dunlap 36-1	36-20N.-3W.
3.	Marathon Oil Co.	Capital Co. No. 1	30-20N.-2W.
4.	R. W. McBurney	Westpet Sunray McBurney Capitol No. 1	30-20N.-2W.
5.	Mobil Oil Corp.	Capital Co. No. 30-1	30-20N.-2W.
6.	Mobil Oil Corp.	Section 29 Unit No. 4	29-20N.-2W.
7.	R. W. McBurney	Section 29 Unit No. 2	29-20N.-2W.
8.	Mobil Oil Corp.	Section 48 Unit No. 4	21-20N.-2W.
9.	Mobil Oil Corp.	Section 48 Unit No. 1	21-20N.-2W.
10.	Standard Oil Co.	Doheny N. G. Unit No. 1	22-20N.-2W.
11.	Mobil Oil Corp.	Section 46 Unit No. 2	27-20N.-2W.
12.	Bettie A. McBurney	Reserve McBurney Unit S1 No. 1	26-20N.-2W.
13.	Bettie A. McBurney	Section 52 Unit No. 3	25-20N.-2W.
14.	Neaves Pet. Devel. Co.	Neaves Capital 53-1	32-20N.-1W.
15.	Mobil Oil Corp.	Dano Seco No. 1	33-20N.-1W.
16.	Calif. Time Pet. Inc.	Butte Dano Seco No. 4	34-20N.-1W.
	Wells on figure 7		
1.	Occidental	Zumwalt K-2	12-13N.-1E.
2.	Shell	Strat. test well	31-14N.-2E.
3.	G. E. Kadane & Sons	Lamb No. 1	30-14N.-2E.
4.	Atlantic	AMKH et al	28-14N.-2E.
5.	Atlantic	Continental Stent No. 1	15-14N.-2E.
6.	Kenneth L. Sperry	Shannon No. 1	10-14N.-2E.
7.	Exxon	Shannon No. 1	14-14N.-2E.
8.	Pearson Sibert	Tom No. 1	5-14N.-3E.
	Wells on figure 8		
1.	Sun Oil Co.	SMC Cameron Dougherty No. 1	31-12N.-3E.
2.	McCulloch Oil & Gas Corp.	McCulloch-Magoon et al. No. 1	32-12N.-3E.
3.	Davis Oil Co.	Aileen Marty No. 1	35-12N.-3E.
4.	Davis Oil Co.	Van Dyke No. 1	35-12N.-3E.
5.	Decalta Int. Corp.	Osterli No. 3	31-12N.-4E.
6.	Sun Oil Co.	Lenert No. 55-29	29-12N.-4E.
7.	Kenneth L. Sperry	Davis No. 1	21-12N.-4E.
8.	Plateau Oil & Gas Co.	Van Dyke No. 1	24-12N.-4E.
9.	Exxon	Bonnefeld No. 1	10-12N.-5E.
	Wells on figure 9		
1.	Rocky Mtn. Drilling Co.	Murdoch No. 1	27-7N.-4E.
2.	McCulloch Oil & Gas Corp.	Scribner No. 1	27-7N.-4E.
3.	Chevron	Correa No. 1	24-7N.-4E
4.	Neaves Pet. Devel. Co.	Neaves-Standard Oil Sims No. 3	18-7N.-5E.
5.	Chevron	Sims Comm. No. 2	18-7N.-5E.
6.	Calif. Expln. Co.	Sacramento Unit A	16-7N.-5E.
7.	Union Oil Co. of Calif.	Union Dow Attorney General No. 1	13-7N.-5E.
8.	Intex Oil Co.	Unit plan No. 1	19-7N.-6E.
9.	D. D. Feldman	Unit plan No. 1	8-7N.-6E.
10.	Norris Oil Co.	Waegell No. 1	6-7N.-7E.
	Wells on figure 10		
1.	Shell Oil Co.	Shell-Brevelli No. 1	4-4N.-6E.
2.	Amerada Hess Corp.	Comm. 1-2	2-4N.-6E.
3.	Amerada Hess Corp.	Comm. 1-1	1-4N.-6E.
4.	Aminoil USA, Inc.	Comm. 2-1	12-4N.-6E.
5.	Union Oil Co. of Calif.	V-B-G No. 1	7-4N.-7E.
6.	Amerada Hess Corp.	Lodi Comm. 8-1	8-4N.-7E.
7.	Amerada Hess Corp.	Lodi Comm. 9-1	9-4N.-7E.
8.	Shell Oil Co.	Ferrera No. 1-10	10-4N.-7E.
9.	Ben Owens Drilling Co.	Thompson No. 1	11-4N.-7E.
	Wells on figure 14		
1.	Solar Drilling Co.	Solar-McElroy-Crawford No. 1	5-23N.-1W.
2.	Pacific Western	Cana No. 1	11-23N.-1W.
3.	Exxon	G. U. Roney No. 1	1-23N.-1W.
4.	Exxon	C. C. Baccala No. 1	28-24N.-1E.
	Wells on figure 20		
1.	Neaves Pet. Devel. Co.	Neaves-Cobb No. 1	11-10N.-3W.
2.	Empire Oil and Gas Co.	Porterfield No. 1	18-10N.-2W.
3.	Texaco, Inc.	Esparto c.h. No. 2	17-10N.-2W.
4.	Dow Chemical Co.	Duncan No. 1	9-10N.-2W.
5.	Gulf Oil Corp.	A. B. Stevens No. 1	7-10N.-1W.
6.	Atlantic Oil Co.	Irrigated Valley Land No. 1	4-10N.-1W.
7.	A. A. Hopkins, Jr.	Manler No. 62-3	3-10N.-1W.
8.	C. K. M. Oil Co.	Mast No. 1	35-11N.-1W.
9.	Gulf Oil Corp.	L. M. Bemmerley No. 1	36-11N.-1W.
10.	E. B. Towne	Slaven No. 1	30-11N.-1W.
11.	Arco	Reiff No. 1	29-11N.-1W.
12.	Arco	Hermle No. 1	15-11N.-1E.

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