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Since Bramlette (1946): The Miocene Monterey Formation of California revisited

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INTRODUCTION

For more than a century the Miocene Monterey Formation has fascinated geologists with its uniquely siliceous composition, complex diagenesis, and importance as both source and reservoir of oil in California. The Monterey's extensive and excellent outcrops, exposed at different stages of alteration, have served as laboratories for countless studies of silica, clay, carbonate, phosphate, organic matter, and petroleum. Bramlette's U.S. Geological Survey Professional Paper 212 (1946) served as the foundation for all of these studies and provided the detailed sedimentology, stratigraphy, and petrology to give them context and meaning. For the most part, Bramlette had it right, and an explosion of new research since the 1970s advanced and refined understanding without disproving many of Bramlette's fundamental observations and assertions. One hypothesis that did eventually fall was that abundant siliceous volcanism was the essential source of the silica incorporated in the frustules of diatoms and in the sediment of the Monterey Formation; we have since learned that within zones of intense upwelling, diatoms or radiolarians can extract enough silica from normal seawater to produce highly siliceous sediments when undiluted by other sedimentary components (Calvert, 1966, 1968).

Research since Bramlette's has broadly focused on diagenesis (especially that of silica, carbonate, and organic matter), petroleum generation and reservoirs, dating and stratigraphic correlation, and the oceanographic context of deposition of the Monterey Formation. Much of this work benefited from technological advances in X-ray diffraction, stable isotopic analysis, electron microscopy, and the results of the Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP). A burst of research, initially proprietary, began about 1970 as oil companies sought to explain and exploit major offshore discoveries in the Monterey Formation following the first sale of Federal Outer Continental Shelf leases in the Santa Barbara Channel in 1966. A tremendous amount of this work was published in a series of American Association of Petroleum Geologists (AAPG) and Society of Economic Paleontologists and Mineralogists (SEPM; now the Society for Sedimentary Geology) special publications and symposium volumes in the 1980s and 1990s (Isaacs, 1981a; Garrison and Douglas, 1981; Williams and Graham, 1982; Isaacs and Garrison, 1983; Garrison et al., 1984; Surdam, 1984; Casey and Barron, 1986; Dunham and Blake, 1987; Schwalbach and Bohacs, 1992; Hornafius, 1994a; Eichhubl, 1998) that coincided with the upturn in industry interest in the petroleum potential of the offshore Monterey (Isaacs, 1984; Crain et al., 1985). An additional major volume focusing on the organic geochemistry of the Monterey Formation (Isaacs and Ruellkötter, 1999) should be published by the time this review is published.

GEOLOGIC SETTING

The Miocene Monterey Formation was deposited along the North American plate boundary during the transition of the California margin from a convergent to transform setting (Blake et al., 1978; Barron, 1986a). Resulting tectonic subsidence and landward transgression of the shoreline during the late Oligocene to middle Miocene led to the development of middle bathyal depocenters in which the Monterey sediments accumulated (Figs. 1 and 2) (Ingle, 1980, 1981a). Presedimentary and synsedimentary tectonic deformation (chiefly extension, shearing, and rotation) during the Miocene has been overprinted by Pliocene-Pleistocene shortening, making palinspastic reconstruction of the location and extent of the Neogene sedimentary basins extremely challenging (Ingersoll and Ernst, 1987; Crouch and Suppe, 1993; Fritsche, 1998; Isaacs, 1999). In

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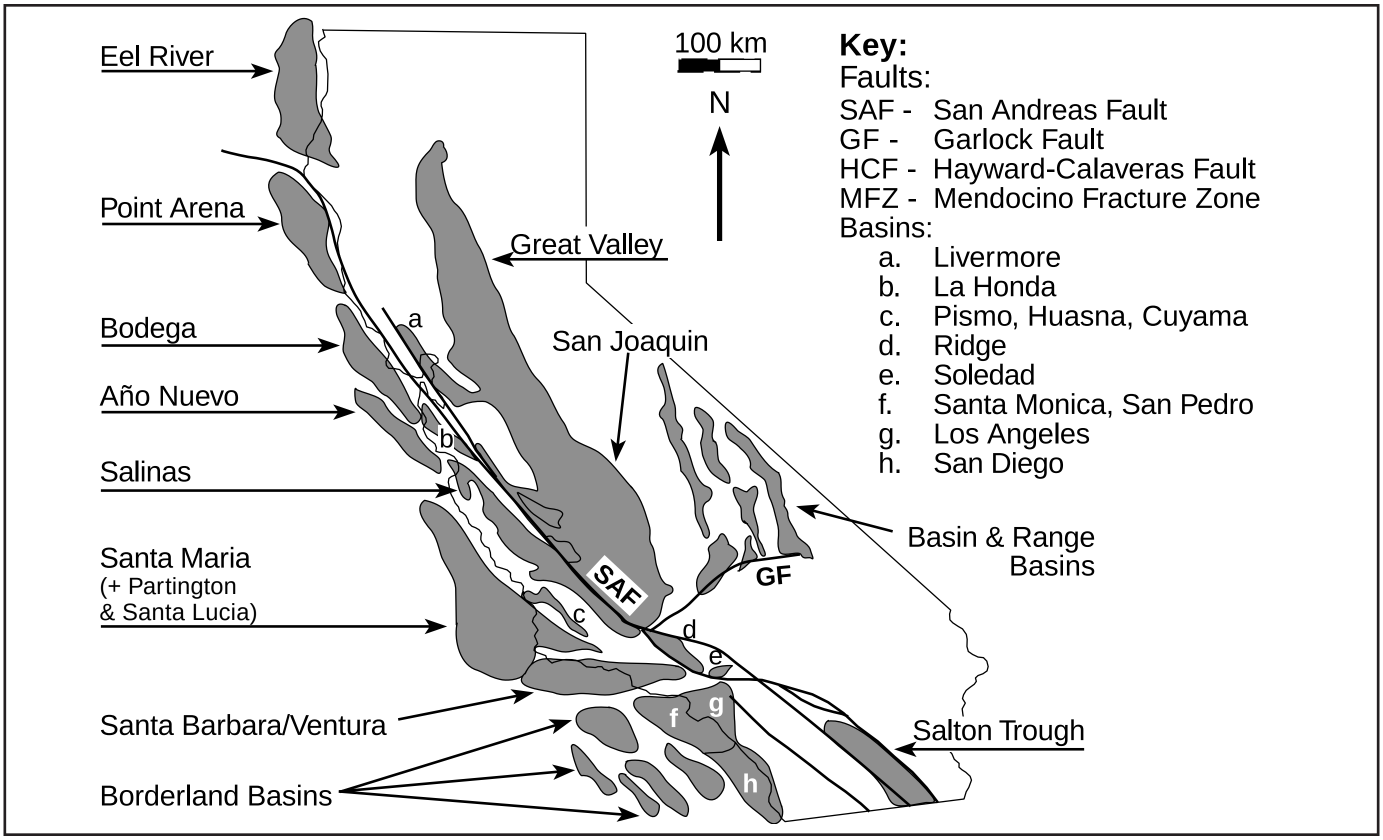

Figure 1. Present location of Neogene depocenters or sedimentary basins (after Biddle, 1991; Dunkel and Piper, 1997).

many cases, the geometry and bathymetry of individual depocenters evolved from the Miocene through the Pleistocene, with earlier deposited sediments, including the Monterey Formation, now forming the folded and faulted flanks of the Pliocene and Quaternary depocenters (Teng and Gorsline, 1989; Blake, 1991).

GEOGRAPHIC EXTENT

The Monterey Formation is part of a discontinuous belt of fine-grained, notably siliceous (diatomaceous) sediments that accumulated around the north Pacific Rim chiefly during the Miocene (ca. 16–4 Ma) (Ingle, 1973, 1980, 1981b). On land, well-studied Monterey strata form extensive outcrops and subcrops in the Coast Ranges and western parts of California (Bramlette, 1946; Pisciotto and Garrison, 1981), extending as an irregular blanket some 1700 km north and south along the continental margin. Offshore equivalents of Monterey siliceous sediments have been cored by deep-sea drilling as far as 300 km seaward from the modern coast and in water as deep as 4200 m (ODP Sites 1010, 1016, 1021) (Lyle et al., 1997). The formation is typically 300–500 m thick on land, but is locally much thinner and thicker (Bramlette, 1946; Isaacs and Petersen, 1987).

AGE OF THE MONTEREY FORMATION

Like most lithostratigraphic units, the age of the middle to late Miocene Monterey Formation varies with location, as sedimentation characteristic of the formation commenced and terminated at different times in separate depocenters. If a typical duration could be specified, it would be from about 16 Ma to 6 Ma (Barron, 1986b). Initiation of Monterey deposition started as early as 17.8 Ma (Saucesian stage, Naples Beach) (DePaolo and Finger, 1991) or as late as 15 Ma (e.g., Relizian, Palos Verdes Hills, Berkeley Hills, Monterey; Obradovich and Naeser, 1981). The youngest Monterey strata at any one location range from about 13 Ma (Luisian, Berkeley Hills) to <5 Ma (Delmontian, Pliocene, Palos Verdes Hills; Woodring et al., 1946; Obradovich and Naeser, 1981). In the Cuyama basin, the Saltos Shale and Whiterock Bluff Shale, often assigned as members of the Monterey Formation (Hill et al., 1958), were apparently deposited entirely before initiation of Monterey sedimentation in the type area (Obradovich and Naeser, 1981).

Biostratigraphy

Microfossils provide the primary basis for biostratigraphy within the fine-grained Monterey Formation, with benthic

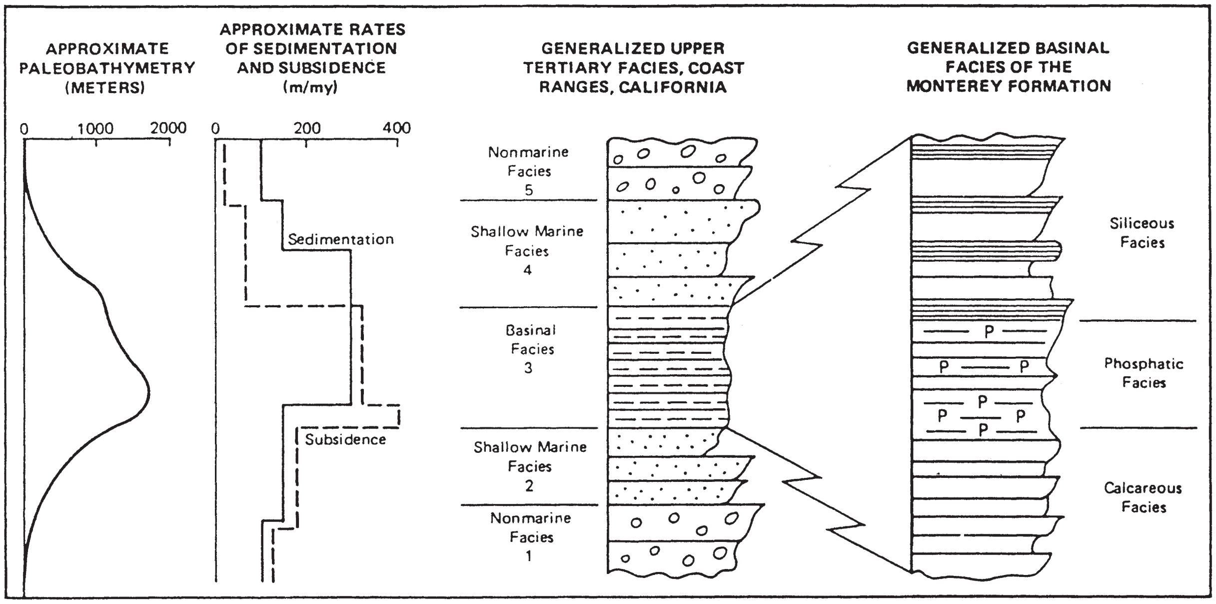

Figure 2. Generalized upper Tertiary sedimentary facies of the California Coast Ranges, showing the position and facies of the Monterey Formation (Pisciotto and Garrison, 1981).

foraminifers remaining the most commonly used taxa for correlation. Because of downsection silica phase transformations and upsection loss or dissolution of carbonate, none of the major biostratigraphic groups are generally useful through the entire formation.

Monterey strata span the late Saucesian, Luisian, Mohnian, and locally, the early Delmontian benthic foraminiferal stages of California (Kleinpell, 1938, 1980). Since development of these Neogene stages, however, it has become evident that benthic assemblages were influenced by local paleobathymetry, character of the impinging water mass, and benthic sedimentology, making them time transgressive and often provincial in nature (Crouch and Bukry, 1979; Ingle, 1980; Obradovich and Naeser, 1981). Although still quite useful within individual fields or basins because of their abundance (Finger, 1995; Blake, 1991), benthic foraminifers had to be integrated with planktonic foraminifers (Keller and Barron, 1981), diatoms (Barron, 1986b; Barron and Isaacs, 1999), nannofossils (Poore et al., 1981), magnetostratigraphy (Omarzai, 1992), radiometric geochronology (Obradovich and Naeser, 1981), and chemostratigraphy (DePaolo and Finger, 1991; Flower and Kennett, 1993, 1994).

LITHOLOGY AND COMPOSITION

The Monterey Formation is distinguished by its overall highly biogenic composition, in which the average contributions of silica (chiefly the tests and frustules of diatoms and radiolarians), carbonate (coccoliths and foraminifers), organic matter (mostly type II kerogen, marine algae) and their diagenetic equivalents greatly exceed those of other Neogene fine-grained sedimentary units (Isaacs, 1985; Isaacs and Petersen, 1987). Although the highly diatomaceous and organic-rich deposit has been interpreted to record unusually great planktonic productivity along the eastern Pacific margin (Barron, 1986a; Ingle, 1980, 1981b; Pisciotto and Garrison, 1981), mass accumulation rates show that the purest biogenic intervals reflect decreased terrigenous input, and consequently less dilution of the biogenic component (Isaacs, 1985, 1999). Overall, the Monterey Formation records sediment starvation in conjunction with surface productivity associated with upwelling along the California Current system. These sedimentary conditions increased the relative proportions of silica, organic matter, phosphate, or carbonate with respect to fine-grained detritus—mainly illite-smectite mixed-layer clay minerals, feldspars, and quartz (Isaacs, 1980; Pollastro, 1990; Compton, 1991). Periods of extremely slow pelagic sedimentation, undiluted by much fine-grained detritus and during which most of the primary biogenic hard components (SiO_2 or $CaCO_3$) dissolved or winnowed away, resulted in the extreme organic richness characteristic of some condensed intervals (e.g., the carbonaceous marl–phosphatic shale facies of the Santa Barbara coastal area) (Isaacs, 1985).

Bramlette (1946) described the typical Monterey lithologies—diatomite, diatomaceous and siliceous mudrocks, porcelanite, chert, calcareous and phosphatic mudrocks, dolostone, and limestone—in remarkable completeness and detail. He also recognized the significance of graded, clastic to biogenic "rhythmites" before the importance of fine-grained turbidites in deep water was generally understood or accepted. Where clastic siltstone and sandstone are common, they are usually assigned to another formation or to a distinct member of the Monterey (e.g., Point Sal Formation, Santa Maria basin, or the Stevens Sands, southern San Joaquin basin) (Williams and Graham,

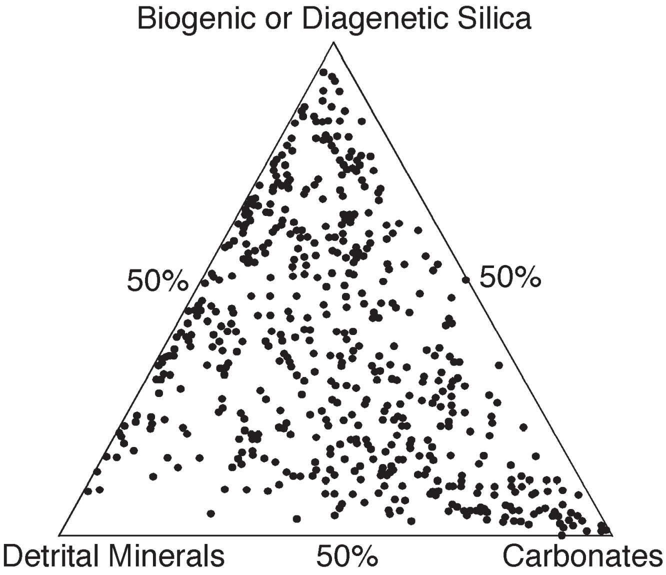

Figure 3. Diagram showing the wide range of sedimentary compositions in the Monterey Formation of the Santa Maria and Santa Barbara basins (Isaacs, 1985).

1982). Conglomerates are even more rare (Garrison and Ramirez, 1989). At scales from less than 1 mm to hundreds of meters, the lithologies of the Monterey are characterized by great compositional variability, making any individual sample usually unrepresentative of its own stratigraphic interval (Fig. 3) (Isaacs, 1985). Compositional variation is expressed by rhythmic alternation of clastic-biogenic, massive-laminated, or diagenetically distinct lithologic cycles (Pisciotto and Garrison, 1981; Govean and Garrison, 1981; Isaacs, 1985). Even with such lithologic variation, large-scale trends in average composition, both vertically and laterally, are relatively consistent within individual regions, giving rise to a number of local stratigraphic subdivisions into informal members (e.g., Canfield, 1939; Woodring et al., 1943; Foss and Blaisdell, 1968; Isaacs, 1981b, 1983; Pisciotto and Garrison, 1981; MacKinnon, 1989a). Although there is a broad similarity to some of the compositional trends—for example, the widespread occurrence of middle Luisian to Relizian calcareous mudstones and late Luisian to early Delmontian diatomaceous sediments—member-scale facies are clearly time transgressive when compared between regions (Blake, 1981; White, 1989; Hornafius, 1991, 1994b; Schwalbach and Bohacs, 1995).

DEPOSITIONAL ENVIRONMENTS

The Monterey Formation was chiefly deposited in lower middle bathyal (1500–2300 m) to upper middle bathyal (500–1500 m) environments (Fig. 2) (Ingle, 1973, 1980; Isaacs, 1999), which shallowed upward in most sequences. Preservation of organic matter, abundance of fine varve-like laminations, and presence of dysaerobic benthic foraminifers indicate that the Monterey was commonly deposited in or associated with an oxygen-deficient environment. Consequently, likely depositional environments for the Monterey include basin plains, slopes, banktops, and shelf edges where they intersect or are influenced by the mid-water oxygen minimum zone (Calvert, 1966; Garrison et al., 1979; Lagoe, 1981; Pisciotto and Garrison, 1981). Possible modern analogues for these settings occur beneath upwelling zones associated with the Southern California Borderland, the Gulf of California, and the Peru and Pakistan margins (Calvert, 1966, 1968; Donegan and Schrader, 1981; Soutar et al., 1981). Although the silled basins of the California Continental Borderland have been most frequently cited as present-day examples, there is little direct evidence for the existence of such steep-sided, silled basins during deposition of the Monterey (Isaacs, 1999).

Thin, millimeter-scale laminations are only intermittently present in the Monterey Formation. They are rare in the predominantly massive and thin- to thick-bedded lower portion of the Monterey, and become increasing prevalent upsection (Mohnian stage), while remaining rhythmically or irregularly interbedded with massive (bioturbated or redeposited) strata (Pisciotto and Garrison, 1981; Govean and Garrison, 1981; Isaacs, 1985; Ozalas et al., 1994). Such alternation suggests continuously fluctuating levels of paleo-oxygenation during deposition (Behl and Kennett, 1996). The overall upward increase in lamination indicates either that bottom water was progressively (if inconsistently) depleted of oxygen through time or that the Monterey Formation depositional environment shoaled into the heart of the mid-water oxygen minimum zone with progradation of the Miocene California margin (Isaacs et al., 1996).

DIAGENESIS

The highly reactive biogenous components of the Monterey Formation (i.e., opaline silica, calcite, phosphate, and organic matter) have undergone a complex paragenetic sequence of alteration with time, burial, and tectonic deformation. Although it is simpler to examine mineralogic and chemical systems in isolation, every stage of dissolution, precipitation, or alteration influenced simultaneous and subsequent reactions by altering pore-water chemistry, water-rock ratios, and permeability (Kastner et al., 1984; Eichhubl and Behl, 1998). Diagenetic modification by chemical migration can enhance or suppress the physical and compositional contrasts that already existed in the originally heterogeneous Monterey sediments, making it a wonderfully complicated unit to work with (Pisciotto and Garrison, 1981; Govean and Garrison, 1981; Grivetti, 1982; Murray and Jones, 1992; Behl, 1992).

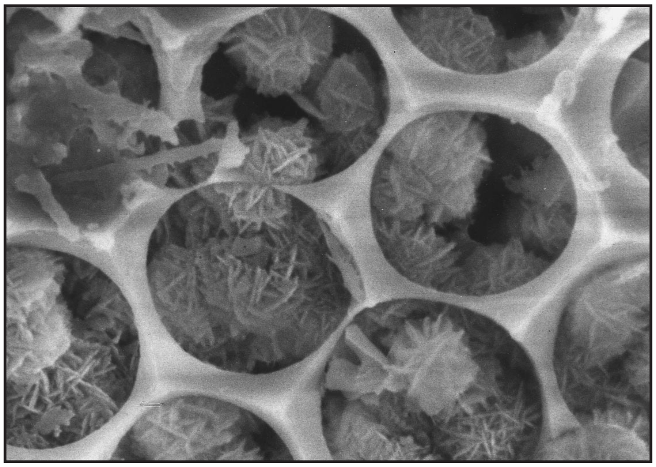

Figure 4. Scanning electron micrograph of nascent opal-CT lepispheres growing within a partially dissolved opal-A test of a diatom. Field of view = 15 µm.

Silica

Although Bramlette (1946) clearly documented the alteration of soft diatomaceous sediments to hard porcelanite and chert, we now know considerably more about the nature, controls, and distribution of silica diagenesis. The sequence of mineralogic alteration involves two steps of complete dissolution and reprecipitation. The first is from biogenic opal-A (hydrous silica that is crystallographically amorphous to X-ray diffraction) to diagenetic opal-CT (hydrous silica composed of interlayered cristobalite and tridymite) (Fig. 4). The second is from opal-CT to diagenetic quartz (generally cryptocrystalline, microcrystalline, or chalcedonic quartz). Transformation is controlled by temperature and burial depth (Murata and Nakata, 1974; Murata and Larson, 1975; Isaacs, 1981c; Pisciotto, 1981a), bulk composition (Isaacs, 1982; Behl and Garrison, 1994), and rock properties, such as porosity and permeability (Behl, 1998; Eichhubl and Behl, 1998). Within sediments of common compositions for the Monterey Formation (i.e., diatomaceous or siliceous mudstones and porcelanites), silica phase conversion takes place within two relatively narrow temperature ranges and burial depths (~40–50 °C and ~0.5–2 km for opal-A to opal-CT and ~65–80 °C and ~1.5–3 km for opal-CT to quartz; Fig. 5) (Pisciotto, 1981a; Keller and Isaacs, 1985). Within an individual stratigraphic sequence, however, the silica phase transformation may not be abrupt, but can occur across a broad transition zone, to 300 m thick, of interbedded lithologies containing different silica phases (Fig. 6) (Isaacs, 1982).

The stratigraphic co-occurrence of silica phases with different thermal stabilities and solubilities is explained by compositionally controlled variation in the kinetics of the phase transformations, in which the opal-A to opal-CT transition is retarded and the opal-CT to quartz transition is accelerated in more detrital- or clay-mineral–rich sediments (Kastner etal., 1977; Isaacs, 1981c, 1982; Williams et al., 1985). The purest siliceous sediments undergo diagenesis even earlier than predicted (Bohrmann et al., 1994), with hard, brittle opal-CT and quartz cherts forming at temperatures as low as 2–33 °C and 36–76 °C, respectively (Fig. 5) (Behl, 1992; Behl and Garrison, 1994). On a larger scale, boundaries between silica phase zones are locally discordant to stratigraphy, reflecting lateral variation in sediment accumulation and burial depth, geothermal gradient, or tectonic deformation (Figs. 6 and 7) (Bramlette, 1946; Murata and Larson, 1975; Pisciotto, 1981a). Within each diagenetic zone, silica becomes increasingly well ordered with depth, temperature, or time, even though there may not be any lithologic change. Opal-A becomes less soluble as higher surface area diatoms dissolve and smaller submicroscopic mineralogic domains give way to larger ones (Williams et al., 1985). Ordering of opal-CT is revealed by decreased spacing of the d_{101} lattice planes (Murata and Nakata, 1974; Murata and Larson, 1975; Cady et al., 1996) and increased crystallite size with growth (Grivetti, 1982; Williams et al., 1985; Behl and Meike, 1990). Progressive growth of crystallite domains in diagenetic quartz is shown by the height and sharpness of X-ray diffraction peaks in the quartz crystallinity index (Murata and Norman, 1976).

Complete dissolution and reprecipitation at the two silica phase transitions produce dramatic changes in the physical properties of the sediment as the rigid, but porous framework collapses, or as internal pore spaces are filled with

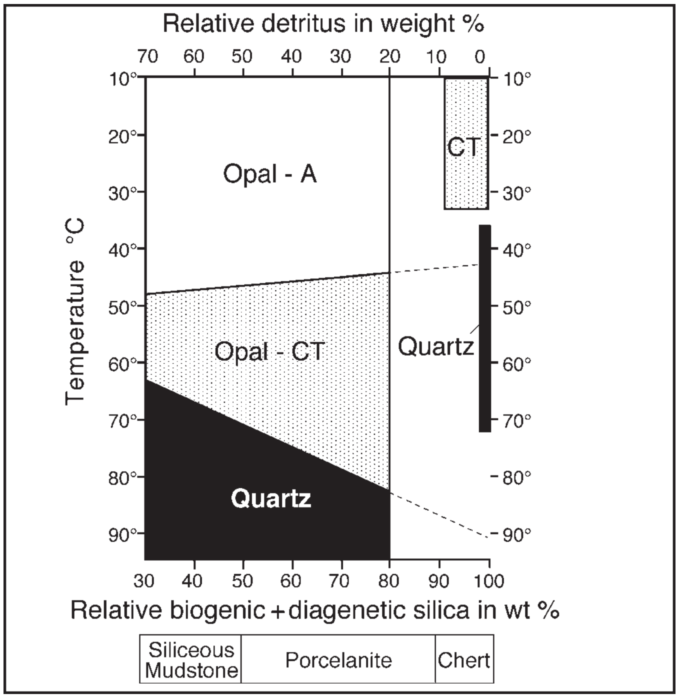

Figure 5. Diagram showing the relative timing and temperatures of silica phase changes (Keller and Isaacs, 1985), modified to include data on the purest diatomites and cherts (Behl, 1992; Behl and Garrison, 1994).

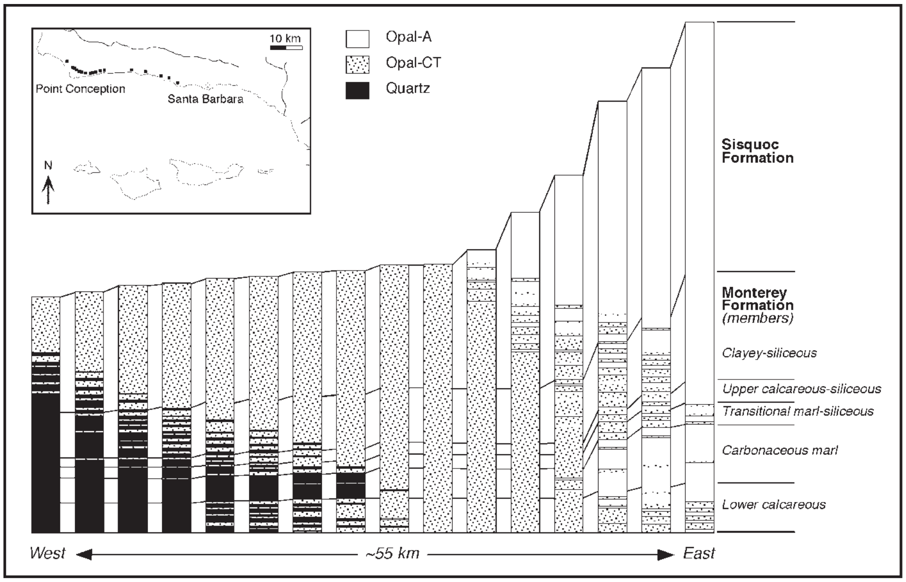

Figure 6. Schematic view of Santa Barbara coastal area, showing silica phase zones cutting across lithostratigraphic boundaries, the interbedded nature of the transition zones, and typical compaction with increased physical and chemical diagenesis (modified from Isaacs, 1981).

silica cement (Fig. 4). These abrupt changes in bulk density can be imaged locally by seismic methods as extensive cross-cutting reflectors in the subsurface (Fig. 7) (Mayerson and Crouch, 1994) and are associated with the expulsion, migration, and trapping of hydrocarbons as well as the potential for forming fractured petroleum reservoirs (McGuire et al, 1983; MacKinnon, 1989b; Mayerson et al., 1995).

Carbonate

Carbonate diagenesis in the Monterey Formation has been studied by a wide variety of geochemical, isotopic, and sedimentological means to determine its paragenesis with organic matter and silica (Murata et al., 1967, 1972; Friedman and Murata, 1979; Pisciotto, 1981b; Garrison et al., 1984; Burns and Baker, 1987; Malone et al., 1996; Eichhubl and Boles, 1998). Although primary carbonate components are mainly calcitic coccoliths and foraminifers, the dominant secondary carbonate phase in the Monterey is calcium-rich dolomite, whether occurring as finely disseminated rhombs, cross-cutting veins, or as tightly cemented concretions and beds (Pisciotto, 1981b). Dolomite forms in anoxic or dysoxic conditions related to the diagenesis of organic matter, within or below the zone of sulfate reduction (Pisciotto and Mahoney, 1981). Low sedimentation rates during early burial diagenesis tend to increase the concentration of dolomite by providing better conditions for continued precipitation in the zone of sulfate reduction (Pisciotto and Mahoney, 1981; Burns and Baker, 1987).

Phosphate

Diagenetic sedimentary phosphate (cryptocrystalline carbonate fluorapatite) forms chiefly with the shallow degradation of organic matter, probably via a number of physical, chemical, and biological mechanisms (Garrison et al., 1990; Föllmi et al., 1991). Most carbonate fluorapatite precipitation occurs within a few tens of centimeters of the sediment surface and during slow sedimentation or depositional hiatuses (Garrison et al., 1994). The most prominent phosphatic facies in the bathyal Monterey Formation are laminated, organic-rich phosphatic marlstones that developed as the condensed residue of slowly deposited, calcareous-siliceous muds and oozes (Garrison et al., 1987) during sediment starvation (Isaacs, 1985, 1999). In this facies, carbonate fluorapatite occurs as small nodules, lenses, laminations, and peloids that formed in place with little or no subsequent reworking. Shelfal and banktop phosphoritic sands occur interbedded with hemipelagic sediments, and consist mostly of phosphatic peloids (Garrison et al., 1987, 1994). Conglomerates and hardgrounds composed of dense, dark phosphatic pebbles, nodules, and concretions are less common in the Monterey Formation, but record repeated episodes of phosphatization, exhumation, winnowing, and reworking by currents, slumping, and sea-level change (Föllmi et al., 1991; Garrison et al., 1994).

SOURCE OF PETROLEUM

The Miocene Monterey Formation is widely considered to be the primary source rock for hydrocarbons in California (Woodring and Bramlette, 1950; Crawford, 1971; Taylor, 1976; Lillis and Lagoe, 1983; Isaacs and Petersen, 1987). Total organic carbon (TOC) in the Monterey can be as high as 23% (34% organic matter by weight), but averages between 2% and 5%, with large sample to sample variation, depending on lithology and depositional setting (Isaacs and Petersen, 1987).

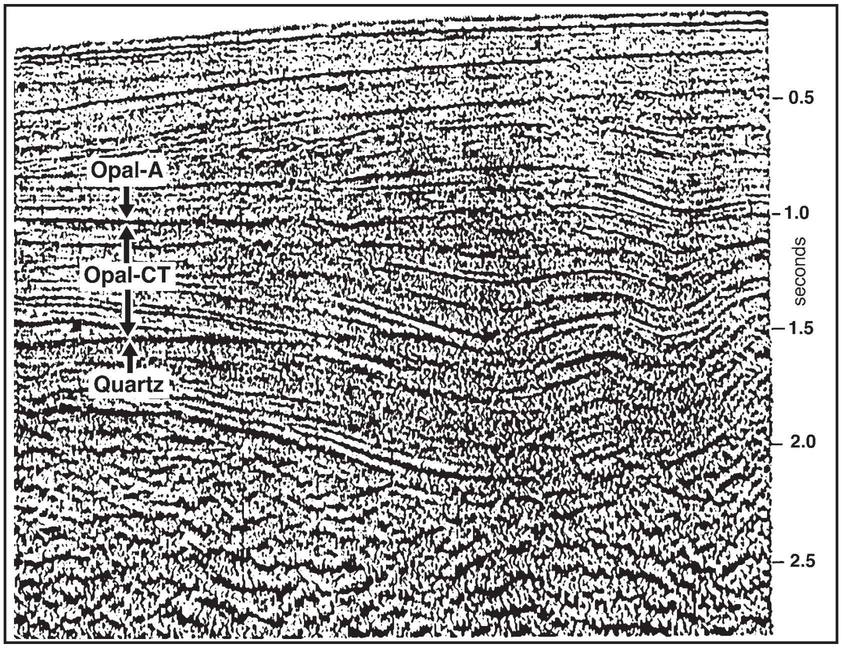

Figure 7. Seismic-reflection profile showing the near-horizontal opal-A to opal-CT and opal-CT to quartz silica phase transitions that cut across stratigraphy. After Crouch, Bachman, and associates, 1991.

Organic matter is overwhelmingly amorphous marine algal debris, but locally includes significant portions of terrestrial origin (Isaacs and Magoon, 1984; Graham and Williams, 1985). Biomarkers in both Monterey oil and rocks also indicate that the organic matter is largely marine (King and Claypool, 1983; Curiale et al., 1985). Kerogens in the highly biogenic Monterey sediments are mostly sulfur-rich, oil-prone type II-S (Surdam and Stanley, 1981; Kruge, 1983; Orr, 1986; Isaacs, 1988; Ruellkötter and Isaacs, 1996).

Much of the oil sourced in the Monterey was generated in rocks considered to be immature or marginally mature by conventional methods of assessment, for example, vitrinite reflectance (R_o <0.4), thermal alteration index (TAI <2.3), Rock-eval pyrolysis (T_{max}, variable and problematic), sapropel fluorescence, hydrogen/carbon ratios, and silica diagenetic grade (Taylor, 1976; McCulloh, 1979; Kablanow and Surdam, 1984; Global Geochemistry Corporation, 1985; Petersen and Hickey, 1987; Ruellkötter and Isaacs, 1996), although some of these indicators may not be reliable indicators of maturity in Monterey-type rocks (Walker et al., 1983). Initiation of catagenesis as early as 60–80 °C is likely related to the high sulfur content (up to 9% by weight) of the kerogen and the weakness of its carbon-sulfur bonds (Hunt, 1979; Orr, 1984, 1986; Isaacs and Petersen, 1987). The generally low API gravity (<20 API°) of Californian oil is also related to early generation, low maximum temperatures, and bacterial degradation, both as organic matter and as hydrocarbons (Petersen and Hickey, 1987; Ruellkötter and Isaacs, 1996). The co-occurrence of both in situ kerogen and migrated hydrocarbons within the rock matrix presents difficulties in assessing the true maturity of source rocks in the Monterey Formation as well as the relative contributions of oil from adjacent or distant (deeper) sources within the formation (Dunham et al., 1991).

Although much of the Monterey Formation has sufficiently high TOC and H/C ratios to be classified as good oil-prone source rock, a proportionally large amount of the oil may come from organic-rich carbonaceous marl (phosphatic shale) strata (Orr, 1984; Dunham et al., 1991; Isaacs and Ruellkötter, 1999) at whatever stratigraphic level and location it is best developed.

PETROLEUM RESERVOIRS

The Monterey Formation is unusual in that it is both source and reservoir of oil (Crawford, 1971; Isaacs and Petersen, 1987). Typically, fine-grained organic-rich rocks lack the effective porosity and permeability to provide commercial petroleum reservoirs. Consequently, petroleum reservoirs generally consist of either adjacent or interfingered sandstone beds, members, or formations, or they consist of naturally fractured, brittle diagenetic siliceous and dolomitic rocks. Oil is also locally produced from highly porous, opal-A diatomite in western parts of the San Joaquin basin through natural and artificially induced fractures.

The high diagenetic potential of the Monterey's fine-grained components (chiefly of silica, carbonate, and organic matter), diagenetic embrittlement (of chert, porcelanite, and dolomite) with burial, and location in a tectonically active set-

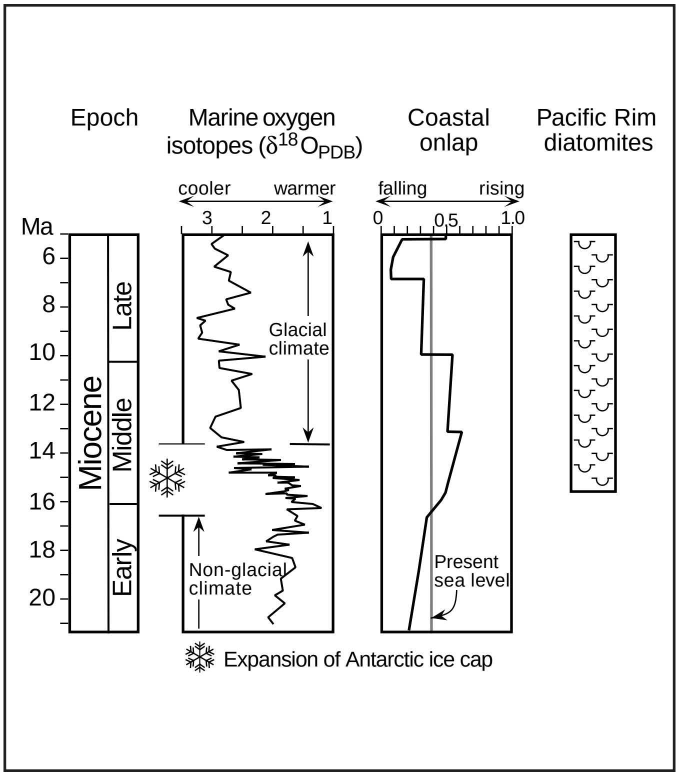

Figure 8. Global Miocene climatic and eustatic events and occurrence of Miocene diatomites in bathyal sequences around the north Pacific margin (modified from Ingle, 1981b). PDB—Peedee belemnite.

ting combined to create many highly fractured or brecciated oil reservoirs in the subsurface (Regan and Hughes, 1949; McGuire et al., 1983; Dunham and Blake, 1987; MacKinnon, 1989b; Eichhubl and Behl, 1998). Depending on original depositional constraints, different lithologies may make up the important fractured reservoirs in individual fields. Whereas fractured siliceous shale and porcelanite provide important production in the San Joaquin basin, chert and dolomite breccias form the most important reservoirs in the onshore and offshore Santa Maria basin (Redwine, 1981; Roehl, 1981; McGuire et al., 1983; Crain et al., 1985; Dunham et al., 1991). In all cases, fractures are critical for fluid flow in the otherwise extremely low permeability (<1 md) Monterey lithologies. The distribution and density of fractures vary with rock type, diagenetic grade, bed thickness, location on tectonic structures, and the regional stress field (Snyder et al., 1983; Belfield et al., 1983; Snyder, 1987; MacKinnon, 1989b; Narr, 1991; Gross, 1993; Gross et al., 1997; Finkbeiner et al., 1997) and are also related to large-scale faulting (Eichhubl, 1997).

In addition to microscopic and macroscopic fracture porosity and permeability, most highly siliceous rocks have substantial (10%–35%) matrix porosity (Isaacs, 1981d), which can form the major part of reservoir storage, but also contributes to a complex production behavior.

While much of the oil generated in the Monterey is produced from associated or overlying clastic reservoirs, fractured reservoirs are locally very important. For example, Monterey fractured reservoirs account for ~75% of the oil produced in the Santa Maria area (Crawford, 1971). In the most recent assessment of hydrocarbon resources of the Pacific Outer Continental Shelf region, fractured Monterey Formation or equivalent strata are estimated to contain more than one-half of the undiscovered conventionally recoverable oil (5.96 billion barrels) and more than one-third of the undiscovered conventionally recoverable gas (6.32 trillion cubic feet) for a total of 7.08 billion barrels of oil equivalent (Dunkel and Piper, 1997).

PALEOCEANOGRAPHIC AND PALEOENVIRONMENTAL SIGNIFICANCE

Deposition of the Monterey and its equivalents coincided with or followed major changes in Miocene ocean circulation, global climate, and tectonics (Ingle, 1980, 1981b; Pisciotto and Garrison, 1981; Vincent and Berger, 1985; Barron, 1986a). The diatomaceous and organic-rich Monterey sediments were deposited following a major switch in marine thermohaline circulation into approximately the modern configuration where deep water that forms in the North Atlantic and circum-Antarctic regions principally upwells in the Pacific and Indian Oceans (Kennett, 1977; Keller and Barron, 1983; Woodruff and Savin, 1989). Monterey deposition also encompassed the important middle Miocene cooling step in which the Southern Hemisphere cryosphere expanded into western Antarctica (Fig. 8) (Kennett, 1977; Miller et al., 1987). Regional intensification of upwelling and increased affinity with higher latitude assemblages in the late Miocene is indicated by most planktonic taxa (Ingle, 1973, 1981b; Weaver et al., 1981; Barron, 1986a). The co-occurrence of all these events in the middle to late Miocene has led many to attribute or relate the character of Monterey deposits to this important reorganization of the Earth's cryosphere-hydrosphere-atmosphere system, both as cause and as effect (Ingle, 1981b; Pisciotto and Garrison, 1981; Vincent and Berger, 1985; Barron, 1986a). In particular, middle Miocene accumulation of organic matter in marine sediments was great enough to perturb the carbon balance of the global ocean and atmosphere and produce a prominent positive excursion in carbon isotopes that has been recognized in deep-sea and Monterey sequences (Vincent and Berger, 1985; Compton et al., 1990; Flower and Kennett, 1993;,1994; Raymo, 1994). Although the accumulation of organic carbon in the Monterey Formation alone was probably insufficient to account for this shift, the Monterey was clearly deposited within the context of an important transition in Cenozoic cooling associated with cryospheric expansion, thermohaline circulation reorganization (Fig. 8), and possibly accelerated flux of nutrients to the ocean related to Himalayan uplift (Richter et al., 1992). The widespread lower calcareous mudstone facies of the Monterey was largely deposited during an interval of early to middle Miocene gradual warming. The phosphatic and organic-rich facies correlate with a middle Miocene sea-level rise and highstand that occurred prior to expansion of the Antarctic ice sheet (Pisciotto and Garrison, 1981), thus are in effect, condensed, "transgressive shales" (Isaacs, 1999).

Recently, major member-scale stratigraphic shifts in bulk composition in the Monterey have been reinterpreted to reflect shoaling and shoring of the Monterey depositional environment as part of a prograding margin, modified by eustatic sea-level changes, rather than regional or global changes in paleoceanography and climate (Isaacs et al., 1996; Isaacs, 1999). In this model, the time-transgressive nature of major compositional lithofacies reflects proximity to loci of coastal or banktop upwelling, sources of terrigenous detritus, as well as periods of sediment starvation (Isaacs et al., 1996; Isaacs, 1999). For example, the generally most siliceous middle to upper members of the Monterey (late Miocene, Mohnian stage) are interpreted to reflect deposition within the direct influence of shallow (~500 m or less) coastal (~20 km from shore) upwelling or bathymetrically induced upwelling, such as that adjacent to shallowly submerged banks. This interpretation is difficult to reconcile, however, with the presence of highly diatomaceous middle to late Miocene deposits in offshore locations from Baja California to the Oregon border that were deposited and remain at middle to lower bathyal depths and are >100 km away from the modern prograded shoreline (Ingle, 1973, 1980; Barron, 1986a, 1986b; Blake, 1981; Lyle et al., 1997). The wide spatial distribution of the important and unusual Monterey-type deposits likely reflects the unique co-occurrence of paleoceanographic, paleoclimatic, and tectonic events during the Miocene epoch.

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