Diagenetic Aspects of Morgan Formation (Pennsylvanian) Shelf Carbonates, Northern Utah and Colorado

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ABSTRACT.—Preliminary study of diagenetic features occurring in carbonate strata of the Middle Pennsylvanian Morgan Formation (northeastern Utah and northwestern Colorado) has revealed a complex history of calcite and silica cementation, dolomite and silica replacement, dedolomitization, fracturing, pressure solution, and late-stage calcite cementation. Fibrous to bladed calcite spar fringe cement grown isopachously upon former aragonite grains is interpreted as former early submarine aragonite or high-Mg calcite, later neomorphosed to low-Mg calcite. Calcite spar grown syntaxially upon echinodermal substrates and equant calcite spar filling interparticle and intraparticle pores both postdate fringe cementation. This later calcite spar precipitation probably occurred in a meteoric phreatic environment. Selective dolomite replacement of micritic grains and matrix probably occurred after most calcite cementation. Dolomite rhombs from the eastern study area are very finely crystalline, limpid, and unzoned. Dolomite rhombs from the western study area, in contrast, are coarsely crystalline and cloudy and display numerous syntaxial ferroan dolomite, dolomite, ferroan calcite, and calcite overgrowths. The precipitational environment for Morgan dolomitization is not entirely clear, but the present data seems in favor of a phreatic, meteoric-marine mixing-zone interpretation. Cathodoluminescence reveals a complex trace element zonation in both the calcite spar cements and the coarsely crystalline dolomite rhombs. Dedolomitization is a relatively rare phenomenon. It either formed as a diagenetic event almost coeval with dolomitization, or during late-stage, near-surface weathering. Silica cementation and replacement probably occurred concurrent with or after dolomitization. Silica cementation of primary depositional porosity is rare; it is most common in secondary solution porosity and consists of two generations of length-fast chalcedony and one later generation of megaquartz. Silica replacement in the form of chert nodules and beds is selective for spiculitic, mud-supported carbonates and for former high-Mg calcite skeletal grains. The environment for Morgan silicification is not entirely clear, but the present data seem in favor of a phreatic, meteoric or meteoric-marine mixing-zone interpretation. Probable pseudomorphs of halite, nodular anhydrite, and gypsum rosettes were first replaced (or possibly filled after early dissolution) by calcite spar, followed by ferroan dolomite, length-slow chalcedony, and megaquartz replacement. Pressure solution postdates all previously mentioned features and is manifested by sutured stylolitic surfaces, interpenetrating grain-to-grain contacts, and nonsutured solution seams. Late-stage cementation by brightly luminescent, nonferroan calcite spar and by later dull-luminescent, ferroan calcite spar fills earlier tectonic(?) fractures, intercrystalline and biomoldic porosity, and some stylolitic surfaces. Porosity preservation is greatest in coarsely dolomitized facies, in which up to 10–15% combined intercrystalline and biomoldic porosity remains; the lowest porosities (1%) occur in calcite spar-cemented grainstones.

INTRODUCTION

The Upper Morgan Formation is a 200-m-thick Middle Pennsylvanian (Atokan–Desmoinesian) sequence consisting largely of 5- to 25-m-thick, fine- to very fine-grained quartzose sandstone intervals (70–80% of total section) interbedded with 0.25- to 11-m-thick, mostly fossiliferous limestone and dolomite units (fig. 1). It is well exposed in northeastern Utah and northwestern Colorado within the Wasatch and Uinta Mountains, and also across parts of the Colorado Plateau (fig. 2). The Morgan Formation was first investigated sedimentologically by Driese and Dott (1984) in order to determine whether or not it represented a form of cyclic sedimentation unique to the western edge of the North American craton. Physical and biogenic sedimentary structures, as well as carbonate lithofacies were both described and interpreted. The wide spectrum of carbonate lithofacies was interpreted as dominantly shallow subtidal in origin. Carbonate lithofacies include (a) oolitic, bioclastic, and peloidal grainstones and packstones, most of which contain normal marine faunal assemblages, but some of which show evidence of faunal restriction; (b) whole-fossil wackestones; (c) intraclastic rudstones and floatstones; (d) chaetetid boundstone masses; (e) laminated, abiotic mudstones containing probable evaporite pseudomorphs; (f) low-angle, cross-bedded, fossiliferous quartz sandstones to well-abraded, sandy grainstones. Sandstone intervals were deposited by eolian dune progradation. Carbonate intervals were deposited after rapid marine transgressive events; carbonate environments were initially stenohaline, but gradually became increasingly restricted to episodically hypersaline as carbonate sediments accumulated up to shoal conditions. With the use of a modified Markov chain analysis procedure described by Carr (1982), repetitive vertical sequences were quantitatively verified. Eustatic changes—possibly related to Gondwana glaciation—were hypothesized as the probable causal mechanism for the repetitive depositional patterns.

Under the paleoenvironmental model presented by Driese and Dott (1984), pore-fluid compositions would have had the potential to fluctuate from meteoric to marine and, at times, to hypersaline, with each marine advance and retreat. Thus, diagenesis in the Morgan Formation might have been directly influenced by sea-level fluctuations and the accompanying lateral shifts in facies mosaics. Recent work by Heckel (1982) on the diagenesis of Pennsylvanian midcontinent cyclothems suggests that many diagenetic features may be controlled by depositional facies. The Morgan Formation has never been the object of any diagenetic study, aside from preliminary subsurface investigations by Picard (1977), who reported extensive diagenetic modification by dolomitization and silicification. The purpose of the present study is fourfold: (1) to describe, in detail, the diagenetic features present in samples from the Morgan outcrop belt, as revealed by petrographic study; (2) to examine the vertical distribution of diagenetic features relative to the repetitive vertical sequences identified by Driese and Dott (1984); (3) to attempt to put these features into a relative time sequence; (4) to examine lateral (regional) variations in diagenesis. Such diagenetic study could aid in further evaluation of the upper Morgan Formation and other lithostratigraphically equivalent rocks as potential hydrocarbon reservoirs. Most of the present Morgan gas production occurs in coarsely dolomitized carbonate intervals and not in the interbedded (tight) quartz sandstones (Verville and others 1973; Picard 1977).

METHODS OF STUDY

During the summer months of 1980 and 1981, the examination of outcrop exposures of the Upper Morgan Formation occupied sixteen weeks in the field. Twenty-two stratigraphic sections were measured (fig. 2). Within the Yampa and Green River Canyons (Dinosaur National Monument), sections were measured at 3- to 10-km intervals, thus permitting detailed study of lateral and vertical facies relationships. Carbonate samples were collected at 0.25- to 1-m intervals within each unit, or wherever a major lithologic change occurred. Approximately 500 samples were slabbed, polished, and examined under low magnification with a binocular microscope. About 200 samples were selected for thin-section examination under a petrographic microscope. Thin sections were stained with alizarin red-S to determine the distribution of calcite and dolomite, and with

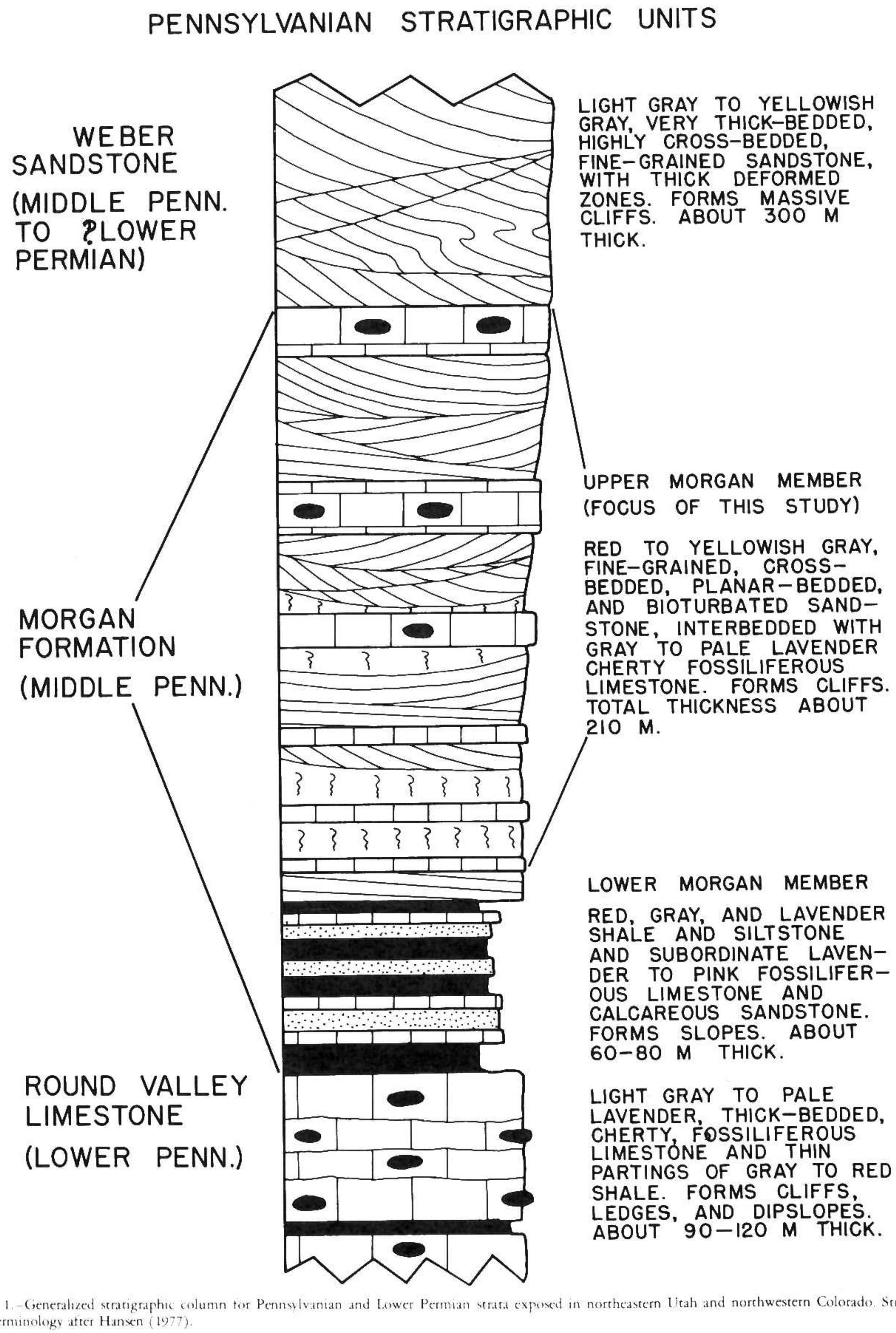

FIGURE 1.–Generalized stratigraphic column for Pennsylvanian and Lower Permian strata exposed in northeastern Utah and northwestern Colorado. Stratigraphic terminology after Hansen (1977).

potassium ferricyanide to determine if the calcite or dolomite was ferroan (Dickson 1966).

The luminescent characteristics of the various carbonate components were revealed by examining uncovered, lightly polished thin sections with a Nuclide Corporation ELM-2A Luminoscope® mounted on a conventional petrographic microscope. Luminoscope operating conditions were: 10 kV beam energy, 0.6–0.8 ma beam current, cold cathode gun type, and variable beam diameter. Cathodoluminescence (CL) petrography was utilized in this study to supplement standard petrographic investigations. Sippel and Glover (1965) and Kopp (1981) have alluded to the usefulness of CL in detecting trace element zonation and "hidden" fabrics in carbonate rocks. Meyers (1974) utilized CL in deducing a cement stratigraphy of Mississippian carbonates in the Sacramento Mountains of New Mexico. Until recently, only qualitative generalizations could be made concerning the relationship between luminescent behavior and trace-element distributions. Frank and others (1982) have demonstrated that the degree of luminescent intensity of a particular calcite zone is controlled by the Fe/Mn ratio, and not by the concentrations of either cation. Frank (1981) similarly proposed that the Fe/Mn ratio also governs luminescence in dolomite. Pierson (1981), in contrast, suggested that the absolute concentration of ferrous iron is the limiting factor for dolomite luminescence.

It is hoped that the present study will serve as a springboard for future diagenetic studies in Pennsylvanian sequences of the western United States. The application of more sophisticated analytical techniques (e.g., electron microprobe, SEM, atomic absorption spectrophotometry, carbon and oxygen isotope ratios, and fluid inclusions) to Morgan Formation diagenetic studies may eventually allow for more precise geochemical interpretations than were possible within the scope of this study.

DESCRIPTION OF DIAGENETIC FEATURES

Micritization of Carbonate Grains

Micritized carbonate grains are observed primarily in bioclastic and peloidal grainstones and packstones, which occur at the base and top of repetitive Morgan sequences (fig. 3). The micritized regions range from a few microns to a half millimeter or so in thickness. A few distinct, micron-scale, tube-shaped microborings filled with micrite or microspar were observed. Typically, the micrite envelopes are internally homogeneous, and no relict internal structures of the host grains are visible. Partially micritized skeletal grains are most common. At least some of the spheroidal grains identified as peloids may represent completely micritized skeletal grains and not fecal pellets.

Discussion

Micrite envelopes in Morgan carbonates probably formed by algal–and/or fungal-boring of carbonate grains in a submarine environment (Bathurst 1966, 1975; Swinchatt 1969).

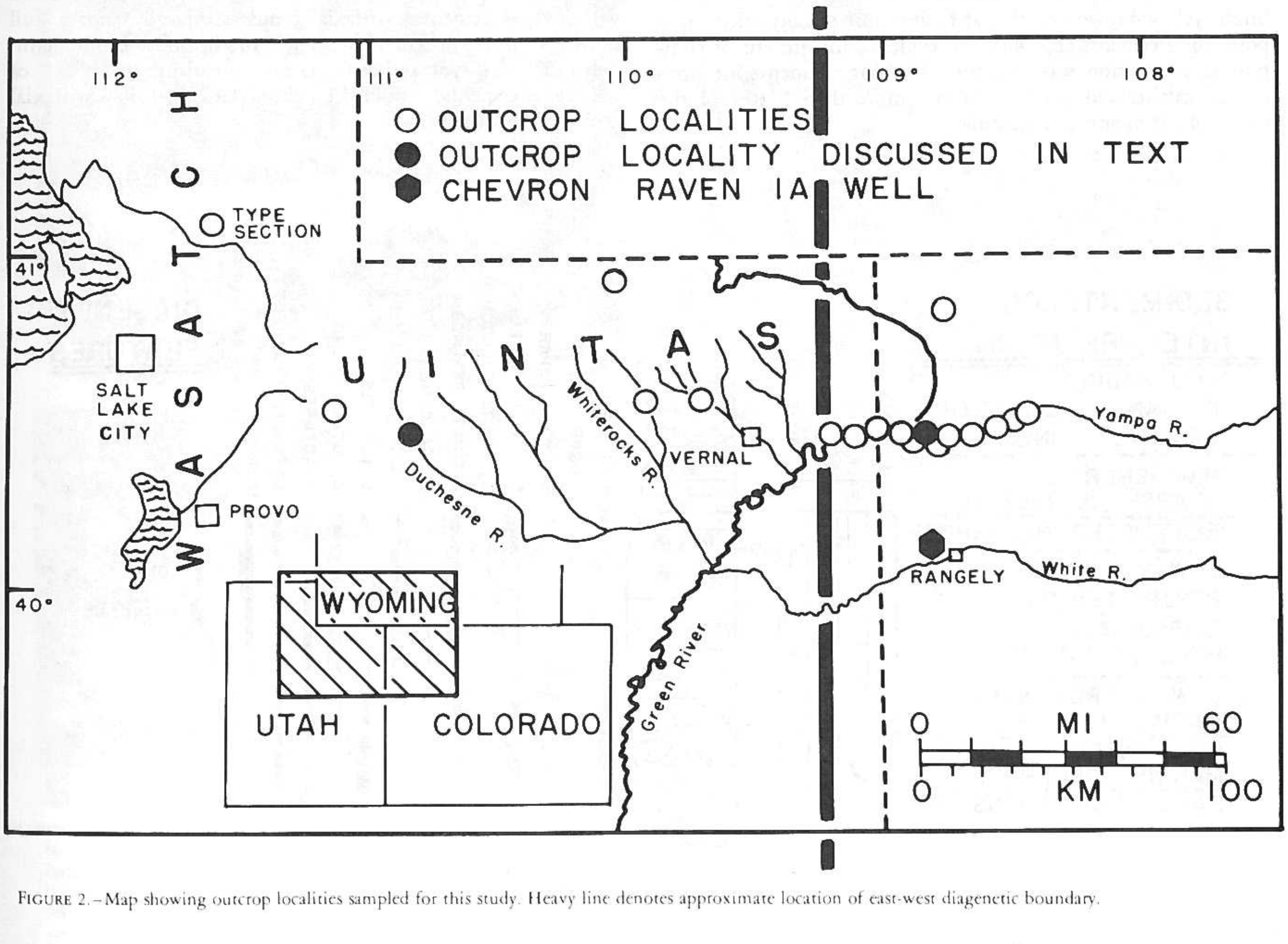

FIGURE 2.–Map showing outcrop localities sampled for this study. Heavy line denotes approximate location of east-west diagenetic boundary.

Because there are no Morgan carbonate facies thought to represent deep-water (below photic-zone) deposition, a photosynthetic algal origin seems most likely, although Friedman and others (1971) cautioned against the overgeneralization that all borings and micrite envelopes were produced by algae.

Micritization could also have occurred beneath the sediment-water interface (May and Perkins 1979). Rare phylloid algal plates in some Morgan facies indicate at least some photic zone deposition. The absence of stromatolites, oncolites, and algal mat structures might reflect an abundance of grazing gastropods. The abundance of micritized grains in basal and upper grain-supported facies of Morgan repetitive sequences (fig. 3) and their near-absence in the central mud-supported facies are probably related to differences in water depth and/or sunlight penetrability.

Fibrous to Bladed Fringe Cements

Fibrous to bladed fringe cements are observed but are rare in grainstone and mud-lean packstone lithologies, which typically occur at the base and top of repetitive Morgan sequences (fig. 3). Crystals are 10–60 μm in length, 2–5 μm in width, and exhibit acute-angle crystal terminations (figs. 4a,b, 5a). Their cloudy nature is probably attributable to an abundance of minute (millimicron-scale) inclusions; whether the inclusions are organic or fluid is uncertain. The best examples of fringe cements occur on substrates of formerly aragonitic composition (e.g., oolitic, peloidal, and molluskan grains). Radiaxial calcite fabrics are generally absent. Fringe cements emit a dark brown, very dull luminescence, which contrasts markedly with the bright yellow-brown colors that typify most equant calcite spar pore-filling cements (fig. 4b). No evidence for growth or compositional zonation was observed. All fringe cements are nonferroan calcite, and never constitute more than 5–10% of the total cement in any given sample.

Discussion

The fibrous fringe cements reported here closely resemble some modern early marine sea floor cements (e.g., Shinn 1969, 1971, Land and Goreau 1970, James and others 1976, Ginsburg and James 1976, Marshall and Davies 1981) or intertidal beach rock cements (Moore 1973, Davies and Kinsey 1973). Their cloudy, inclusion-rich nature and dominant elongation normal to growth substrates suggest precipitation in a magnesium-rich or saline environment in which sideward growth was inhibited by the poisoning effect of magnesium and other foreign ions (Folk 1974) or by differential charge development on crystal faces (Lahann 1978). The isopachous growth pattern and lack of meniscus (Dunham 1971) or gravitational (Muller 1971) growth features suggest precipitation in a phreatic, rather than a vadose environment. The preference of fringe cements for former aragonitic host substrates implies that there was some host control on early cement nucleation, and it appears to be the primary control on their vertical distribution in the Morgan Formation.

Syntaxial Calcite Spar Cements

Syntaxial calcite spar cements are most commonly observed in echinoderm-rich, bioclastic grainstones (fig. 4c), which typically occur at the base of repetitive Morgan sequences (fig. 3). Most syntaxial cements are clear to faintly turbid and are nonferroan calcite. Cement growth is preferentially elongate in directions parallel to the c-axes of monocrystalline host substrates (fig. 4d). No microdolomite inclusions were ever observed within these cements. Syntaxial cements typically emit a dull brown luminescence or none at all. In the most echinoderm-rich lithofacies, syntaxial cements may constitute up to 95% of the calcite cements; in mixed bioclastic facies, 10–40% syntaxial cement is more typical.

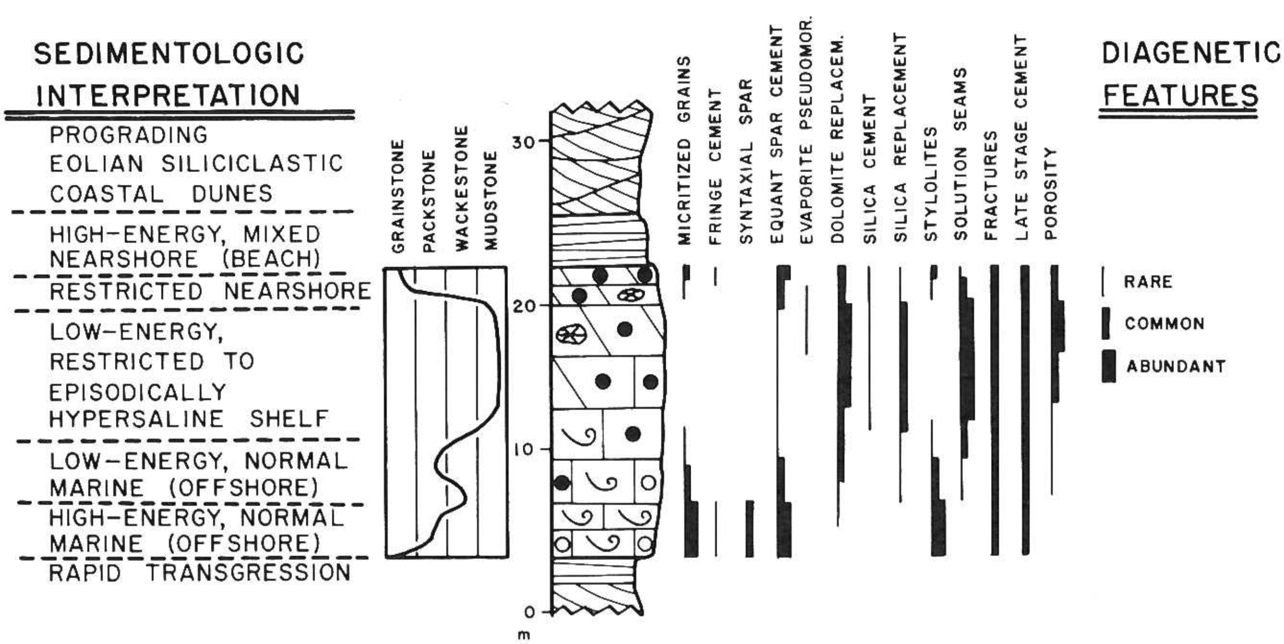

FIGURE 3.—Generalized vertical distribution of diagenetic features as present within repetitive Morgan sequence previously defined by Driese and Dott (1984).

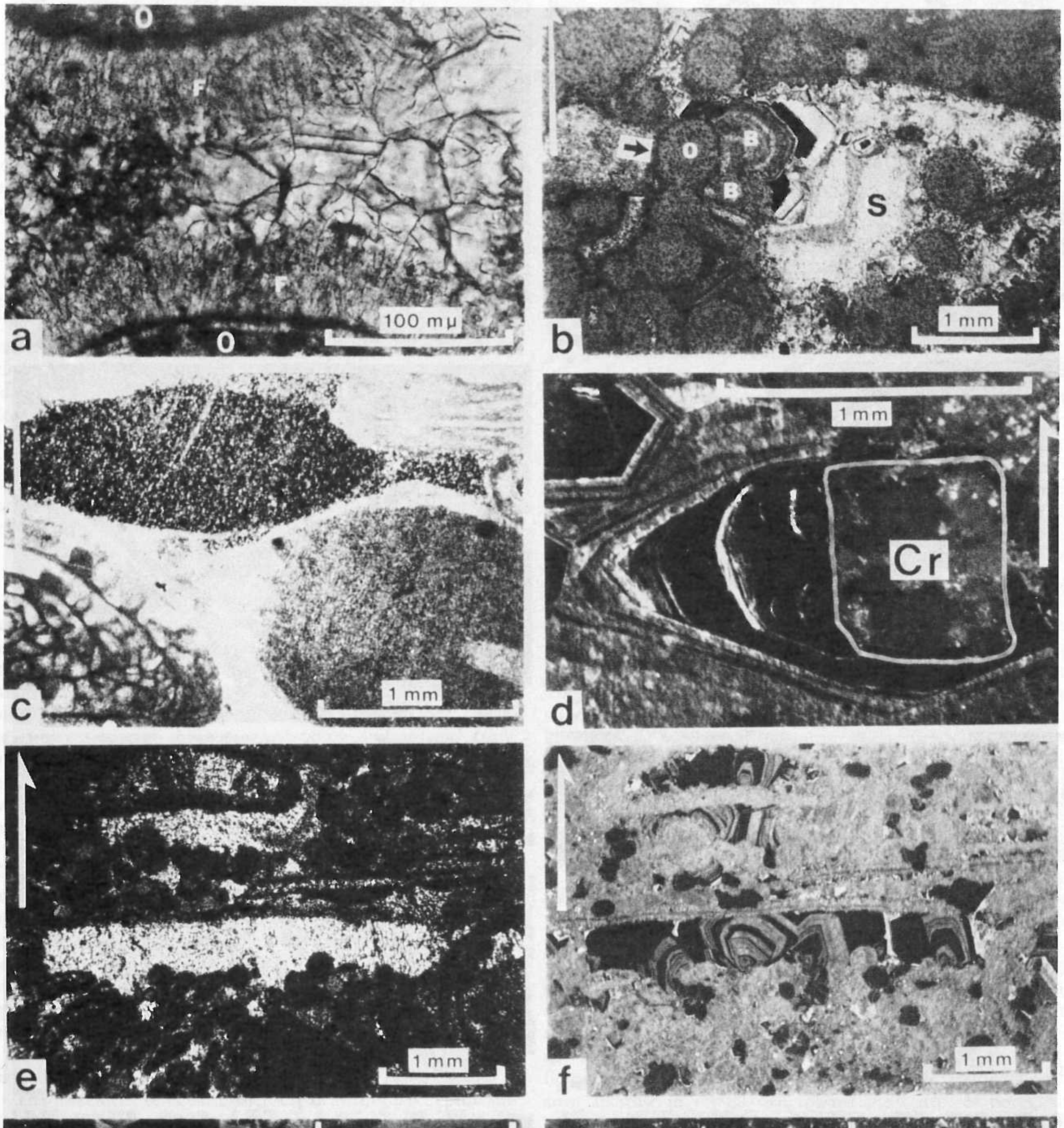

FIGURE 4.–Examples of calcite and quartz cementation: (a) Fibrous to bladed calcite fringe cements (F) rich in inclusions, isopachously coat ooids (O), followed by later equant calcite spar pore-filling cements (S). Thin section, plane light. UW 1740/23. (b) Same oolitic grainstone seen in 4a, as viewed under cathodoluminescence (CL). Ooids (O) luminesce dull yellow-brown, fringe cements (black arrow) are dark brown, botryoidal calcite cements (B) are dull brown, calcite spar cements (S) are bright yellow-brown to nonluminescent. UW 1740/23. (c) Syntaxial calcite spar-cemented crinoidal grainstone; overly close packing due to later solution compaction. Thin section, crossed nicols. UW 1740/24. (d) Syntaxial calcite spar cements viewed under CL. Early nonluminescent cement zones (black) have grown preferentially along direction of c-axis of crinoid columnal (Cr, outlined in white), followed by later zones of yellow-brown luminescing cements (lighter colors). UW 1740/25. (e) Calcite spar cements precipitated in sheltered pores in peloid-bioclast grainstone. Thin section, plane light. UW 1740/26. (f) Same field of view as in 4e under CL. Spar cement consists of alternating yellow-brown luminescing (light color) and nonluminescing (dark color) zones; latest zone is bright yellow and occludes remaining intercrystalline porosity. UW 1740/26. (g) Two generations of length-fast, fibrous chalcedonic cements (1 = clear, 2 = inclusion-rich) isopachously coat dolomitized crinoid grains (CR), followed by later equant megaquartz cements (3). Thin section, plane light. UW 1740/27. (h) Same field of view as in 4g, under crossed nicols. UW 1740/27.

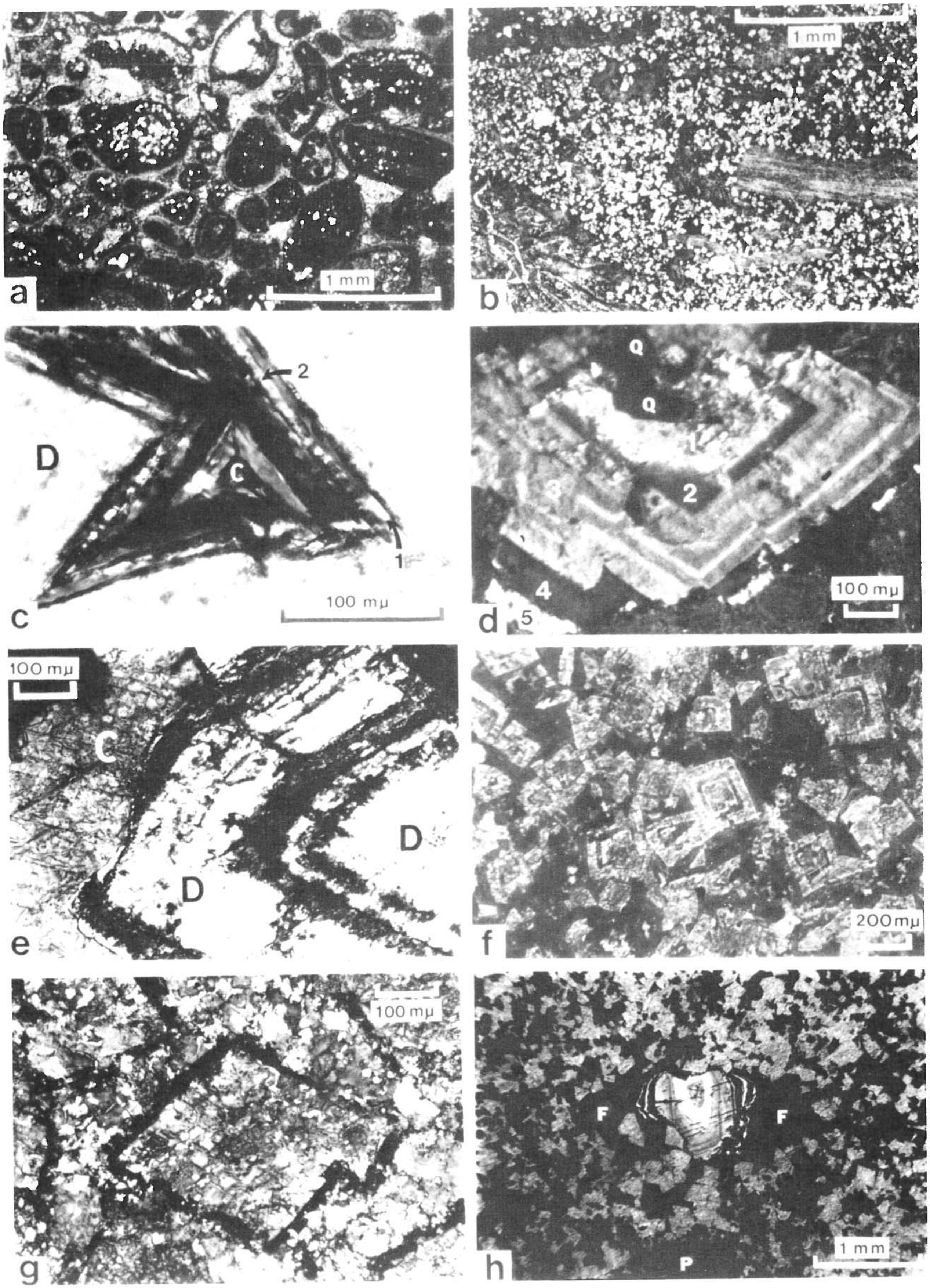

FIGURE 5.–Examples of dolomitization and dedolomitization: (a) Selective partial dolomite replacement of micritic grains in peloidal-intraclast grainstone. Dolomite rhombs are very light colored. Note unaltered fringe and sparry calcite cements. Stained thin section, plane light. UW 1740/28. (b) Selective dolomitization of micritic matrix in bioclastic wackestone. Note unreplaced skeletal grains. Stained thin section, plane light. UW 1740/29. (c) Coarse intergrown dolomite rhombs (D) and syntaxial overgrowths, 1 = calcite, 2 = ferroan dolomite. Intercrystalline space filled with calcite spar or pseudospar (C). Stained thin section, plane light. UW 1740/30. (d) Dolomite rhomb exhibiting complex zonation under CL. Zone 1 luminesces bright yellow orange; Zone 2 is dull red-brown; Zone 3 is medium-bright orange with faint red-brown bands; Zone 4 is dull red-brown; Zone 5 is bright yellow; nonluminescing bodies (Q) are detrital quartz sand. UW 1740/31. (e) Coarse dolomite rhomb (D) containing syntaxial ferroan calcite overgrowths (dark bands). Rhomb partly replaces(?) calcite spar cement (C); rhomb border appears partly corroded. Stained thin section, plane light. UW 1740/32. (f) Zoned dolomite rhombs viewed under CL. UW 1740/33. (g) Rhombic forms, formerly dolomite, now completely replaced by calcite pseudospar. Note opaque iron oxides and organic matter displaced to outer margins of rhombs. Stained thin section, plane light. UW 1740/34. (h) Porous dolomite viewed under CL. Note pore (biomold ?) occluded with bright yellow-luminescing calcite spar cements that display complex zonation; remainder of pore cemented by nonluminescent ferroan calcite spar (F). Other black areas represent either empty pore space (P) or detrital quartz sand. UW 1740/35.

Discussion

Syntaxial spar cement growth was discussed by Evamy and Shearman (1965, 1969) and Bathurst (1975). The clear, inclusion-free nature of most syntaxial cements in the Morgan and the absence of vadose characteristics suggest precipitation in a meteoric phreatic environment (Meyers 1974, 1978; Folk 1974). The absence of microdolomite inclusions probably rules out a former marine, high-Mg calcite precursor for these cements (Lohmann and Meyers 1977, Meyers and Lohmann 1978). Fabric relationships suggest that syntaxial spar precipitated after fringe cement growth but before equant spar cementation. Thus the fringe cements are probably early submarine and may have originally precipitated as aragonite or high-Mg calcite. Syntaxial cements, in contrast, might be early burial phreatic and probably originally precipitated from meteoric waters as low-Mg calcite. The distribution of echinoderm grains within normal marine grainstone facies appears to be the primary control on the vertical distribution of syntaxial spar cements in the Morgan Formation.

Equant Calcite Spar Cements

Clear, 0.05- to 0.5-mm-diameter crystals of equant calcite spar occur as both inter- and intraparticle pore-fillings throughout all parts of repetitive Morgan sequences (fig. 3). The coarsest cement crystals grew within cavities formed by larger skeletal grains (chiefly brachiopods and ostracods) or in sheltered pores (figs. 4b,e). Intercrystalline boundaries are planar, crystal size increases toward the centers of pores, and enfacial triple junctions predominate. Equant spar cements are always nonferroan calcite. Standard petrographic examination did not reveal any evidence of compositional or growth zonation. CL, in contrast, revealed multiple increments of growth; dull brown or nonluminescing zones alternate with yellow-brown to bright yellow luminescent ones (figs. 4b,f). The widths of discrete luminescent zones range from a few microns up to a millimeter. In general, the spar cements do not display a distinct "stratigraphy" of luminescent zones; although correlation of zones within the same sample was possible, both sample-to-sample and outcrop-to-outcrop correlation proved impossible. Equant spar cements typically constitute 60–100% of all calcite cements in any given sample.

Discussion

The placement of equant calcite spar in original depositional porosity, together with the high percentage of plane intercrystalline boundaries and the increase in crystal size toward the center of pores, suggests its precipitation as a void-filling cement in the Morgan Formation (Bathurst 1975). The clear, inclusion-free nature of all Morgan equant spar and the absence of vadose characteristics suggest precipitation in a meteoric phreatic environment (Meyers 1974, 1978; Folk 1974). A former marine, high-Mg calcite precursor for the equant spar seems unlikely on the basis of the absence of microdolomite inclusions (Lohmann and Meyers 1977, Meyers and Lohmann 1978); precipitation as low-Mg calcite is more probable. Fabric relationships suggest that the equant calcite spar precipitated mostly after the syntaxial spar. Whether or not both were derived from the same pore-fluids is not certain, but their different luminescent properties do imply differences in Fe/Mn ratios (Frank and others 1982) and, therefore, differences in pore-fluid chemistries. The complex compositional zonation in Morgan equant spar cements, as revealed by CL, indicates cement growth in many different stages, as opposed to a single generation of growth for most fringe and syntaxial calcite spar cements. The widespread distribution of equant calcite spar cements in all carbonate facies within Morgan repetitive sequences seems to imply that substrate preference was not an important factor governing their distribution, unlike fringe and syntaxial calcite spar cements; instead, the availability of open pore spaces was most important.

Chalcedony and Megaquartz Cements

Most silica present in Morgan Formation carbonates, aside from detrital quartz sand and silt, seems to be of replacive origin. However, some pore-filling silica cements are patchily distributed within porous, dolomitized grainstone and packstone facies from the western part of the study area (fig. 2). They are relatively insignificant compared to calcite cements and only occur at the top of repetitive Morgan sequences (fig. 3). Three distinct generations of silica cements are generally observed in Morgan carbonates (figs. 4g,h): (1) clear, isopachous, length-fast chalcedony containing micron-scale microdolomite inclusions; (2) cloudy, isopachous, length-fast chalcedony containing abundant opaque, minute (millimicron-scale) ferric iron inclusions; (3) clear, 0.05- to 0.25-mm-diameter, equant megaquartz containing large (5–20 μm) fluid and/or vapor-filled inclusions. In both chalcedonic cements, the fiber-bundles are grown elongate normal to the pore walls. The equant megaquartz crystals increase in size toward the centers of the pores. This general tripartite zonation of silica cements was also observed within the interiors of many chert nodules; some nodule interiors still retain unfilled porosity. The first chalcedonic cement exhibits a dull brown luminescence, whereas the second, darker one luminesces pale lavender. The megaquartz cement does not luminesce. Silica cements generally constitute less than 1–2% of all cement mineralogies present in any given sample.

Discussion

The chalcedony and megaquartz are here interpreted as pore-filling cements on the basis of their distinct growth fabrics. The isopachous growth pattern, elongation of fiber-bundles normal to substrate walls, and lack of vadose characteristics all suggest that the length-fast chalcedony grew as a phreatic cement within an open cavity; length-fast chalcedony, according to Folk and Pittman (1971), occurs primarily as a cavity filling. The high percentage of plane intercrystalline boundaries, increase in crystal size toward the center of pores, and lack of vadose characteristics suggest that the megaquartz also precipitated as a phreatic void-filling cement. The tripartite pattern of silica cementation in the Morgan Formation closely resembles the pattern III silica cementation (substrate wall → chert → chalcedony → megaquartz) reported as occurring in Permian pelecypods and brachiopods by Schmitt and Boyd (1981). Length-fast chalcedony and megaquartz in the Morgan Formation resemble cements interpreted as having precipitated in a phreatic groundwater system by Meyers (1977) and Meyers and James (1978). The timing for emplacement of the three generations of silica is uncertain. Three interpretations are possible: (1) Silica cements precipitated in original depositional porosity; either they formed before calcite cementation (less likely) or after calcite cementation *and* after dolomitization (more likely). (2) Silica cements precipitated into pores formed by the leaching out of calcite cements, probably after dolomitization. (3) These three silica phases are not cements at all but formed by the direct replacement of carbonate. In this interpretation, the isopachous chalcedony, some of which contains microdolomite inclusions, probably represents a replacement of a botryoidal, high-Mg calcite, early submarine cement, whereas the equant megaquartz might represent replacement of equant calcite spar. Because the same tripartite zonation of silica phases also occurs in the centers of chert nodules, which clearly postdate calcite cementation (see later section on silica replacement), a cement origin for both the length-fast chalcedony and megaquartz still seems most probable. The predominant occurrence of silica cements at the top of repetitive Morgan sequences and their greater abundance in western outcrop areas imply there are both vertical and lateral controls on their distribution. Perhaps quartz sandstones served as local sources of silica for cementation. If this were the case, then proximity to superjacent quartz sandstone intervals might explain the observed vertical restriction, whereas the general westward decrease in carbonate thickness (and concomitant increase in sandstone thickness) could be responsible for the observed lateral variations.

Dolomitization

Lithologic and Regional Controls

Both lithologic and regional factors have affected dolomite replacement in the Morgan Formation. A strong positive relationship exists between original depositional micrite content and the amount of replacement; 25–100% replacement typifies wackestone and mudstone lithologies, whereas 0–25% characterizes most grainstones and packstones. Consequently, the central, mud-support-dominated part of repetitive Morgan sequences (fig. 3) is the most pervasively dolomitized. In addition to this fabric control, a significant regional (east-west) control seems apparent; samples from the eastern part of the study area (fig. 2) are dominated by clear, very finely crystalline dolomite, whereas those from the western area (fig. 2) are cloudy (inclusion-rich) and coarsely crystalline and display complex compositional zonation.

Eastern Region

Partial replacement by 2- to 100-μm diameter, euhedral to subhedral dolomite rhombs occurs within grain-supported, mud-lean (<5% depositional micrite) lithofacies from the eastern area (fig. 5a). Typically the dolomite is of a clear, "limpid" variety; staining indicates a slightly ferroan composition. The rhombs display strong host-selectivity, nucleating only within micrite and micritic grains (e.g., micrite envelopes on skeletal grains, foraminifera, peloidal grains, grapestone lumps, carbonate mudstone intraclasts). Conversely, coarsely crystalline skeletal grains, fibrous fringe cements, and syntaxial and equant spar cements are not replaced.

Selective replacement of both micritic grains and matrix by very finely crystalline dolomite also occurs in mud-rich (>5% depositional micrite) carbonate lithofacies (fig. 5b). Replacement ranges from 10 to 100%; the greater the amount of coarsely crystalline skeletal material and sparry calcite cement present, the less the dolomite replacement. Petrographic evidence for compositional zonation or overgrowth is rare, although in some of the largest rhombs cloudy central cores are visible. Most dolomite rhombs from the eastern region luminesce a bright red-orange color, and display dull red-brown central cores.

Western Region

Dolomitization is so much more pervasive in samples from the western area that original depositional textures and grain types are difficult to impossible to identify. A few types of skeletal grains (e.g., crinoid columnals and brachiopod valves) retain enough distinctive morphologic form so as to be identifiable after dolomitization. Dolomite rhombs in the western region are much coarser than any observed in the eastern area, ranging from 0.1 to 1.5 mm in diameter; in spite of their large sizes, no saddle-shaped or "baroque" crystals were ever observed. Rhombs are typically cloudy (inclusion-rich) and euhedral to subhedral and display calcite, ferroan calcite, dolomite, and ferroan dolomite overgrowths (figs. 5c,f). Overgrowths range in thickness from 5 to 50 μm, and all are in optical continuity with host rhombs. The general trend is for dolomite composition to become increasingly ferroan toward the outer peripheries of rhombs. In several samples, 10- to 50-μm diameter, cloudy cores are visible within the centers of the coarsest rhombs. Cathodoluminescence petrography reveals an even more complex compositional zonation than was initially recognized in stained thin sections (figs. 5d,f). Calcite zones luminesce bright yellow to yellow-orange; ferroan dolomite is a dull red-brown color; non-ferroan dolomite is a medium-bright to bright intensity, red to red-orange color. Subzones within major compositional zones are common (e.g., zone 3 in fig. 5d). Round central cores are very much more conspicuous under CL as well. The basic CL zonation observed in all dolomite rhombs is (1) dull to medium-bright intensity, red to red-orange central core, (2) bright red-orange to orange middle zone, (3) dull red-brown outer zone. The coarsest rhombs always exhibit the greatest number of compositional bands.

Discussion

Selective dolomite replacement was previously reported by Land (1967), Shinn (1968), and Neal (1969), among others. Neal (1969) observed an order of relative susceptibility of components to dolomitization that is almost identical to the one observed in the Morgan Formation: lime mud matrix > ferroan calcite spar > nonferroan calcite spar > calcite bioclasts. Selective dolomite replacement of micrite is apparently related to the greater number of potential reaction sites in the micrite and the relative stability of low-Mg calcite cements and bioclasts. Models for dolomitization generally invoke either some form of brine reflux (e.g., Adams and Rhodes 1960, Deffeyes and others 1965, Hsü and Siegenthaler 1969) or the mixing of meteoric and marine waters (e.g., Hanshaw and others 1971, Badiozamani 1973, Folk and Land 1975). A newer model by Baker and Kastner (1980, 1981) proposes that dolomitization is favored by a reduction in both dissolved sulfate and silica. It is difficult to ascertain, at the present level of investigation, the mechanism(s) by which Morgan carbonates were dolomitized. However, the evidence does not seem in favor of penecontemporaneous or evaporative dolomitization, for several reasons: (1) There is no evidence for the former presence of any extensive evaporite sequences in the Morgan Formation, although rare evaporite pseudomorphs were reported by Driese and Dott (1984). (2) There is not a strong tie between evaporite pseudomorph occurrences and pervasiveness of dolomitization (e.g., a stenohaline wackestone lithology exhibits the same degree of dolomite replacement as a laminated, evaporitic mudstone). (3) Morgan carbonates are almost exclusively shallow subtidal and were rarely emergent (Driese and Dott 1984). A mixing-zone model seems more likely for the dolomitization of Morgan Formation carbonates based on dolomite crystal morphologies. The finely crystalline, euhedral, limpid dolomite rhombs characteristic of the eastern study area probably formed under slow rates of crystallization, an absence of foreign ions competing for structural sites, and low Mg/Ca ratios (Folk and Land 1975). Folk and Land (1975) suggested that fluctuation from dolomite to calcite precipitation occurs in waters with Mg/Ca ratios around 1; hence, the alternation of calcite and dolomite overgrowths grown syntaxially on coarse, inclusion-rich rhombs in the western study area could reflect slight shifts in the Mg/Ca ratio above and below the critical 1:1 ratio. The common occurrence of ferroan dolomite and ferroan calcite as overgrowths suggests that at least these later stages of dolomitization occurred within a low Eh, phreatic environment containing abundant reduced iron (Evamy 1969).

The reasons for the observed east-west gradient in dolomitization of Morgan carbonates are also unclear. Dunham and Olson (1978, 1980) alluded to an east-west paleogeographic control on dolomitization of Middle Paleozoic carbonates in the Cordilleran miogeocline of central Nevada. They proposed that meteoric water derived from subaerially exposed tracts in the eastern part of the miogeocline migrated westward to mix repeatedly with marine waters; hence, there is a consistent temporal and spatial distribution of limestone towards the west and dolomite towards the east. Although an east-west paleohydrologic control may have governed Morgan dolomitization as well, the observed spatial distributions of less extensive dolomitization in the east (shelfward) and more pervasive towards the west (miogeocline) are not as predicted by Dunham and Olson's (1978, 1980) model. Another possible explanation is that lateral variation in dolomitization is related to east-west changes in depositional facies–Morgan carbonates thin greatly at the expense of intervening quartz sandstones when traced westward, and the quartz sandstones assume a more shallow-marine rather than eolian character (Driese and Dott 1984). The decrease in thicknesses of carbonate packets and increase in thicknesses of intercalated (permeable?) quartz sandstone units might have facilitated dolomitization. Alternatively, an east-west gradient in maximum burial temperatures, as deduced from conodont alteration indices (Epstein and others 1977), may have had an influence. Conodonts from eastern Morgan carbonate samples suggest maximum burial temperatures in the range of $<$50–80°C, whereas in the western area, temperatures were as high as 85–170°C. However, if the dolomite had formed at these elevated temperatures during deep burial, then saddle-shaped or "baroque" crystals would have been expected to form (Radke and Mathis 1980), yet none were ever observed; it therefore seems unlikely that the dolomite is of high-temperature, deep-burial origin.

The timing of dolomitization cannot be precisely determined. Although it clearly postdates calcite spar cementation, it could either predate or be coeval with silica replacement. If a mixing-zone interpretation for Morgan dolomitization is correct, then dolomite could have precipitated early while meteoric-marine fronts repeatedly passed through the section with each transgression and regression. If the dolomite formed at deep-burial depths (which seems less likely), then it might have precipitated very late, but probably before Cenozoic tectonism.

Dedolomitization

Dedolomite occurs rarely within the coarsely dolomitized facies of the western study area. The most common expression of dedolomite is as a polycrystalline mosaic of calcite pseudospar and microspar in which the rhombic outlines of the former dolomite rhombs are still preserved (fig. 5g). Incomplete dedolomitization commonly appears as a mottled, or "swiss-cheese," fabric. Dedolomite is very sporadic in occurrence; it typically grades both laterally and vertically into unreplaced dolomite in outcrop, over distances of no more than a few meters. It is volumetrically insignificant compared to dolomite, generally replacing much less than 1% of the dolomite in the samples examined. Organic inclusions and ferric iron that were probably originally contained within the dolomite appear to have been forcibly displaced toward the outer periphery of each rhomb, thus forming opaque rims (fig. 5g). Calcite-replaced dolomite rhombs luminesce a dull brown to medium-bright intensity, yellow-brown color that is clearly distinguishable from the bright red to red-orange luminescence of the unreplaced dolomite; opaque outer rims do not luminesce.

Discussion

Dedolomitization is believed to occur when solutions with a high Ca/Mg ratio react with dolomite to form calcite (Evamy 1967). Three mechanisms were proposed for the process: (1) surficial chemical weathering, (2) reaction of dolomite with solutions containing sulfate, (3) changes in the Mg/Ca ratio of dolomitizing pore-fluids (Katz 1968). The mode of formation of Morgan dedolomite is difficult to ascertain because of its sporadic and seemingly unpredictable distribution. The common occurrence of Morgan dedolomite as coarse rhombs composed of polycrystalline aggregates of calcite pseudospar and microspar points to direct replacement of dolomite by calcite, rather than by dissolution-precipitation (Evamy 1967). The opaque rims fringing dedolomite rhombs might have formed when excess ferrous iron was expelled from the rhomb during calcite replacement and reprecipitated as colloidal ferric hydroxide (Evamy 1963). Alternatively, the ferric iron rims could simply be a consequence of the fact that the ferrous iron content of Morgan dolomite rhombs typically increases outward from the centers of rhombs. The timing for the formation of calcitized dolomite rhombs is interpretable in two different ways: (1) Early diagenetic dedolomitization could have occurred almost coevally with dolomitization. If the chemical environment was unstable, a slight shift in the Mg/Ca ratio produced a change from dolomite precipitation to dedolomitization. (2) Late-diagenetic dedolomitization could have occurred as a near-surface weathering feature, which therefore postdates all other diagenetic events, including late-stage calcite spar cementation. Vertical distributional controls on dedolomitization in repetitive Morgan sequences are not apparent; the restriction of dedolomite to western, coarsely dolomitized carbonate facies suggests some lateral control(s), perhaps related to the same factor(s) governing the observed east-west lateral variation in dolomite replacement.

Silicification

Lithologic and Regional Controls

Silica replacement of carbonate grains, matrix, cements, and dolomite appears to have been governed by depositional texture and geographic location in much the same manner as was the dolomite; 0–5% silica replacement typifies most grainstone and packstone lithologies, whereas some wackestones and mudstones are up to 30% chert replaced. Consequently, the mud-support-dominated central parts of repetitive Morgan sequences are the most extensively replaced by silica, whereas basal and upper grain-support-dominated parts are generally not replaced (fig. 3). Silicification is also much more pervasive in the western part of the study area, where whole beds of carbonate of any texture are replaced.

Eastern Area

Skeletal grains are partially to completely replaced by chert and/or chalcedony in the eastern area (fig. 6a). Skeletal grain type susceptibility to silica replacement is echinoderm $>$

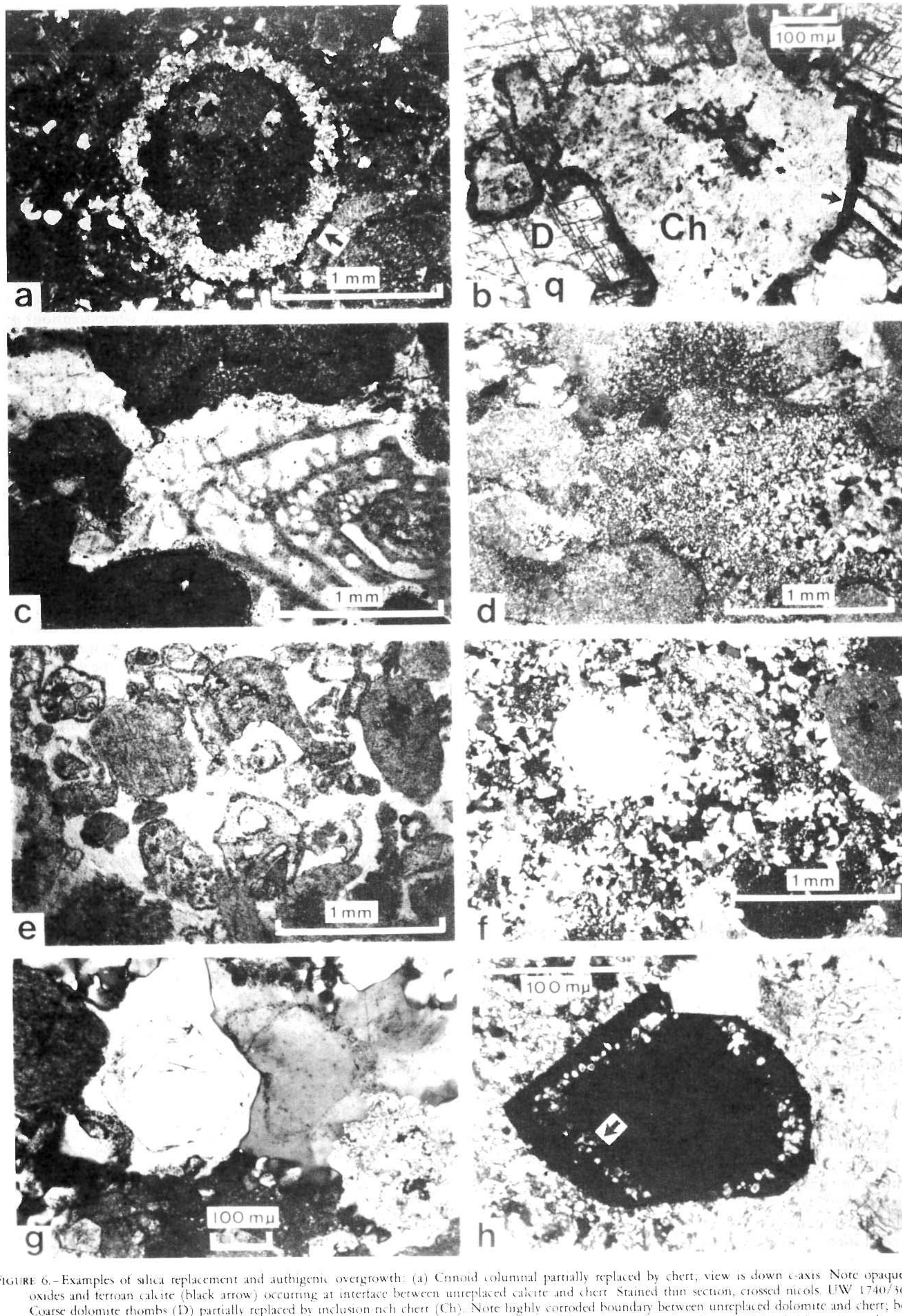

FIGURE 6.–Examples of silica replacement and authigenic overgrowth: (a) Crinoid columnal partially replaced by chert; view is down c-axis. Note opaque iron oxides and ferroan calcite (black arrow) occurring at interface between unreplaced calcite and chert. Stained thin section, crossed nicols. UW 1740/36. (b) Coarse dolomite rhombs (D) partially replaced by inclusion-rich chert (Ch). Note highly corroded boundary between unreplaced dolomite and chert; boundary accentuated by abundant opaque iron oxides and organic matter (black arrow). Scattered detrital quartz sand grains (q). Stained thin section, plane light. UW 1740/37. (c) Fusulinid replaced by chert; note adjacent dolomitized crinoid grains (dark color). Stained thin section, plane light. UW 1740/38. (d) Same field of view as in 6c, under crossed nicols. UW 1740/38. (e) Chert and megaquartz replacement of former carbonate cements (?) and some grains in bioclastic grainstone. Stained thin section, plane light. UW 1740/39. (f) Same field of view as in 6e, under crossed nicols; note dolomitized crinoid grains at various extinction positions. UW 1740/39. (g) Anhedral quartz overgrowths on detrital quartz sand grains in sandy dolomite. Stained thin section, crossed nicols. UW 1740/39. (h) Detrital quartz sand grain displaying subhedral authigenic quartz overgrowth (at extinction position). Outline of original grain marked by band of dolomite inclusions (black arrow). Coarse dolomite completely envelopes grain. UW 1740/40.

brachiopod > bryozoan > all other grains. The chert and chalcedony are colorless to pale yellow-brown in transmitted light; both contain rare 2- to 10-μm-diameter microdolomite inclusions. Chalcedony fiber-bundles are randomly oriented within replaced grains and are both length-slow and length-fast. Concentrations of opaque ferric iron, organic material, and ferroan calcite all occur at the boundaries between unaltered carbonate and replacive silica (figs. 6a,b).

Silica replacement occurs on a megascopic scale in the form of 1- to 40-cm-diameter chert nodules and 1- to 15-cm-thick, laterally discontinuous chert beds. Depositional stratification, burrow structures, and skeletal grains are all directly traceable from the chert nodules and beds into adjacent carbonate. Locally abundant, laterally discontinuous concentrations of skeletal grains were favored sites for silica replacement, as were spiculitic wackestone and mudstone lithologies. Some nodules contain length-fast chalcedony as pore-filling cements filling shrinkage cracks within their interiors; they were generally followed by equant megaquartz cements. All silica-replaced carbonate exhibits dull brown luminescence or is nonluminescent.

Western Area

In the western part of the study area, silica replaces not only carbonate grains and matrix, but also calcite cements and dolomite. Dolomite replacement was probably preceded by dissolution or corrosion (fig. 6b). Silica replacement of skeletal grains typically preserved even the most delicate details of skeletal microstructure (figs. 6c,d). Though not provable, wholesale replacement of former carbonate cements is also inferred in figures 6e and 6f because of a lack of fabrics suggestive of growth into open pores, as are evident in figures 4g and 4h. Authigenic quartz overgrowth on quartz sand and silt grains is restricted to the western area. Anhedral overgrowths predominate in very sandy carbonate lithologies (fig. 6g), whereas euhedral to subhedral overgrowths are best developed in lithologies in which the quartz grains are not in direct grain-to-grain contact with each other (fig. 6h). Outlines of host detrital grains are commonly accentuated by rings of disseminated microdolomite inclusions (fig. 6h). The crystal habit adopted by most overgrowths is a doubly terminated hexagonal prism. Overgrowths do not luminesce under the beam-current conditions normally used to examine the carbonate facies, but many detrital nuclei luminesce a bright blue color.

Discussion

The primary controls on the distribution of replacive silica in repetitive Morgan sequences appear to have been skeletal grain composition and micrite matrix content. Selective silica replacement of skeletal grains and lack of replacement of neighboring intraclasts and micrite matrix might be explained by the presence of organic decomposition products within the skeletal grains that created favorable microenvironments conducive for silica replacement (Jacka 1974). The occurrences of both length-slow and length-fast chalcedony as skeletal replacements in the Morgan are somewhat puzzling; the presence of length-slow chalcedony might serve to differentiate cases of direct replacement of skeletals from those in which dissolution/precipitation (i.e., cementation) probably occurred (Folk and Pittman 1971). The abundant chert replacement observed in Morgan wackestones and mudstones might be a function of their higher sponge-spicule contents; the sponge spicules may have served as local sources of silica (Meyers 1977). It is difficult to ascertain, at the present level of investigation, the mechanisms by which Morgan carbonates were silicified. Previous researchers have postulated silica replacement within meteoric phreatic lenses (Meyers 1977) or by mixing marine and meteoric waters (Knauth 1979). The relative timing of silicification is likewise unclear. Fabric relationships in which silica replaces dolomite or in which dolomite inclusions float within silica seem to imply that at least some silicification postdates dolomitization. However, many of the microdolomite inclusions, which are common within silica-replaced grains of former high-Mg calcite composition (e.g., echinoderms, brachiopods, bryozoa), could have formed by dolomite exsolution during silica replacement (Jacka 1974); thus their presence need not require that silicification postdate dolomitization. A recent dolomitization model proposed by Baker and Kastner (1981) seems to imply that silicification may in fact be conducive to dolomitization—thus it is also possible that the two occurred coevally. The greater pervasiveness of silicification observed in the western study area appears to coincide with greater dolomitization in that same region; hence, the factor(s) responsible for the lateral variability of dolomitization may also have affected silicification.

Evaporite Diagenesis

Rare evaporite pseudomorphs were observed within laminated mudstones and in wackestones containing abundant peloidal grains or restricted faunal assemblages (e.g., ostracods, lingulid brachiopods, gastropods, and encrusting foraminifera). The pseudomorphs typically occur in the upper part of repetitive Morgan sequences, directly beneath the uppermost grain-supported facies (fig. 3). They are generally more common in carbonates from the eastern part of the study area than from the western part. They do not occur in extensive, laterally continuous beds but in 1- to 10-cm-thick, patchy zones. There are three principal morphologic forms of pseudomorphs, each of which are replaced to varying degrees by nonferroan calcite, ferroan dolomite, length-slow chalcedony, and megaquartz. The first morphologic type is a 0.25- to 1.5-mm-diameter cubic form, which is probably pseudomorphous after halite (fig. 7a). The second is a globular, irregularly shaped, massive form, possibly pseudomorphous after nodular mosaic anhydrite (figs. 7b,f). The third and most common type is a stellate form that consists of 1- to 20-mm-diameter, radiating swarms or aggregates of lathlike crystals, many of which exhibit "swallowtail" terminations (figs. 7c,d,e,g). It is most likely pseudomorphous after gypsum (selenite) rosettes. No relict evaporite minerals per se were ever observed in Morgan carbonates. The pseudomorphs, in general, constitute less than 10% of most samples in which they are observed. Fabric relationships suggest replacement of evaporites in the following sequence: nonferroan calcite pseudospar → ferroan dolomite → length-slow chalcedony and megaquartz (fig. 7f). Microdolomite inclusions occur in some chalcedony- and megaquartz-replaced pseudomorphs (figs. 7c,d). Although calcite-crystal fabrics in a few pseudomorphs are suggestive of void-fillings, the majority suggest a direct replacement origin for the calcite (fig. 7g). Under CL, calcite pseudospar luminesces a dull to medium-bright intensity, brown to yellow-brown color; ferroan dolomite luminesces dull red-brown; chalcedony and megaquartz generally do not luminesce (fig. 7a). Many pseudomorphs also contain cloudy, semiopaque masses (organic material?) that luminesce an intense bright blue color.

Discussion

Although marine evaporites occur most commonly in supratidal (sabkha) environments (Shearman 1963, Murray 1964, Kinsman 1969), care should be exercised against over-

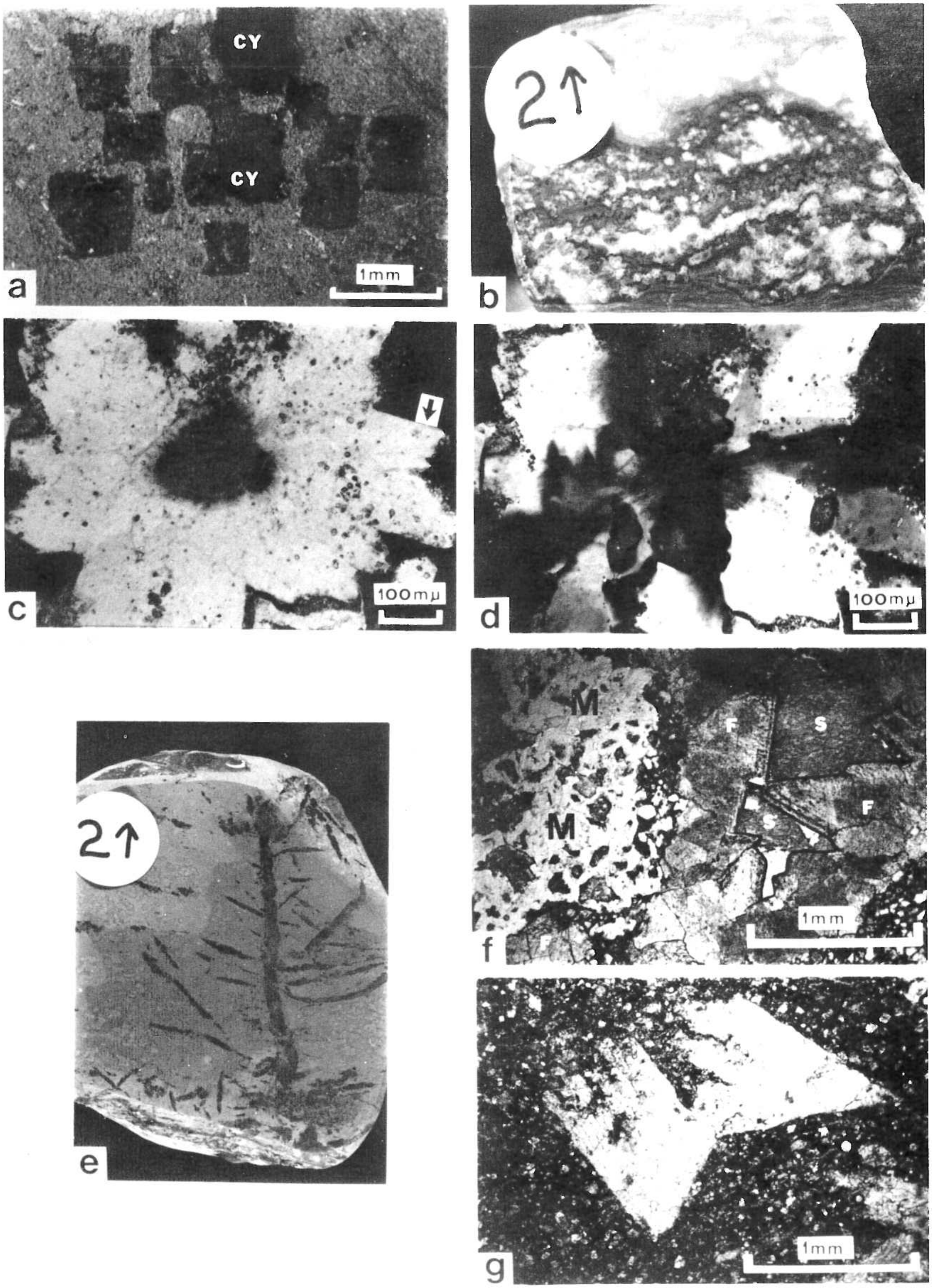

FIGURE 7.- Examples of evaporite replacement. (a) Calcite-replaced, cubic halite pseudomorphs, viewed under CL. Calcite luminesces dull yellow-brown; matrix (partially dolomitized) is bright red-orange; patches of chalcedony (CY) do not luminesce. UW 1740/41. (b) Dolomitic mudstone (darker color) containing irregularly shaped sparry calcite bodies (lighter color), which are probably pseudomorphs after nodular mosaic anhydrite. White circle is 2 cm in diameter. Polished slab cut normal to bedding. UW 1740/22. (c) Megaquartz-replaced, rosettelike pseudomorph after gypsum. Note "swallowtail" terminations on some crystals (black arrow), abundant microdolomite inclusions floating within megaquartz, and cloudy core region. Thin section, plane light. UW 1740/42. (d) Same field of view as in 7c, under crossed nicols. Length-slow megaquartz pseudomorphs lathlike habit of former gypsum crystals. UW 1740/42. (e) Calcite and chert replacement, probably pseudomorphous after selenite, in a dolomitic mudstone. White circle is 2 cm in diameter. Polished slab cut normal to bedding. UW 1740/43. (f) Former anhydrite nodule replaced by calcite pseudospar (S), ferroan dolomite (F), and megaquartz (M); note abundant calcite and dolomite inclusions floating within megaquartz. Stained thin section, plane light. UW 1740/44. (g) "Swallowtail" pseudomorph of gypsum replaced by calcite pseudospar. Matrix is partially dolomitized. Stained thin section, plane light. UW 1740/45.

generalizing that their presence in a stratigraphic sequence is indicative exclusively of supratidal deposition (Dean and others 1975). Driese and Dott (1984) proposed that Morgan carbonate facies containing evaporite pseudomorphs formed under low-energy, shallow subtidal conditions in hypersaline lagoons or bays on the basis of an absence of probable supratidal or intertidal features (e.g., desiccation polygons, fenestral or birdseye fabrics, algal-mat structures, stromatolites). Evaporite replacement might have been concurrent with several of the diagenetic events previously alluded to on the basis of similarity of relative time order, luminescent properties, and mineralogy: (1) It is possible that calcite pseudospar replacement of evaporites and calcite spar cementation are related. (2) Ferroan dolomite replacement of evaporites could have been coeval with the more widespread dolomitization that affected most Morgan carbonate facies. (3) Length-slow chalcedony and megaquartz replacement of evaporites probably was concurrent with silica replacement of carbonate grains, matrix, and cements. The lateral restriction of evaporite pseudomorphs to the eastern part of the study area seems to indicate more arid conditions and more impaired marine circulation on this part of the shelf. Their vertical restriction in repetitive Morgan sequences, according to Driese and Dott (1984), probably reflects upward shoaling of carbonate environments coincident with increasingly impaired water circulation.

Pressure Solution

Pressure solution shows as stylolites, fitted-fabric textures, and nonsutured solution seams. These features generally occur within Morgan carbonate strata that were relatively unaffected by either extensive dolomitization or silicification; hence, pressure-solution phenomena are more abundant in the eastern part of the study area. Evidence for physical compaction (e.g., crushed ooids, broken micrite envelopes) is relatively rare. Sutured stylolitic surfaces are the most easily recognized form of pressure solution (fig. 8a,b,g). Stylolites are most abundant in the basal grain-support-dominated lithofacies within repetitive Morgan sequences (fig. 3); about 25% of the grain-rich carbonates examined contain some stylolites. Most stylolites have a dominant horizontal to subhorizontal orientation; maximum observed vertical amplitudes are about 2 cm. The insoluble residues that occur along stylolitic surfaces consist of very fine detrital quartz sand and fine-to-coarse quartz silt, heavy mineral grains, very finely crystalline dolomite rhombs, phosphatic skeletal material, and opaque iron oxides. Many dolomite rhombs are truncated against stylolites (fig. 8g). Calcite spar cements associated with stylolitic surfaces are distinctly different from all other cements when viewed under CL, because they emit a very bright luminescence of an intensity rivaled only by that of calcite spar fracture-filling cements (e.g., compare figs. 8b and 8h). Solution compaction is also expressed as overly close packed or fitted fabrics, in which all grain-to-grain (or grain-cement) boundaries are interpenetrative, yet distinctly sutured stylolitic surfaces are absent (figs. 4c, 8c). This feature is also most common in the basal part of repetitive Morgan sequences (fig. 3); almost every bioclastic grainstone showed some grain-to-grain interpenetration.

An overly close packed fabric can also be the final product of solution compaction in the mud-supported lithologies that dominate the central part of repetitive Morgan sequences (fig. 3). Most grains are not in direct grain-to-grain contact, but are separated by thin (micron-scale) microstylolite swarms that contain abundant insoluble residue (fig. 8d). Where significant loss of micrite matrix is inferred, a crude fabric is developed that resembles a foliation or a fissility. This fabric, roughly parallel to bedding, is evident in figure 8d; hand samples show a tendency to break along these planes. These features are also observed in outcrop and are recognized where carbonate strata thin across rigid, insoluble chert nodules and lenses. Solution compaction in very mud-rich lithologies ($>80\%$ depositional micrite) is typically expressed as swarms of wispy, anastomosing solution seams, only a few microns wide (fig. 8e); fine insoluble residues, including some clay, occur along solution seams. About 50% of the mud-rich carbonates examined contain some nonsutured solution seams.

Discussion

Pressure solution is an important diagenetic process in many carbonate rocks, especially with respect to the reduction of inter- and intraparticle porosity (Wagner and Matthews 1982). Although experimental work indicates that modern carbonate sediments are compactable (Shinn and others 1977), this does not appear to have been an important process in Morgan carbonates. The vertical distribution of pressure-solution phenomena in repetitive Morgan sequences is probably a function of the different types of responses of different lithologies to vertically directed overburden (burial) stresses (Wanless 1979); sutured seam solution (stylolitization, fitted fabric textures) dominates in carbonates that had structural resistance to stress, whereas nonsutured seam solution occurs in structurally responsive, grain-poor carbonates that contained appreciable amounts of clay and platy silt. Fabric relationships suggest that most pressure solution postdated calcite spar void-filling (e.g., figs. 4c, 8a) and dolomitization (fig. 8g), although a minor amount of very finely crystalline dolomite could have formed from Mg^{2+} released during pressure solution (Wanless 1979). The lateral restriction of pressure-solution features to the eastern part of the study area is a consequence of the greater pervasiveness of dolomitization and silicification in the western area and is further evidence that pressure solution postdated these features. The similarity of luminescent properties between the calcite spar associated with stylolitic surfaces and that in late-stage fracture-fillings is interpretable in two ways: (1) Pressure solution served as a local source of calcite for late-stage cementation. Some of the calcite reprecipitated along the stylolites and some migrated into late-stage porosity (e.g., fractures, intercrystalline and biomoldic porosity). (2) The brightly luminescent calcite spar came into the stylolites much later and therefore is not related to pressure solution.

Late-Stage Cementation

Vertically to subvertically oriented, 0.1- to 5-m-wide fractures crosscut carbonate grains, matrix, cements, and dolomite rhombs. They occur equally in all facies within repetitive Morgan sequences (fig. 3) and across all parts of the study area. Typically the fractures are filled with 0.05- to 0.5-mm-diameter, bladed to equant calcite spar cements. The wider fractures (2–5 mm) commonly display two generations of cement growth–an early bladed cement, elongate normal to the fracture wall, followed by equant calcite spar cements, which occlude most of the remaining porosity. Staining indicates that both ferroan and nonferroan calcite spar cements occur within the same fractures; in general, ferroan calcite is the latest form. Nonferroan calcite fracture-filling cements luminesce an intense bright yellow to yellow-orange color under CL (fig. 8h), far brighter than any other cement type, except for cements associated with stylolitic surfaces (fig. 8b) or occluding intercrystalline porosity within coarsely dolomitized facies (fig. 5h). In several samples,

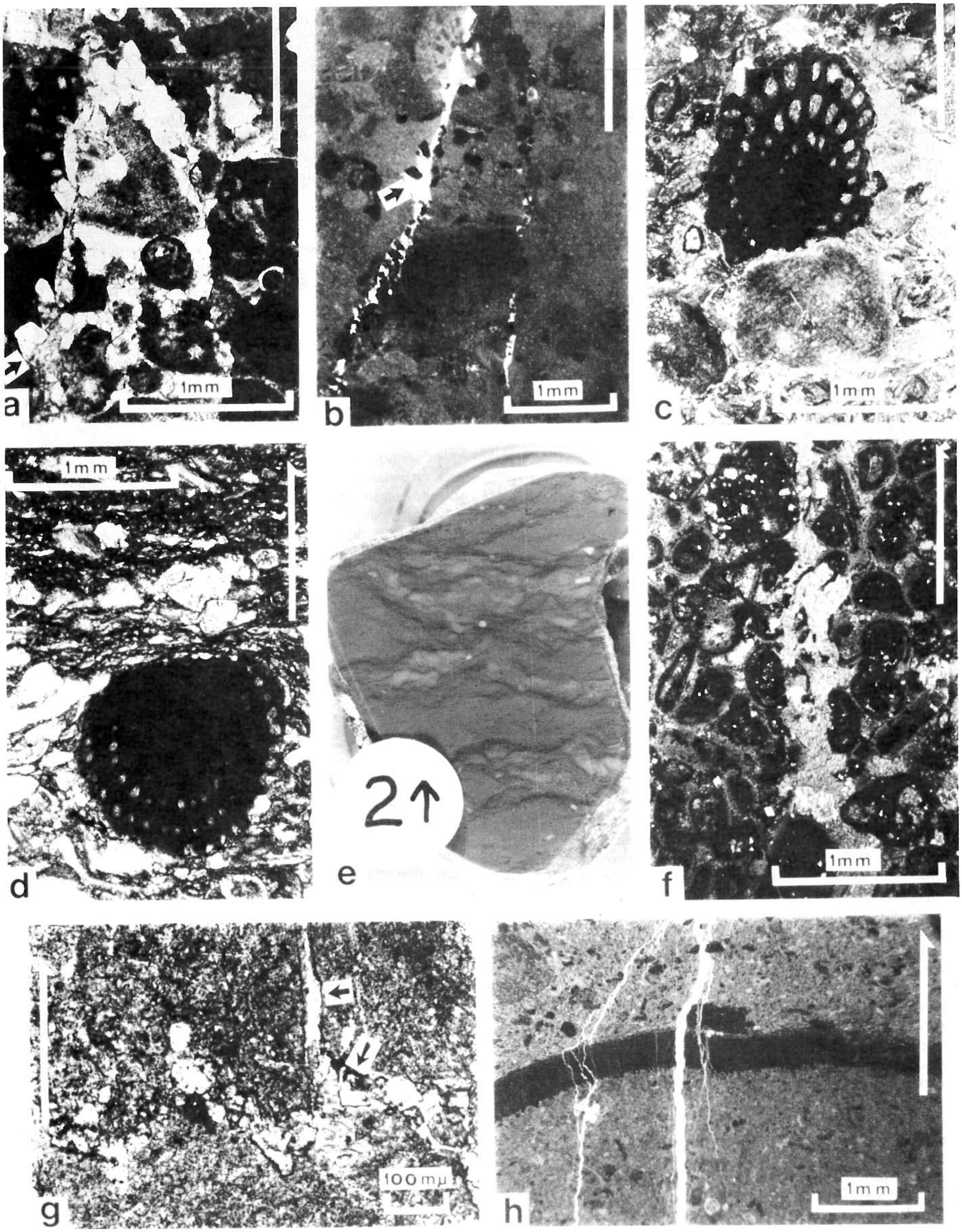

FIGURE 8.–Examples of pressure solution and late-stage cementation: (a) Macrostylolite developed within bioclastic grainstone. Note truncated skeletal grains, insoluble residue (chiefly detrital quartz sand), and late calcite spar cement (black arrow) associated with stylolitic surface. Thin section, plane light. UW 1740/46. (b) Part of same stylolite seen in 8(a), viewed under CI. Late calcite spar cement luminesces an intense bright yellow color; most of the adjacent carbonate grains are dull yellow-brown. Black arrow points to same location as in 8a. UW 1740/46. (c) Interpenetrating contact between fusulinid grain (dark) and crinoid columnal (bright). Stained thin section, crossed nicols. UW 1740/47. (d) Solution compaction along microstylolite swarms in bioclastic wackestone, resulting in overly close packed fabric, note how fabric wraps around relatively insoluble fusulinid. Thin section, plane light. UW 1740/48. (e) Wispy, anastomosing, nonsutured solution seams in burrowed wackestone. Dark coloration of solution seams caused by insoluble residue. White circle is 2 cm in diameter. Polished slab cut normal to bedding. UW 1740/49. (f) Late calcite spar cement-filled fracture cross-cutting peloids, intraclasts, fringe cements, and earlier equant calcite spar cements. Stained thin section, plane light. UW 1740/28. (g) Sutured stylolitic surface containing abundant very finely crystalline dolomite rhombs. Note how several rhombs are truncated by surface (black arrows). Stained thin section, plane light. UW 1740/24. (h) Late-stage fracture-filling crosscutting brachiopod valve (dark body) and micrite matrix, viewed under CI. Late calcite spar cement luminesces an intense bright yellow color, matrix is dull yellow-brown, brachiopod does not luminesce. UW 1740/50.

CL revealed the presence of more than one generation of filled fractures, each generation luminescing a slightly different color. "Hidden" fracture-filling cements grown in optical continuity with associated early equant calcite spar cements were also revealed by luminescence.

Discussion

Fracture-filling cementation constitutes the latest diagenetic event, except for possible near-surface calcite dissolution and dedolomitization. Fracturing might be related to regional Cenozoic tectonism because of its widespread lateral and vertical distribution in Morgan carbonates. On the basis of the strong similarity of their luminescent properties, ions released during pressure solution are hypothesized as possible sources for the brightly luminescent calcite cements that fill fractures and intercrystalline and biomoldic porosity; the intensity of their luminescence suggests Fe/Mn ratios less than 0.5 (Frank and others 1982). The occurrence of ferroan calcite juxtaposed in the same pores with luminescent, nonferroan calcite implies two generations of late-stage cementation within differing chemical environments–the late ferroan calcites probably precipitated in low Eh, meteoric phreatic environments containing abundant reduced iron (Evamy 1969).

Porosity Evolution

Porosity evolution in Morgan carbonates has been exceedingly complex, and many questions remain unanswered concerning the mechanisms of porosity formation and occlusion, and their relative timing. Almost all original depositional porosity of both inter- and intraparticle types was probably occluded by early fibrous to bladed fringe cementation and later syntaxial and equant calcite spar pore-filling. Consequently, the grainstones and mud-lean packstones from the eastern part of the study area are tight. Possible episodes of development of secondary porosity in the Morgan and their approximate relative timing include (a) evaporite dissolution (pre-equant calcite spar cementation), (b) calcite and dolomite dissolution/corrosion (postdolomitization, pre-silica cementation and/or replacement), (c) calcite dissolution (post-silica cementation and/or replacement, pre-late-stage calcite spar cementation), (d) fracturing (post-pressure solution, pre-late-stage calcite spar cementation), (e) near-surface leaching of calcite and weathering of dolomite (post-late-stage calcite spar cementation). Of these, events (c) and possibly (e) are most responsible for the observed porosity that still is unfilled. Coarsely dolomitized carbonates of all textures from the western part of the study area have the highest present porosities; combined intercrystalline and biomoldic porosities of 2–5% are typical but range as high as 10–15% in 100% dolomitized facies (fig. 5h). This type of outcrop facies closely resembles those which constitute the producing zones in the subsurface Morgan Formation of the Western Overthrust Belt (Verville and others 1973; Picard 1977).

SUMMARY

Relative Timing of Diagenetic Events

The relative time-order for Morgan diagenetic features is summarized in figure 9. Morgan diagenesis is divided into four major phases. During the earliest part of the early phase, micritization and fibrous to bladed fringe cementation occurred in a submarine environment. Syntaxial and equant calcite spar cements precipitated later, probably in a meteoric phreatic environment. The middle diagenetic phase was first dominated by dolomitization, possibly occurring in a meteoric-marine mixing-zone, and then by silica replacement/cementation, which could have formed in either a meteoric phreatic or mixing-zone environment. The late phase was dominated by pressure solution, fracturing, and late-stage calcite spar cementation; pressure solution was probably caused by overburden (burial) stresses, whereas the fractures may be of tectonic origin. A near-surface diagenetic phase is also postulated to include surficial weathering processes.

Influence of Depositional Environments upon Diagenesis

An important question is whether deposition in different environments had a perceivable influence on diagenetic history. Heckel (1982) recently proposed that diagenetic differences do exist between the various depositional facies that comprise midcontinent Pennsylvanian cyclothems. He suggested that the thin, basal "transgressive" limestones were more deeply buried and compacted before cementation than were the thick, upper, "regressive" units, which experienced a much greater variety of diagenetic events. Diagenetic features unique to upper regressive limestones included early marine fringe cementation, leaching by meteoric groundwater, and blocky calcite spar, ferroan calcite, and ferroan dolomite cementation, respectively. The black shale "core" facies of midcontinent cyclothems, which occurs sandwiched between lower transgressive and upper regressive carbonate facies, was inferred to have behaved as an impermeable seal, effectively separating two distinctly different diagenetic environments.

The vertical progression of Morgan carbonate facies described by Driese and Dott (1984) bears some resemblance to midcontinent cyclothems in that it is conspicuously repetitive and consists of a thin, lower transgressive interval and a thicker, upper regressive one. From base to top, the vertical sequence consists of (fig. 3) (a) sharp, basal transgressive surface, (b) shallow subtidal, grain-supported, normal marine facies, (c) shallow subtidal, mud-supported, normal marine facies, (d) shallow subtidal(?), mud-supported, restricted to episodically hypersaline facies, (e) shallow subtidal to rare intertidal, grain-supported, restricted facies, (f) intertidal (foreshore), mixed terrigenous-carbonate facies, gradational into (g) overlying siliciclastic eolian dune sandstones. However, vertical diagenetic changes solely related to differences in Morgan depositional environments do not appear as important as changes related to depositional texture, micrite matrix content, and grain type. There are several lines of evidence to support this: (1) Grain-rich carbonate lithofacies sampled from different positions within the same Morgan carbonate interval (repetitive sequence) and from the same region all seem to have experienced an identical calcite cementation history, irrespective of their environmental interpretation. Equant calcite spar cements are widespread because their distribution was dependent solely upon pore-space distribution and not upon the types of substrates available, in contrast to fringe and syntaxial spar cements. (2) Mud-rich carbonate lithofacies sampled from different positions within the same Morgan carbonate interval and from the same region all seem to exhibit the same degree of dolomite replacement, irrespective of their environmental interpretation. For example, a bioclastic wackestone deposited in a low-energy, normal marine environment displays the same degree of selective dolomite replacement of micritic matrix as an abiotic, evaporite pseudomorph-bearing wackestone deposited in a low-energy, hypersaline environment. There does not appear to be a strong positive relationship between hypersaline or restricted carbonate facies and pervasiveness of dolomitization. (3) Silica replacement and cementation were controlled by depositional texture, micrite matrix content, and grain types present. All mud-rich wackestones and mudstones display at least some chert replacement of carbonate, irrespective of stratigraphic position and environmental interpretation. Environmental factors were responsible for the distribution of high-Mg calcite skeletal grains, which are most susceptible to silica replacement. Silica cementation is primarily confined to grain-supported, mud-lean lithofacies, within original depositional porosity or in secondary solution porosity. The occurrence of silica cements is not confined to any one particular environment or to a single stratigraphic position.

It would be difficult to ascertain, with the present data, whether or not specific diagenetic features are correlative with discrete transgressive and regressive events or represent the sum total effect of many such events overprinted one upon the other. The complexity of Morgan paragenetic relationships and my inability to correlate certain features (e.g., luminescent zonation in calcite cements and dolomite rhombs) suggest that there has been much overprinting of one diagenetic event upon another, making interpretation difficult.

Lateral Distribution of Diagenetic Features

The lateral distribution of Morgan diagenetic features is summarized in table 1. Many early-phase diagenetic features (e.g., micritization, calcite cementation) appear equally distributed over both the western and eastern regions. Middle-phase features (e.g., dolomitization and silicification) are much more pervasive in the western study area, where they tend to obscure earlier diagenetic events. Late-phase features (e.g., pressure solution, fracturing, late-stage cementation) appear equally distributed in both regions. Porosity is generally higher in the western area, especially within dolomitized facies. Paleohydrologic differences, depositional facies changes, and burial temperature/depth differences are all postulated as possible factors that might account for observed regional differences in diagenesis.

Vertical Distribution of Diagenetic Features

The vertical distribution of diagenetic features, as it occurs within the repetitive Morgan sequence previously defined by Driese and Dott (1983), is summarized in figure 3. The primary vertical controls on diagenesis appear to be depositional texture, which influences porosity, and permeability, micrite matrix content, and grain type. The distribution of micritized grains was probably influenced by light penetrability. Fringe cement distribution seems related to the distribution of former aragonitic host substrates. Syntaxial spar cementation required abundant echinodermal host substrates. Equant calcite spar cement distribution was influenced only by the availability of open pore-spaces. Evaporites (now pseudomorphs) seem confined to facies interpreted as low-energy, shallow subtidal, hypersaline lagoon or tidal pond deposits. Dolomite replacement was strongly selective for mud-supported facies that contained abundant micrite matrix or micritic grains. Selective silica replacement for mud-supported facies probably reflects their higher initial sponge-spicule contents. Silica replacement was also selective for former high-Mg calcite skeletal grains and former evaporites. Stylolites were restricted to grain-rich, grain-supported facies, whereas nonsutured solution seams and microstylolite swarms dominated matrix-rich, mud-supported

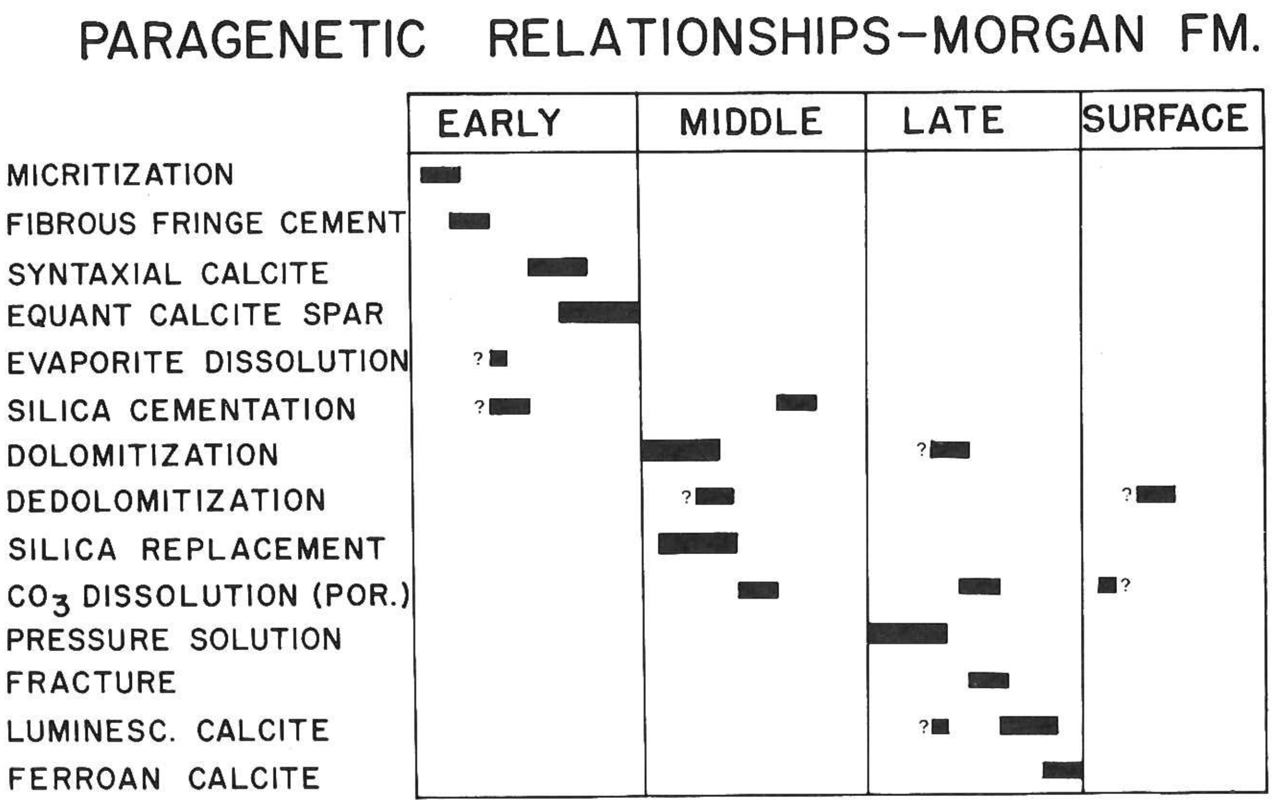

FIGURE 9.—Summary of relative time order inferred for Morgan diagenetic features. Events whose occurrence or timing are in question are accompanied by question marks. "Early," "middle," and "late" are used only in a relative time sense.

TABLE 1
Summary of Regional Distribution of Diagenetic Features

	*Eastern Area**	*Western Area**
Micritized carbonate grains	Common	Common
Fibrous to bladed fringe cement	Rare	Not observed
Syntaxial calcite spar cement	Common	Rare
Equant calcite spar cement	Common	Common
Evaporite pseudomorphs	Rare	Not observed
Silica cements	Not observed	Rare
Dolomite replacement	Common (very fine x̄lline)	Abundant (coarsely x̄lline)
Dedolomite	Not observed	Rare
Silica replacement of carbonate	Common	Abundant
Authigenic quartz overgrowths	Not observed	Common
Pressure solution features	Abundant	Common
Fractures	Common	Common
Late-stage calcite spar cements	Common	Common
Porosity	Low (<5%), primarily in dolomitized facies	Moderate (2-5%, as high as 10-15%), primarily in dolomitized facies

*"Eastern" and "western" areas are defined in figure 2.

rocks. Fractures and late-stage calcite cementation showed no preferred vertical distribution. Present porosity is lowest in calcite-cemented grainstones and highest in pervasively dolomitized wackestones.

ACKNOWLEDGMENTS

This paper constitutes part of a Ph.D. dissertation completed while at the University of Wisconsin–Madison, under the supervision of Professor Robert H. Dott, Jr. The author gratefully acknowledges the generous financial assistance provided by the Atlantic Richfield Company. Lloyd C. Pray (University of Wisconsin–Madison) read earlier drafts of this manuscript and provided many useful comments and suggestions.

APPENDIX

All specimens figured in this paper are on repository at the Department of Geology and Geophysics, University of Wisconsin–Madison, under catalog number 1740.

REFERENCES CITED

Adams, J. E., and Rhodes, M. L., 1960, Dolomitization by seepage refluxion: American Association of Petroleum Geologists Bulletin, v. 44, p. 1912-20.

Badiozamani, K., 1973, The dorag dolomitization model: application to the Middle Ordovician of Wisconsin: Journal of Sedimentary Petrology, v. 43, p. 965-84.

Baker, P. A., and Kastner, M., 1980, The origin of dolomite in marine sediments: Geological Society of America, Abstracts w/Programs, v. 12, p. 381-82.

———, 1981, Constraints on the formation of sedimentary dolomite: Science, v. 213, p. 214-16.

Bathurst, R. G. C., 1966, Boring algae, micrite envelopes and lithification of molluscan biosparites: Journal of Geology, v. 5, p. 15-32.

———, 1975, Carbonate sediments and their diagenesis, 2nd edition: Elsevier, New York, 658p.

Carr, T. R., 1982, Log-linear models, Markov chains and cyclic sedimentation: Journal of Sedimentary Petrology, v. 52, p. 905-12.

Davies, P. J., and Kinsey, D. W., 1973, Organic and inorganic factors in recent beach rock formation, Heron Island, Great Barrier Reef: Journal of Sedimentary Petrology, v. 43, p. 59-81.

Dean, W. E., Davies, G. R., and Anderson, R. Y., 1975, Sedimentological significance of nodular and laminated anhydrite: Geology, v. 3, p. 367-72.

Deffeyes, K. S., Lucia, F. J., and Weyl, P. K., 1965, Dolomitization of Recent and Plio-Pleistocene sediments by marine evaporite waters on Bonaire, Netherland Antilles: In Dolomitization and limestone diagenesis: Society of Economic Paleontologists and Mineralogists Special Publication 13, p. 71-88.

Dickson, J. A. D., 1966, Carbonate identification and genesis as revealed by staining: Journal of Sedimentary Petrology, v. 36, p. 491-505.

Driese, S. G., and Dott, R. H., Jr., 1984, Model for sandstone-carbonate "cyclothems" based upon the Upper Morgan Formation (Pennsylvanian) of northern Utah and Colorado: American Association of Petroleum Geologists Bulletin, in press.

Dunham, J. B., and Olson, E. R., 1978, Diagenetic dolomite formation related to Paleozoic paleogeography of the Cordilleran miogeocline in Nevada: Geology, v. 6, p. 556-59.

———, 1980, Shallow subsurface dolomitization of subtidally deposited carbonate sediments in the Hanson Creek Formation (Ordovician-Silurian) of central Nevada): In Zenger, D. H., Dunham, J. B., and Ethington, R. L. (eds.), Concepts and models of dolomitization: Society of Economic Paleontologists and Mineralogists Special Publication 28, p. 139-61.

Dunham, R. J., 1971, Meniscus cement: In Bricker, O. P. (ed.), Carbonate cements: Johns Hopkins University Press, Baltimore, p. 297-300.

Epstein, A. G., Epstein, J. B., and Harris, L. D., 1977, Conodont color alteration–an index to organic metamorphism: U.S. Geological Survey Professional Paper 995, p. 1-27.

Evamy, B. D., 1963, The application of a chemical staining technique to a study of dedolomitization: Sedimentology, v. 2, p. 164-70.

———, 1967, Dedolomitization and the development of rhombohedral pores in limestone: Journal of Sedimentary Petrology, v. 37, p. 1204-15.

———, 1969, The precipitational environment and correlation of some calcite cements deduced from artificial staining: Journal of Sedimentary Petrology, v. 39, p. 787-821.

Evamy, B. D., and Shearman, D. J., 1965, The development of overgrowths from echinoderm fragments: Sedimentology, v. 5, p. 211-33.

———, 1969, Early stages in development of overgrowths on echinoderm fragments in limestones: Sedimentology, v. 12, p. 317-22.

Folk, R. L., 1974, The natural history of crystalline calcium carbonate: Effect of magnesium content and salinity: Journal of Sedimentary Petrology, v. 44, p. 40-53.

Folk, R. L., and Land, L. S., 1975, Mg/Ca ratio and salinity: two controls over crystallization of dolomite: American Association of Petroleum Geologists Bulletin, v. 59, p. 60-68.

Folk, R. L., and Pittman, J. S., 1971, Length-slow chalcedony: a new testament for vanished evaporites: Journal of Sedimentary Petrology, v. 41, p. 1045-58.

Frank, J. R., 1981, Dedolomitization in the Taum Sauk Limestone (Upper Cambrian), southeast Missouri: Journal of Sedimentary Petrology, v. 51, p. 7-18.

Frank, J. R., Carpenter, A. B., and Oglesby, T. W., 1982, Cathodoluminescence and composition of calcite cement in the Taum Sauk Limestone (Upper Cambrian), southeast Missouri: Journal of Sedimentary Petrology, v. 52, p. 631-38.

Friedman, G. M., Gebelein, C. D., and Sanders, J. E., 1971, Micritic envelopes of carbonate grains are not exclusively of photosynthetic algal origin: Sedimentology, v. 16, p. 89-96.

Ginsburg, R. N., and James, N. P., 1976, Submarine botryoidal aragonite in Holocene reef limestones, Belize: Geology, v. 4, p. 431-36.

Hansen, W. R., 1977, Geologic map of the Jones Hole Quadrangle, Uintah County, Utah, and Moffat County, Colorado: U.S. Geological Survey Quadrangle Map GQ-1401.

Hanshaw, B. B., Back, W., and Deike, R. G., 1971, A geochemical hypothesis for dolomitization by groundwater: Economic Geology, v. 66, p. 710-24.
Heckel, P. H., 1982, Diagenetic model for carbonate rocks in mid-continent Pennsylvanian eustatic cyclothems: American Association of Petroleum Geologists Bulletin, v. 66, p. 580.
Hsü, K. J., and Siegenthaler, C., 1969, Preliminary experiments on hydrodynamic movement induced by evaporation and their bearing on the dolomite problem: Sedimentology, v. 12, p. 11-25.
Jacka, A. D., 1974, Replacement of fossils by length-slow chalcedony and associated dolomitization: Journal of Sedimentary Petrology, v. 44, p. 421-27.
James, N. P., Ginsburg, R. N., Marszalek, D. S., and Choquette, P. W., 1976, Facies and fabric specificity of early subsea cements in shallow Belize (British Honduras) reefs: Journal of Sedimentary Petrology, v. 46, p. 523-44.
Katz, A., 1968, Calcian dolomites and dedolomitization: Nature, v. 217, p. 439-40.
Kinsman, D. J. J., 1969, Modes of formation, sedimentary associations, and diagnostic features of shallow-water and supratidal evaporites: American Association of Petroleum Geologists Bulletin, v. 53, p. 830-40.
Knauth, L. P., 1979, A model for the origin of chert in limestone: Geology, v. 7, p. 274-77.
Kopp, O. C., 1981, Cathodoluminescence petrography: a valuable tool for teaching and research: Journal of Geological Education, v. 29, p. 108-13.
Lahann, R. W., 1978, A chemical model for calcite crystal growth and morphology control: Journal of Sedimentary Petrology, v. 48, p. 337-44.
Land, L. S., 1967, Diagenesis of skeletal carbonates: Journal of Sedimentary Petrology, v. 37, p. 914-30.
Land, L. S., and Goreau, T. F., 1970, Submarine lithification of Jamaican reefs: Journal of Sedimentary Petrology, v. 40, p. 457-61.
Lohmann, K. C., and Meyers, W. J., 1977, Microdolomite inclusions in cloudy prismatic calcites: a proposed criterion for former high-magnesium calcites: Journal of Sedimentary Petrology, v. 47, p. 1078-88.
Marshall, J. F., and Davies, P. J., 1981, Submarine lithification on windward reef slopes: Capricorn-Bunker Group, southern Great Barrier Reef: Journal of Sedimentary Petrology, v. 51, p. 953-60.
May, J. A., and Perkins, R. D., 1979, Endolithic infestation of carbonate substrates below the sediment-water interface: Journal of Sedimentary Petrology, v. 49, p. 357-78.
Meyers, W. J., 1974, Carbonate cement stratigraphy of the Lake Valley Formation (Mississippian) Sacramento Mountains, New Mexico: Journal of Sedimentary Petrology, v. 44, p. 837-61.
__________, 1977, Chertification in the Mississippian Lake Valley Formation, Sacramento Mountains, New Mexico: Sedimentology, v. 24, p. 75-105.
__________, 1978, Carbonate cements: their regional distribution and interpretation in Mississippian limestones of southwestern New Mexico: Sedimentology, v. 25, p. 371-400.
Meyers, W. J., and James, A. T., 1978, Stable isotopes of cherts and carbonate cements in the Lake Valley Formation (Mississippian), Sacramento Mountains, New Mexico: Sedimentology, v. 25, p. 105-24.
Meyers, W. J., and Lohmann, K. C., 1978, Microdolomite-rich syntaxial cements: proposed meteoric-marine mixing zone phreatic cements from Mississippian limestones, New Mexico: Journal of Sedimentary Petrology, v. 48, p. 475-88.
Moore, C. H., Jr., 1973, Intertidal carbonate cementation, Grand Cayman, West Indies: Journal of Sedimentary Petrology, v. 43, p. 591-602.
Muller, G., 1971, 'Gravitational' cement: an indicator for the vadose zone of the subaerial diagenetic environment: In Bricker, O. P. (ed.), Carbonate cements: Johns Hopkins University Press, Baltimore, p. 301-2.
Murray, R. C., 1964, Origin and diagenesis of gypsum and anhydrite: Journal of Sedimentary Petrology, v. 34, p. 512-23.
Neal, W. J., 1969, Diagenesis and dolomitization of a limestone (Pennsylvanian of Missouri) as revealed by staining: Journal of Sedimentary Petrology, v. 39, p. 1040-45.
Picard, M. D., 1977, Petrography and stratigraphy of productive beds in the Morgan Formation, Church Buttes Unit No. 19, southwest Wyoming: In Rocky Mountain Thrust Belt geology and resources: Wyoming Geological Association, 29th Annual Field Conference, p. 179-96.
Pierson, B. J., 1981, The control of cathodoluminescence in dolomite by iron and manganese: Sedimentology, v. 28, p. 601-10.
Radke, B. M., and Mathis, R. L., 1980, On the formation and occurrence of saddle dolomite: Journal of Sedimentary Petrology, v. 50, p. 1149-68.
Schmitt, J. G., and Boyd, D. W., 1981, Patterns of silicification in Permian pelecypods and brachiopods from Wyoming: Journal of Sedimentary Petrology, v. 51, p. 1297-1308.
Shearman, D. J., 1963, Recent anhydrite, gypsum, dolomite, and halite from the coastal flats of the Arabian shore of the Persian Gulf: Geological Society of London Proceedings, v. 1607, p. 63-65.
Shinn, E. A., 1968, Selective dolomitization of recent sedimentary structures: Journal of Sedimentary Petrology, v. 38, p. 612-16.
__________, 1969, Submarine lithification of Holocene carbonate sediments in the Persian Gulf: Sedimentology, v. 12, p. 109-44.
__________, 1971, Aspects of diagenesis of algal cup reefs in Bermuda: Gulf Coast Association of Geological Societies Transactions, v. 21, p. 387-94.
Shinn, E. A., Halley, R. B., Hudson, J. H., and Lidz, B., 1977, Limestone compaction: an enigma: Geology, v. 5, p. 21-24.
Sippel, R. F., and Glover, E. D., 1965, Structures in carbonate rocks made visible by luminescence petrography: Science, v. 150, p. 1283-87.
Swinchatt, J. P., 1969, Algal boring: a possible depth indicator in carbonate rocks and sediments: Geological Society of America Bulletin, v. 80, p. 1391-96.
Verville, G. J., Sanderson, G. A., Brenckle, P. L., and Lane, H. R., 1973, Upper Paleozoic biozonation in the Unit No. 19 well, Church Buttes field, Uinta County, Wyoming: In Symposium and core seminar on the geology and mineral resources of the greater Green River Basin: Wyoming Geological Association Guidebook, 25th Field Conference, p. 165-71.
Wagner, P. D., and Matthews, R. K., 1982, Porosity preservation in the Upper Smackover (Jurassic) carbonate grainstone, Walker Creek field, Arkansas: response of paleophreatic lenses to burial processes: Journal of Sedimentary Petrology, v. 52, p. 3-18.
Wanless, H. R., 1979, Limestone response to stress: pressure solution and dolomitization: Journal of Sedimentary Petrology, v. 49, p. 437-62.