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ORIGIN OF THE MIOCENE MONTEREY FORMATION IN CALIFORNIA

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INTRODUCTION

Diatomaceous rocks and their diagenetic equivalents, chert, porcelanite and siliceous mudstone are abundant in Miocene deposits of the Pacific region. Of these, the Monterey Formation is the best known and most extensive. It is present throughout large parts of the state of California (Figure 1) and is in places over three kilometers thick. In addition, it includes the largest diatomite quarry in the world and more importantly, is the source and reservoir for much of the oil produced in California (Figure 2).

The origin of Miocene siliceous deposits in the Pacific region is tied to a fortuitous combination of tectonic, climatic and oceanographic conditions that promoted biogenic productivity while reducing terrigenous input to coastal basins. This resulted in rapid accumulation of relatively undiluted biogenic sediment, not only siliceous but calcareous as well, during the Miocene times.

Early work on the Monterey was dominated by Bramlette (1946) who documented the extent of the formation and recognized its basic origin. This work still remains as the most detailed statewide stratigraphic summary.

Following Bramlette's work, surprisingly little work was done on the Monterey Formation until the early 1980's when a flood of publications began to appear. The increased interest was due in part to new analytical methods that were capable of extracting some of the detailed paleoclimatic and oceanographic information that some Monterey Formation rocks contain, and in part due to a strong surge in petroleum exploration for Monterey Formation fractured reservoirs. Much of the new work is included in symposium or guidebook volumes edited by Garrison and Douglas (1981), Issacs (1981a), Williams and Graham (1982), Isaacs and Garrison (1983), Garrison and others (1984), and Casey and Barron (1986). More recent work with up to date reference lists include: Graham and Williams (1985) on the San Joaquin Valley; Barron (1986) on dating and paleoceanography; Garrison and others (1987 and in press) on phosphatic rocks and paleoceanography; Isaacs and Petersen (1987) on petroleum geology; and Snyder (1987) on structure.

GEOLOGIC SETTING

California has been part of a subduction or transform plate margin since the Late Paleozoic. Subduction appears to have been the dominant mode at least from the Late Jurassic to Mid Tertiary times. In Late Oligocene times (ca 29 Ma) subduction began to cease due to the collision of a spreading ridge with the California margin trench (Atwater, 1970).

The change from a subduction to a transform margin resulted in a dramatic change of depositional patterns in California that helped set the stage for Monterey Formation deposition. In the Late Oligocene-Early Miocene, the California margin experienced rapid downdropping and the development of deep marine coastal basins (Tennyson, 1986; Bachman, 1988). This event is documented by a transgressive sequence (non-marine to shallow-marine to deep-marine) that is present throughout the Coast and Transverse Ranges of California (Pisciotto and Garrison, 1981; Ingle, 1981). It is not clear whether the initial downdropping was related wholly to transform tectonics or to some aspect of the trench-transform transition. Regardless, extension was the dominant mechanism as evidenced by normal faulting and vulcanism which occurred throughout much of California in Early Miocene times (Yeats, 1987).

Clastic sedimentation dominated the early phases of fill in the newly formed basins. Then approximately 17.5 Ma, Monterey deposition began abruptly throughout much of California. The major controlling factor was an abrupt and major reduction in the supply of terrigenous material to reach the offshore areas and a high biogenic sedimentation rate (Isaacs, 1985; Graham and Williams, 1985).

The sudden reduction in sediment supply was apparently caused by tectonic, and possibly eustatic events (Figure 3). A major tectonic reorganization of the borderland, at least in Southern California, apparently occurred around the onset of Monterey deposition. Evidence for this includes the beginning of rotation of parts of the Transverse Ranges and Mojave area, clear evidence of transform faulting, rapid downdropping and uplift in coastal areas, and local volcanism, all occurring approximately 16 to 18 Ma (Tennyson, 1986, Crowell 1987; Luyendyk and Hornafius, 1987). By Middle Miocene times, the offshore area was probably composed of a number of small, deep (1000-2000 m; 3300-6600 ft) basins similar to the Southern California borderland today. Outer basins were protected from sediment influx by trapping of coarse clastics in inner basins. Onshore, considerable sediment was trapped in newly formed interior basins (Figure 1). In addition, a high eustatic sea level during the Early and Middle Miocene may have caused flooding of coastal areas, thereby expanding shelf areas where additional sediment could be trapped.

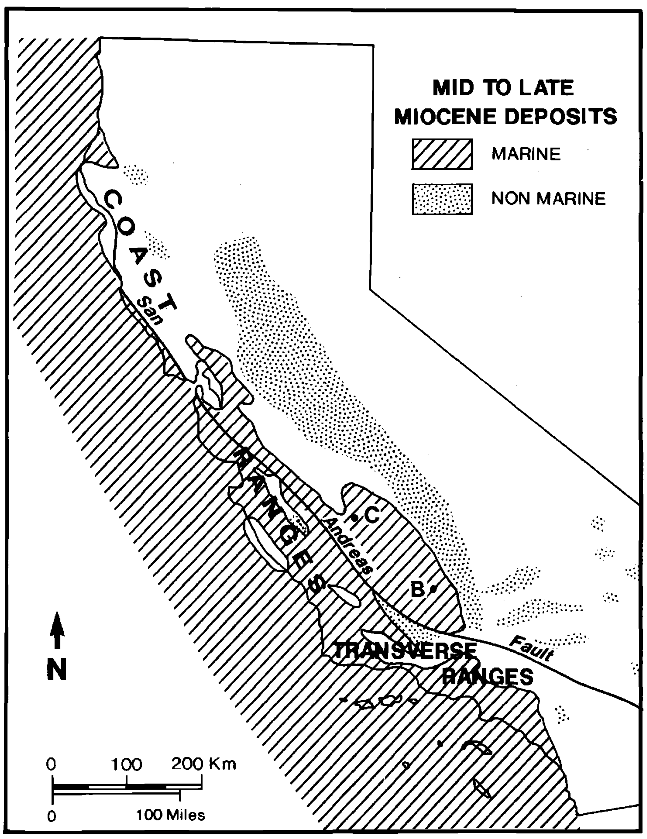

Figure 1. Present generalized location of Late Miocene marine and nonmarine deposits in California. Modified from Reed (1933) and Addicott (1968).
C = Coalinga; B = Bakersfield

In addition to reduced sediment supply, high productivity of siliceous organisms in the Pacific region was an important factor promoting Monterey deposition. Worldwide siliceous sedimentation apparently switched from low latitude North Atlantic areas to the North Pacific approximately 17-18 Ma due to a change in ocean circulation (Keller and Barron, 1983). Furthermore, major polar cooling began about 15 Ma (Woodruff and others, 1981); this probably resulted in increased ocean circulation and upwelling which enhanced productivity, particularly of diatoms. Average silica accumulation rates during the Miocene are comparable with present rates in highly productive oceanic areas (Isaacs, 1985). High ocean productivity and low terrigenous input continued into the Late Miocene resulting in thick accumulation of hemipelagic calcareous

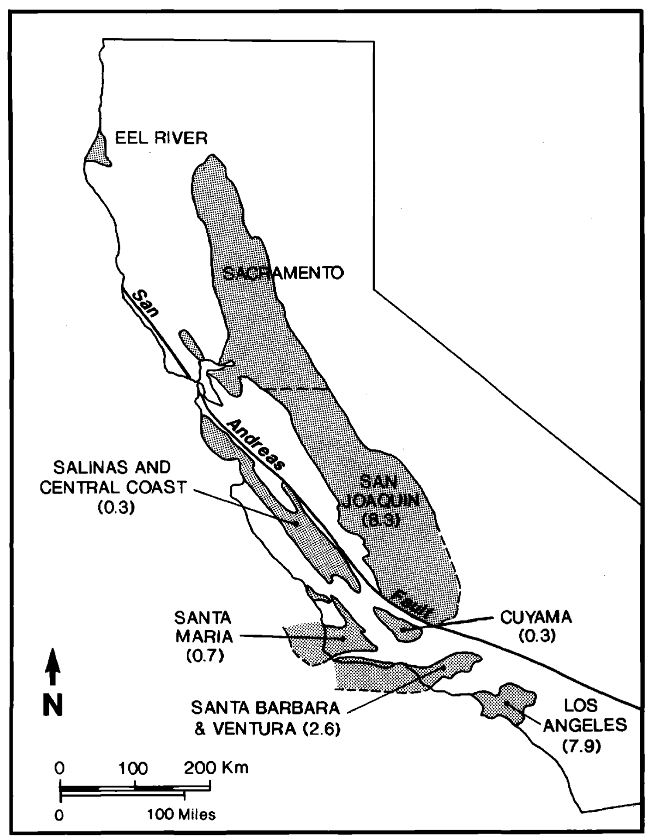

Figure 2. Location of major onshore Neogene depocenters (basins) in California. The Monterey Formation is present in parts or all of each basin (compare Figure 1). Numbers indicate cumulative oil production in billions of barrels; approximately 90% was produced from either the Monterey Formation or clastic rocks of Middle Miocene through Pliocene age (Taylor, 1976). The Monterey Formation is the source of most of the oil.

and siliceous deposits throughout much of California.

Approximately 6 million years ago, a sudden influx of terrigenous material signaled the end of Monterey deposition in most areas of California. This influx was mainly due to the onset of compression and mountain building throughout most of coastal California and may have also been influenced by falling sea level. The Transverse Ranges and part of the Coast Ranges were initially uplifted at this time, while coastal basins continued to subside and fill, setting the scene for Monterey hydrocarbon generation and entrapment.

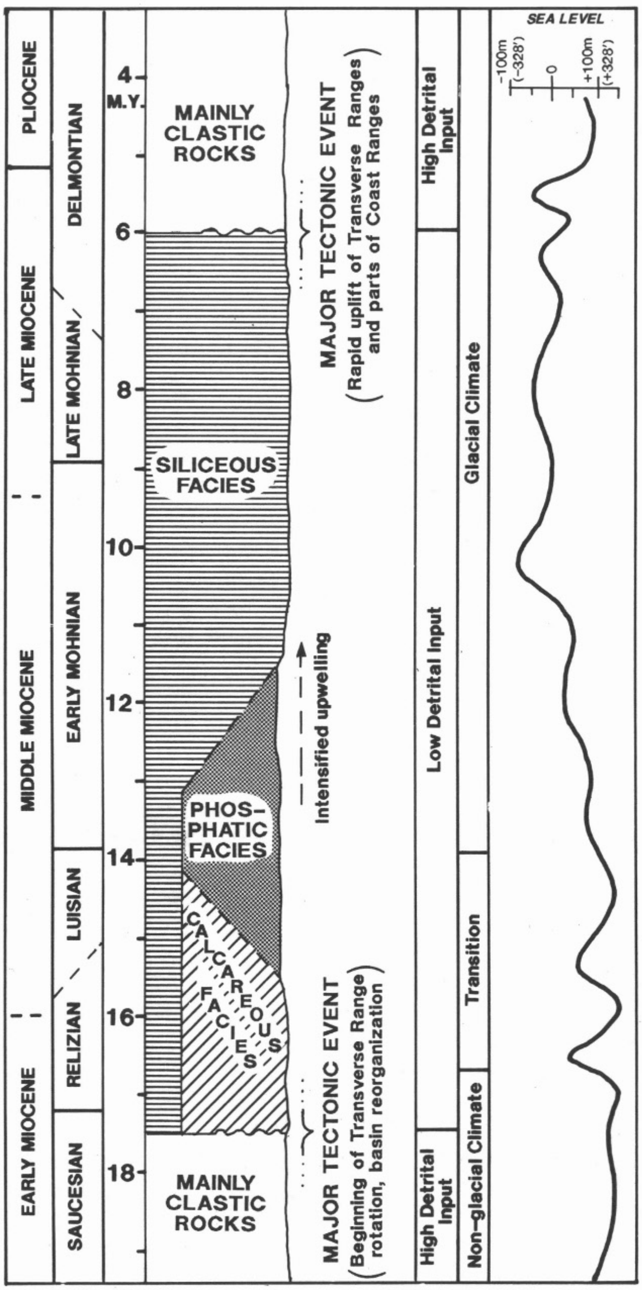

Figure 3. Generalized stratigraphic column of the Monterey Formation where uninterrupted by coarse clastics. Climatic interpretation from Woodruff and others (1981); sea level curve from Haq and others (1988); tectonic events referenced in text; ages from Isaacs (1983), Barron (1986), and Dumont (pers. comm.).

STRATIGRAPHY

The Monterey Formation was originally laid down as diatomaceous and calcareous oozes with variable detrital content. Subsequently, most of these sediments were buried deep enough so that the diatomaceous oozes were converted, through diagenesis, into the familiar siliceous rock types -

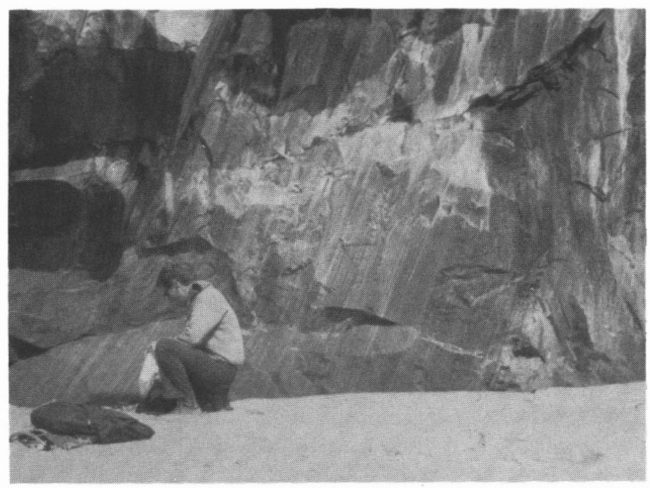

Figure 4. Diatomaceous sequence, coastal California.

Figure 5. Thin-bedded sequence of chert, porcelanite, mudstone, and dolostone; siliceous facies, coastal California.

TABLE 1 Common Monterey Formation Rock Types

CHERT: An aphanitic siliceous rock (either opal CT or quartz) with a glassy, vitreous, or waxy luster; hardness >5; a knife will not scratch it; fractures concoidally or into angular fragments; color varies in shades of white, black, brown, to nearly colorless.

PORCELANITE: An aphanitic siliceous rock, (either quartz or opal CT) with a matte luster resembling unglazed porcelain; commonly laminated; hardness <5; a knife will scratch it; fractures concoidally or into splintery fragments; light colored when weathered-brown to gray when fresh.

MUDSTONE OR SHALE (SILICEOUS): A fine-grained, massive to crudely laminated rock which dents on impact, fractures irregularly or along bedding, and shows a waxy impression when scratched.

DOLOSTONE OR MARL: Dolostone: an aphanitic to sucrosic rock which, when powdered, fizzes vigorously in dilute HCl; usually gray to brown when fresh-weathers to a yellow color; scratches easily but very tough, rings when hit with a hammer. Marl: A fine-grained, light to dark-colored rock that fizzes vigorously in HCl and is similar in appearance to mudstone or shale as described above.

DIATOMACEOUS ROCK: A massive to laminated soft rock that is noticeably lightweight and very porous; fresh samples are brownish-weathered samples are chalk white; detritus-rich varieties show a waxy impression when scratched; diatom-rich varieties powder when scratched.

chert, porcelanite, and siliceous mudstone (Figures 4 and 5), and calcareous oozes became calcareous or dolomitic mudstone or dolostone (Table 1). At present, diatomaceous rocks remain only in areas where burial was minimal, such as along basin margins or at the stratigraphic top of the Monterey Formation.

The thickness of the Monterey Formation is highly variable. The thickest deposits are found in basin centers, such as in the southwest part of the San Joaquin basin where the Monterey is over 3000 m (9800 ft) thick. More typically the formation ranges from 300 to 1000 m (980-3300 ft) thick (Figure 6) with thinner deposits preferentially located on what were persistent bathymetric highs and basin margins.

The distinctive, thin-bedded biogenic character of the Monterey Formation is best developed in areas that were most protected from terrigenous input from land. In general, the purest biogenic sections are found in the Coast Ranges and the western Transverse Ranges, areas that were probably located furthest from land in Miocene times. Clastic-dominated sections coeval with biogenic Monterey deposits, are typically found in easternward outcrops or in areas that had access to turbidites.

An example of distal to proximal facies relationships in southern California is shown in Figure 6. In this area, the Monterey changes from nearly pure hemipelagic deposits near Point Conception - to subequal clastics in the coeval Modelo Formation in the Santa Monica Mountains - to dominantly clastics in the Puente Formation in the Puente Hills.

Examples of turbidite sandstone bodies within the Monterey Formation include the highly oil productive Stevens sandstone in the San Joaquin Valley (Graham and Williams, 1985) and lower Monterey sandstone in the western Santa Barbara Channel offshore (Ogle and others, 1987).

The base of the Monterey, where not obscured by coarse clastics, is generally sharp and dates at approximately 17.5 Ma (Barron, 1986). The upper contact of the Monterey Formation, where not erosional, is sharp in some areas and apparently gradational in others. Where the contact is gradational, overlying strata are siliceous but are generally more detritus-rich than underlying Monterey rocks. Age of the youngest Monterey Formation rocks is approximately 6 Ma (Dumont and others, 1986; Barron, 1986).

In many areas of California the Monterey Formation can be subdivided into three facies: a lower calcareous facies; a middle phosphatic facies; and an upper siliceous facies. Additional subdivisions are recognized locally (e.g., Figure 6).

The CALCAREOUS FACIES consists mainly of massive to laminated dolostone, and calcareous or dolomitic mudstone, siliceous shale, porcelanite, and rare chert. Sections are generally thin-bedded but lithologic variation is wide with some sections composed mainly of carbonate-rich shale, others dominated by dolostone, and still others containing abundant carbonate-rich siliceous mudstone and porcelanite.

The PHOSPHATIC FACIES is characterized by dark, laminated, organic-rich, phosphatic, calcareous or dolomitic mudstone or marl. Dolostone nodules or beds are typically present. The phosphate most commonly occurs in light-colored laminations or in blebs up to a few centimeters long.

The SILICEOUS FACIES is characterized by the predominance of laminated siliceous rock types - mainly porcelanite, siliceous mudstone, and chert. In addition, dolostone, and carbonate-rich mudstone are common in many areas. Two distinct end-member outcrop types are recognizable. On one extreme are outcrops that are distinctly heterogeneous with widely varying rock types, abundant chert and dolostone, and highly variable bed thickness. On the other extreme are outcrops composed almost entirely of uniformly thin-bedded siliceous mudstone and porcelanite.

The three facies are not present everywhere and are partly age controlled (Bramlette, 1946; Pisciotto and Garrison, 1981). The siliceous facies is the most distinctive and widespread. Its age spans the age range of the Monterey formation (17.5 to 6 Ma), but it is best developed in most areas starting around 12-13 Ma ago. The calcareous and phosphatic facies are best developed in the Central Coast Ranges and Western Transverse Ranges; they are

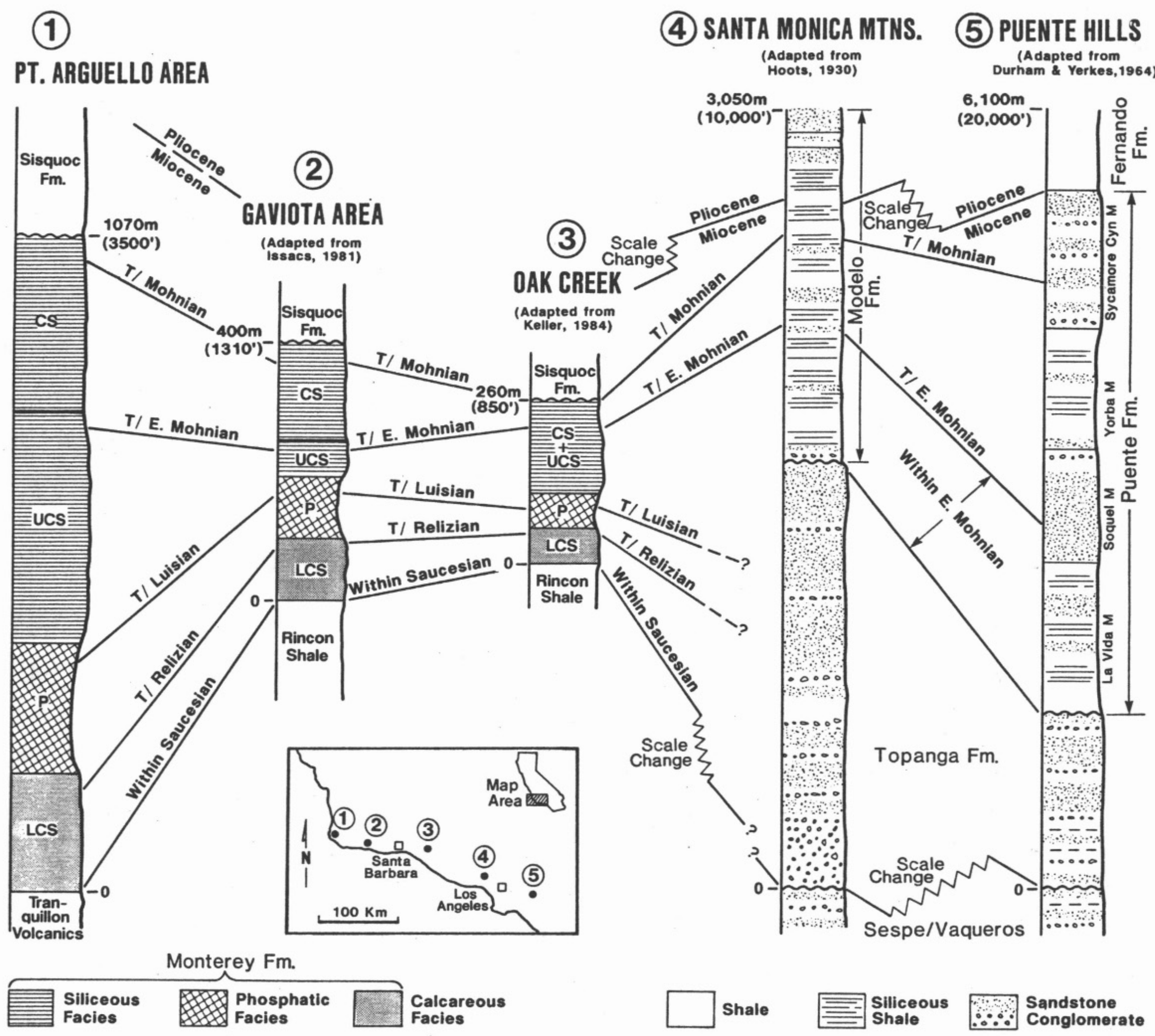

Figure 6. Comparison of five Miocene sections in Southern California. The Monterey Formation in the western three sections contain easily recognizable calcareous, phosphatic and siliceous facies (shaded). Local terminology shown by letters: CS = clayey-siliceous member; UCS = upper calcareous-siliceous member; P = phosphatic member; LCS = lower calcareous-siliceous member. Note that the two eastern sections (paleo-inboard) are dominantly sandy but still contain siliceous rocks, especially of Mohnian age. Age control is from references cited in Figure and M. P. Dumont (pers. comm.).

poorly represented or absent in many other areas, most notably in parts of the San Joaquin Valley. Where present, the calcareous facies typically extends from 17.5 Ma (base of Monterey) to 13 to 14 Ma; the phosphatic member most commonly ranges from 12-15 Ma (Figure 3).

SILICEOUS ROCK TYPES AND DIAGENESIS

The shells of diatoms, radiolaria and other siliceous organisms are composed of opal A, a hydrous silica phase ($SiO_2.nH_2O$) that is inherently unstable. With burial, and mainly due to increasing temperature, opal A dehydrates and converts to a silica phase similar to crystobalite-tridymite called opal CT (SiO_2). With a further increase in burial, opal CT converts to ordinary quartz.

The transformation from opal A to opal CT to quartz results in major changes in rock properties and appearance (Murata and Larson, 1975; Isaacs, 1981a, 1981b). The most dramatic change occurs in the transformation of soft, punky diatomaceous rocks (opal A) to chert, porcelanite or siliceous mudstone (opal CT) which are typically hard and brittle. The change is accompanied by a sharp drop in porosity (Isaacs, 1981b) and an increase in grain and bulk density.

A further decrease in porosity and density values occurs during the transformation of opal CT to quartz. However, in this transformation, the hand specimen appearance of rocks does not usually change appreciably. For example, an opal CT chert normally cannot be distinguished from a quartz chert in hand specimen. Generally, the only sure way to distinguish opal CT and quartz phase rocks is by x-ray diffraction, though in some cases, particularly with cherts, thin section examination may suffice.

Temperature appears to be the major factor controlling silica phase transformations. Various studies (summarized in Isaacs, 1987) have shown that the transformation temperatures for opal A to CT are approximately 45° to 50° C (113° to 122° F) and for opal CT to quartz are approximately 75° to 85° C (167° to 185° F). In addition, composition has a minor but important control on transformation temperature - rocks with high silica/detritus ratios convert from opal A to opal CT at lower temperatures and from CT to quartz at higher temperatures than rocks with low silica/detritus ratios (Isaacs, 1982). Outcrop and well studies show that the stratigraphic thickness of both the opal A to opal CT and opal CT to quartz transition intervals varies from 20 to 300 m (66 to 980 ft) (Keller and Isaacs 1985).

Because temperature is the main control, burial depth required for silica transformations depends on local heat flow. The geothermal gradient typically varies from approximately 25° to 50° C/km (1.4° to 2.7° F/100 ft) in Monterey Formation basins in California. This translates into a burial depth range of 600 to 1400 m (1970 to 4590 ft) for the opal A to opal CT transition and 1200 to 2800 m (3940 to 9190 ft) for the opal CT to quartz transition, for onshore sections (using surface temperature = 15°C, (59° F) and assuming a constant past geothermal gradient; Figure 7). These calculations are consistent with well data.

ROCK CLASSIFICATION

Hand specimen characteristics of diagenetically altered Monterey rocks (Table 1) are in large part controlled by the relative proportion of three main

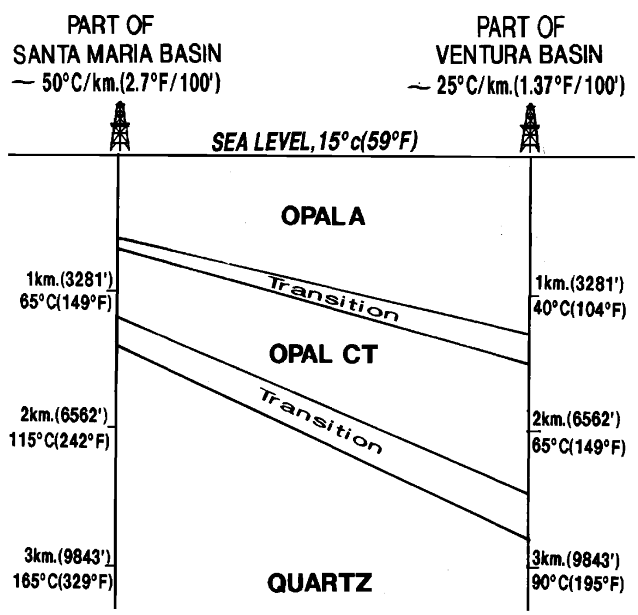

Figure 7. Diatomaceous rocks (opal A) are transformed to opal CT at approximately 45° to 50° C (113° to 122° F) and to quartz at approximately 75° to 85° C (167° to 185° F) (Isaacs, 1987). Paleo-geothermal gradients will control the depth of these transformations. In this diagram, the entire section is imagined to be diatomaceous and typical present-day geothermal gradients are used.

components: (1) biogenic silica; (2) detritus; and (3) carbonate (Isaacs, 1981a). For example, geochemical analysis consistently show that cherts contain greater than four parts biogenic silica to one part detritus. As another example, analyses of typical porcelanites give silica-detritus ratios of from 1:1 to 1:4. Similarly, siliceous mudstones and dolostones contain high percentages of detritus and carbonate respectively. Such data provide a means of classifying siliceous rocks quantitatively.

The proportion of biogenic silica, detritus, and carbonate is determined using a geochemical method such as described by Carpenter (this volume). In this procedure, the elemental composition of siliceous samples is determined by x-ray fluorescence and other methods. The elemental data are then converted to mineral composition using formulas based on detailed geochemical analyses of control samples from nearby wells or outcrops. Finally, the proportions of biogenic silica, detritus and carbonate are calculated from the mineralogy.

The proportions of biogenic silica, detritus, and carbonate can conveniently be plotted on a triangular diagram to serve as a quantitative basis for classification. Our work shows that cherts, porcelanites, shale/mudstone, and marlstone/dolostone, as

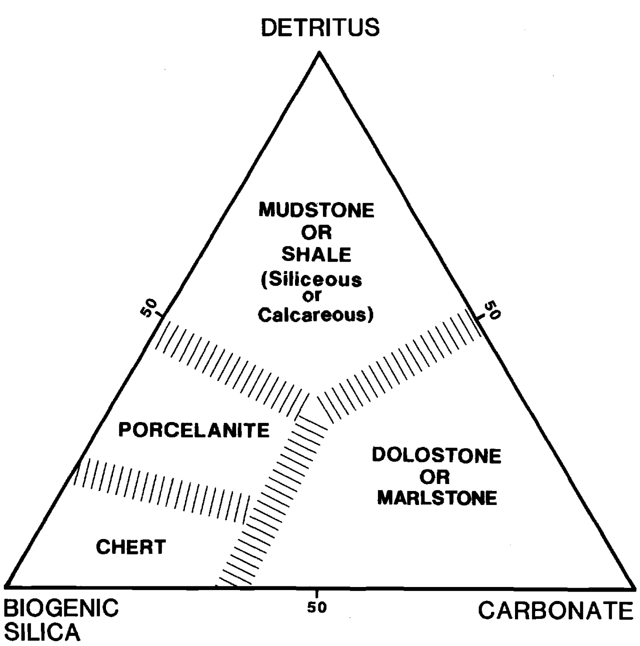

Figure 8. Generalized geochemical classification of Monterey rock types. Rock identification from field characteristics is somewhat operator dependent and therefore the boundaries between rock types is approximate.

identified by field characteristics, consistently fall in the broadly defined compositional fields shown in Figure 8.

Isaacs (1982) pointed out potential problems with a compositional-based scheme for rock classification. She noted that the textures of chert, porcelanite and siliceous mudstone result in part from the crystallization character of silica. She noted that clay type, rather than simply the amount, might affect crystallization character. She further noted that coarse silt and sand-sized detrital grains are less likely to influence silica crystalline character than an equal volume of dispersed clay-sized particles. We have not noted any correlation of clay type to texture, however, it is obvious that the presence of coarser detrital material in an otherwise fine-grained rock could give erroneous results. Therefore, bimodal rocks such as sandy diatomites and their diagenetic equivalents may require special consideration. Regardless of these problems, our compositional-based scheme (Figure 8) provides a useful basis for quantifying and comparing field-based definitions.

OCEANOGRAPHIC, CLIMATIC, AND TECTONIC CONTROLS ON SEDIMENTATION

The complex interbedding of different rock types that characterizes the Monterey Formation is the result of variations in the amount of detritus, biogenic silica and biogenic carbonate deposited at a given time. The relative abundance of these three components depends on a complex interplay of climatic, oceanographic and tectonic conditions (Garrison, 1981; Kennett, 1982).

Detrital input in the marine environment is largely from fluvial sources (Gorsline and Douglas, 1987) and is strongly influenced by sea level changes, tectonics, and climate. For example, sediment supply will increase if sea level falls, uplift of mountains is initiated or accelerated, and generally in periods of higher rainfall or floods. These factors must have played an important role in controlling sediment input into Monterey Formation basins.

Biogenic silica and carbonate consist mainly of microscopic, shell-forming plants and animals. The plants live in the photic zone, or upper 50 to 100 m (160 to 330 ft) of the water column, and consist mainly of diatoms (siliceous) and coccolithophores (calcareous). The animals consist mainly of foraminifera (calcareous) and radiolaria (siliceous); these live anywhere in the water column or on the ocean floor, but many choose to live in the photic zone because of high nutrient levels there.

Total productivity of the biogenic population is controlled by the supply of nutrients available. Areas of high productivity are generally associated with upwelling zones which form when winds or ocean currents displace surface waters, allowing underlying waters to upwell (Figure 9). Upwelling waters bring nutrients, mainly phosphorus and nitrogen derived from the decay of biogenic organisms, from depths of a few hundred meters to the surface waters where they can be reutilized. Under such conditions, total productivity can be an order of magnitude greater than in areas where upwelling is not present.

An important feature of some upwelling areas is an underlying layer of oxygen-depleted water known as the oxygen minimum zone. These zones are created by oxygen depletion due to the decay and oxidation of dead organisms as they fall through the water column. Above this zone, surface mixing provides oxygen; below this zone, there are fewer organisms present to use up availiable oxygen by respiration or decay. Well developed oxygen minimum zones range in depth from roughly 150 to 1500 m (490 to 4900 ft) (Figure 9). Where an oxygen minimum zone impinges on the ocean floor, little oxidation can take place and organic matter is more likely to be preserved. In addition, large burrowing organisms cannot survive, and thin beds or laminations are preserved rather than destroyed by bioturbation.

Of particular importance to the Monterey Formation is the observation that diatoms out compete all other shelled microorganisms

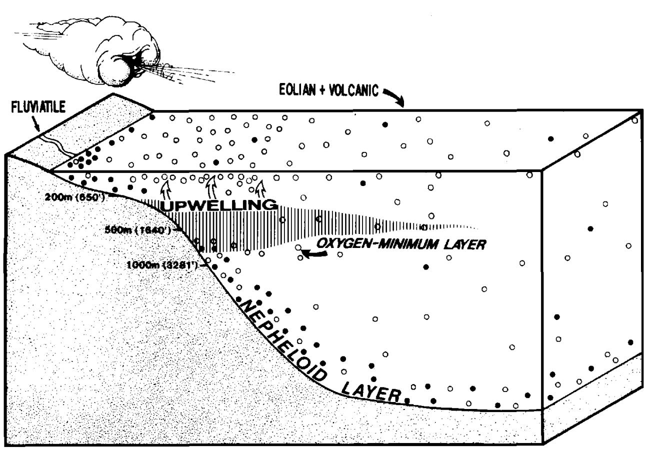

Figure 9. Model for hemipelagic sedimentation in a nearshore upwelling regime; adapted from Rullkötter (1983).

in oceanic areas of high nutrient concentration associated with upwelling areas (Garrison, 1981). In areas of low to moderate productivity, calcareous organisms generally predominate.

Preservation of shell material is also an important variable in determining sediment composition (Garrison, 1981; Kennett, 1982). Calcareous shells are subject to increased dissolution with depth. The depth at which calcareous shells cannot survive (calcium carbonate compensation depth-CCD) averages approximately 4500 m (15,000 ft) in the world's oceans. However, the CCD rises towards the continents and may possibly be very shallow locally, particularly in areas of high productivity where decay of abundant organic matter generates CO2 which increases carbonate dissolution. Siliceous shells composed of unstable opal A are also subject to dissolution. In many areas of low productivity, siliceous shell material is absent in sediments because dissolution occurs before burial can take place. Protection of both siliceous and calcareous material from dissolution is enhanced by incorporation into fecal pellets.

Cycles

Because the input of biogenic and detrital material is dependent on so many variables, frequent variations should be expected. The thin-bedded, laminated and unlaminated rocks of the Monterey Formation attest to this and provide a detailed record of paleoceanographic events. Some of these changes were cyclic in nature as evidenced by the presence of laminations and cyclic bedding in the Monterey Formation (Bramlette, 1946; Pisciotto and Garrison, 1981; Isaacs, 1983).

Laminations in the Monterey Formation are examples of clastic-biogenic cycles of climatic origin. Most of the laminations are apparently varves reflecting seasonal changes in sediment supply (Bramlette, 1946). Dark layers are typically detritus-rich and represent periods of high rainfall and runoff on land (winter), whereas the light-colored layers are concentrations of relatively undiluted biogenic siliceous material (summer). Many modern analogues have been recognized, such as oxygen-starved areas of the Santa Barbara Channel and parts of the Gulf of California (Donegan and Schrader, 1981).

Somewhat longer term climatic fluctuations appear to have created much of the thin-bedded character of the Monterey Formation. For example, alternating detritus-rich versus biogenic-rich beds, typically from a few to 30 cm (1 to 12 in) in thickness are thought to represent time periods of 10's to 100's of years, as determined from counting laminations inferred to be varves (Pisciotto and Garrison 1981). In some cases, the clastic part of these cycles appears to be of turbidite origin. In other cases the clastic portion appears to represent sustained rather than a sudden influx of detrital material. In both cases, climatic fluctuation, such as periodic floods or dry-wet cycles of longer duration may be the main controlling factor.

An additional explanation for the thin bedded character of the Monterey Formation includes short term changes in oceanographic conditions. For example, silica-carbonate variations may be due to preferential preservation or changing temperature-nutrient conditions that favor the growth of certain organisms. As an example, unusually pure diatomaceous beds (chert precursors?) may have been deposited during short periods of intense upwelling, which created diatom blooms while promoting dissolution of carbonate.

Alternating laminated and non-laminated intervals represent still another type of cycle probably controlled by oceanographic conditions. These intervals are generally from a meter (3.3 ft) to tens of meters (33 ft +) in thickness and are most clearly visible in diatomaceous sequences. Most of the non-laminated intervals have clearly been bioturbated; they appear to represent periods of oxygenated conditions during which time large burrowing organisms could survive. Laminated intervals represent low oxygen conditions and the absence of burrowing organisms. Such periodic fluctuations in oxygen are probably due to ocean circulation changes.

ORIGIN OF DOLOMITE

The origin of dolomite in the Monterey Formation has been the subject of extensive research in recent years (e.g., Garrison and others, 1984). Most of the dolomite occurs in three forms: (1) nodules; (2) beds; and (3) as disseminated crystals.

Disseminated dolomite is the most common type of occurrence in the Santa Barbara-Santa Maria area (Isaacs, 1984) and perhaps in the Monterey Formation in general. In thin section, this dolomite is typically evenly dispersed as silt and clay-sized crystals in mudstone, porcelanite and chert.

Dolostone nodules and beds make up only a small portion of the Monterey Formation but they are conspicuous because they are hard and resistant to erosion and often weather to an unusual yellow-orange color. Dolostone nodules are generally less than a meter (3 ft) thick and up to several meters (10 ft +) in length. Dolostone beds are generally less than a meter (3 ft) thick but range up to 3 m (10 ft) in thickness.

Most dolostone beds and nodules apparently formed early, while the host sediment was still soft and uncompacted. Field evidence for this include differential compaction features best displayed by nodules - bedding in the host rock commonly wraps around nodules (Figure 10) and furthermore, laminations in nodules are typically further apart than in laterally equivalent host rock. Laboratory evidence for early formation includes oxygen isotope studies which show that many beds and nodules apparently formed in the 20° to 30° C (68° to 87° F) range (see papers in Garrison and others, 1984). This corresponds to a burial depth of no more than a few hundred meters (several 100 ft).

Modern analogs of early formed dolomite have been recognized by ocean drilling in areas of organic-rich sediments as found, for example, off Peru and in the Gulf of California. In these areas, nodules and beds are forming today at very shallow burial depths and possibly on the sea floor itself. The origin of early dolomite is apparently promoted by the oxidation of organic matter which results in excess CO2.

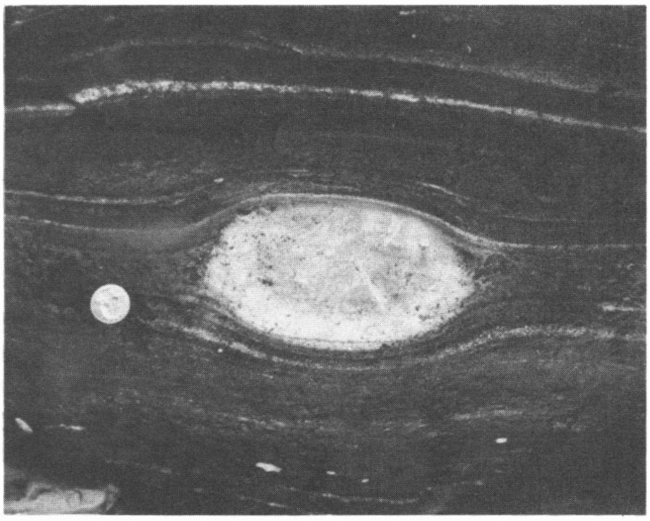

Figure 10. Small dolostone nodule in phosphatic mudstone. Note how laminations bend around nodule indicating that the nodule formed prior to full compaction of the host sediment.

Furthermore, reduction of sulfate to sulfide is considered important because sulfate is a major inhibitor of dolomite formation. Magnesium may be derived directly from sea water (see papers in Garrison and others, 1984).

An important aspect of early formed dolomite is that it helps preserve geologic features that are usually destroyed in other rock types. Dolostone beds and nodules commonly contain; (1) detrital remnant magnetism useful in paleomagnetic tectonic studies; and (2) dateable siliceous microfossils in otherwise barren sequences.

Though early dolomite is common, oxygen isotope data indicate a later higher temperature origin for much of the disseminated dolomite and at least the outer parts of some nodules and beds. Some of this dolomite apparently formed by later replacement of calcite rather than direct precipitation. In addition, dolomite also occurs locally as a late stage diagenetic vein filling (Redwine, 1981; Roehl, 1981).

ORIGIN OF PHOSPHATIC BEDS

Phosphatic beds in the Monterey Formation are of particular interest because they typically have very high total organic carbon (TOC) values and are excellent hydrocarbon source rocks. In addition, phosphatic material is a key environmental indicator.

Two main phosphatic facies have been recognized in the Monterey Formation by Garrison and others (in press). These are: (1) a phosphatic marlstone facies; and (2) a pelletal-oolitic phosphorite facies.

Phosphatic marlstone is by far the most common type. It is typically laminated, commonly contains 50% or more calcite or dolomite and is thought to have originally been deposited as a coccolith-foraminiferal-diatom mud (Figure 11). Fossilized bacterial mats are commonly present and may be a major component of the organic matter (Williams and Reimers, 1983). The phosphate is found in blebs and pellets or as disseminated crystals along laminations. The phosphate appears to have formed prior to significant compaction as indicated by field evidence for soft sediment deformation of phosphate nodules and the inclusion of nodules in early formed dolostone.

Garrison and others (1987) suggest that phosphatic marlstone formed by slow sedimentation in a low oxygen environment (oxygen minimum zone) dominated by calcareous sedimentation. Such conditions may be met in a variety of enviroments such as outer shelf, slope, basin floor, and on

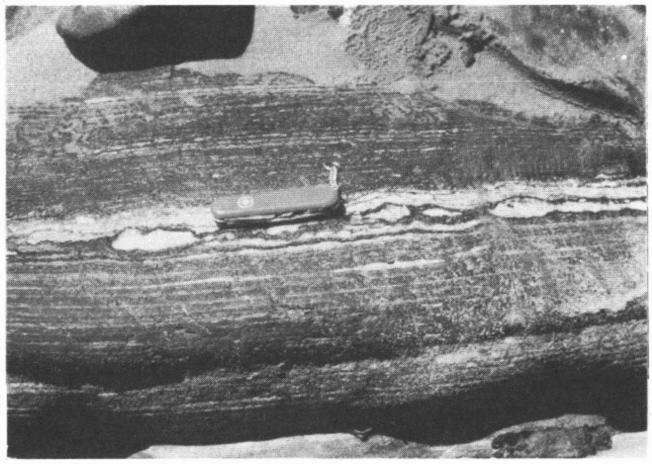

Figure 11. Laminated phosphatic mudstone containing a layer of phosphatic nodules.

deeply submerged banks. Modern analogues cited by Garrison and others (1987) include the upwelling areas off Peru and on the west African shelf. In these two areas, phosphorites are forming today on or within a few centimeters (inches) of the seafloor and within oxygen minimum zones associated with upwelling.

Pelletal-oolitic phosphorites, as described by Garrison and others (in press), consist of phosphatic sandstones some of which contain glauconite, terrigenous material, phosphatic micronodules, intraclasts, and fish fragments. These phosphorites are typically interbedded with siliceous mudrocks, and in some cases probably formed by winnowing of sand-sized phosphatic grains by wave or current action. In other cases the phosphate may have formed on the sea floor without significant reworking. Pellatal-oolitic phosphorites are though to have formed on shelves and bank tops; some of these deposits were apparently redeposited in deeper water by turbidity currents.

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REFERENCES

Combined references are at end of the following paper.