

# Overview of Porosity Evolution in Carbonate Reservoirs\*

By  
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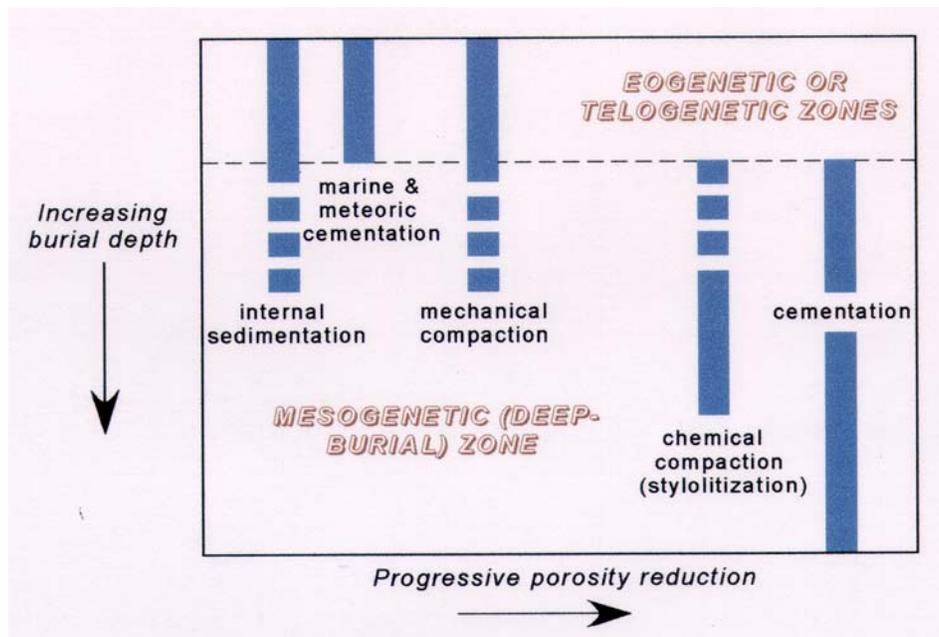
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## Introduction

Carbonate rocks (limestone and dolomite) account for approximately 50% of oil and gas production around the world. Of the carbonates, a slightly greater percentage of world hydrocarbon reserves has been produced from dolomites because such rocks commonly, but not always, have more porosity and permeability than limestones (Halley and Schmoker, 1983). Unlike most sandstone reservoirs, which typically are single-porosity systems (i.e., interparticle pores) of relative uniform (homogeneous) nature, reservoirs in carbonate rocks commonly are multiple-porosity systems that characteristically impart petrophysical heterogeneity to the reservoirs (Mazzullo and Chilingarian, 1992). Hence, the specific types and relative percentages of pores present, and their distribution within the rocks, exert strong control on the production and stimulation characteristics of carbonate reservoirs (e.g., Jodry, 1992; Chilingarian et al., 1992; Honarpour et al., 1992; Hendrickson et al., 1992; Wardlaw, 1996). Pore types in carbonate rocks can generally be classified on the basis of the *timing* of porosity evolution (Choquette and Pray, 1970) into: (1) primary pores (or depositional porosity), which are pores inherent in newly-deposited sediments *and* the particles that comprise them. Such pore types include *interparticle* pores in, for example, carbonate sands (but also in muddy carbonates), *intraparticle* pores (within particles such as foraminifera or gastropod shells), *fenestral* pores (formed by gas bubbles and sediment shrinkage in tidal-flat carbonates), and *shelter* and *growth-framework* pores (common in reef buildups); and (2) secondary pores, which are those that form as a result of later, generally post-depositional dissolution. Such pore types include all of those mentioned above, but only when it can be demonstrated that primary pores which subsequently were occluded by cement later had all or some of that cement dissolved (resulting in the generation of exhumed pores - Figure 2), as well as *vugs* (large pores that transect rock fabric, that is, dissolution was not fabric-selective) and *dissolution-enlarged* fractures. Most of these primary and secondary pore types can readily be identified in cores, and with the possible exception of shelter and growth-framework pores, also in well cuttings samples.

Because the natural tendency in most carbonate sediments is that primary porosity is substantially reduced by cementation and compaction during post-depositional burial (Figure 1; Halley and Schmoker, 1983), many workers would argue that most porosity in limestone and dolomite reservoirs is of secondary origin (e.g., Mazzullo and Chilingarian, 1992). Exceptions to this statement are cases where primary porosity is preserved because of the early influx of hydrocarbons into pores (e.g., Feazel and Schatzinger, 1985). Early on in the study of carbonate sediments and their diagenesis, the *subaerial meteoric diagenetic (freshwater) model* was promoted as a means of explaining porosity evolution in carbonates, specifically in shallow-water carbonates that lie beneath unconformities in paleo-vadose and paleo-phreatic freshwater zones (e.g., Friedman, 1964; Land, 1967). This model still is heavily applied today, especially in sequence stratigraphic-related diagenetic studies of reservoirs. This model presupposes that if porous carbonate rocks are present beneath unconformities, then that porosity must have been created by freshwater dissolution during subaerial exposure. Of course, as explorationists, we all can probably list a large number of wells that were drilled into non-porous carbonate rocks beneath unconformities. Hence, the corollary pertaining to subaerial exposure is not true - meteoric exposure does not always create porosity, and even if it did, that porosity may be occluded during later burial (Figure 1). A most critical constraint on evaluating, or more importantly, on predicting porosity in carbonate rocks utilizing only the subaerial meteoric diagenetic model is that one must call upon fluids capable of dissolving carbonate to come from *above*; that is, from rain water percolating down into sediments or rocks beneath unconformity surfaces. Ostensibly, then, many might consider that if there is not an unconformity in the section, then the carbonates will not be porous. Again, as explorationists we can all probably compile a list of wells in which porous carbonates that were not associated with unconformities were encountered in the subsurface.



**Figure 1. Processes by which porosity is reduced in carbonate rocks. Syndepositional marine cementation occurs only in the eogenetic zone, and mechanical compaction is unlikely to affect telogenetically-exposed older carbonate rocks.**

The foregoing analysis therefore begs the following questions: (1) can secondary porosity in carbonate rocks be generated by processes *other than* subaerial meteoric exposure, and if so, what are those processes?; (2) how might reservoir porosity formed by such alternative processes differ from reservoirs created by meteoric dissolution along unconformities?; (3) how can we recognize and determine the origin of reservoir porosity?; and (4) can the subsurface occurrence of porosity formed by any such alternative models of reservoir origin be predicted? The purpose of this contribution is to address these questions by demonstrating the multi-faceted evolution of secondary porosity in carbonate hydrocarbon reservoirs. In the following discussions attention will be focused on the recognition and origin of pore types in shallow-water limestones and dolomites, as observed mainly in cores and well cuttings samples.

### **Secondary Porosity Beneath Unconformities: The Subaerial Meteoric Model**

The generation of secondary porosity in carbonate sediments or rocks in this model is a direct consequence of dissolution by freshwater (ultimately rain-derived), which dissolves carbonates because the water is undersaturated with respect to calcium carbonate. The extent of dissolution and secondary porosity formation are controlled by factors such as the acidity of freshwater (e.g., rain water percolating down through a soil zone will be more acidic than in areas where soils are not present), the amount of porosity or fractures within the affected carbonates, the residence time of the freshwater in the diagenetic system, the mineralogy of the sediments or rocks, and so forth (Longman, 1980; Moore, 1989). Secondary porosity generation via dissolution can occur relatively soon after deposition, in unconsolidated sediments, in what Choquette and Pray (1970) refer to as the *eogenetic* zone; or it can occur much later, in rocks, in the *telogenetic* zone as a consequence of uplift of older, formerly buried carbonates (Figure 2). In newly-deposited carbonate sediments that subsequently are subaerially exposed, it is the difference in original mineralogies of particles in the sediments that drives the relatively rapid, selective dissolution of particles. Fragments of corals, pelecypods and gastropods, and oolites, for example, are originally composed mostly of the mineral *aragonite* ( $\text{CaCO}_3$ , orthorhombic), which is very soluble. It is for this reason that formerly aragonitic particles in limestones usually are represented by pores (in this case, fabric-selective pores) or cement-filled pores. In contrast, particles such as forams, crinoid fragments, and bryozoans are originally composed of the mineral high-magnesium calcite ( $\text{CaCO}_3$ , hexagonal-rhombohedral), which is calcite with up to 23 mole%  $\text{MgCO}_3$  in the crystal lattice. With exposure to freshwater such particles tend merely to lose  $\text{MgCO}_3$  and not to dissolve like aragonitic particles. Other particles, such as brachiopod shells and some pelecypods, build their skeletons out of low-Mg calcite (also  $\text{CaCO}_3$ , hexagonal-rhombohedral), which is calcite with less than 4 mole%  $\text{MgCO}_3$ . Particles of original high-magnesium calcite and low-magnesium calcite mineralogy tend not to dissolve unless the freshwater is quite undersaturated with respect to calcium carbonate, and it is for this reason that some particles (crinoids, bryozoans, brachiopods) often are well-preserved in ancient rocks. The eogenetic exposure to freshwater of newly-deposited carbonate sediments, which are generally highly porous and *polyminerallic* (i.e., as

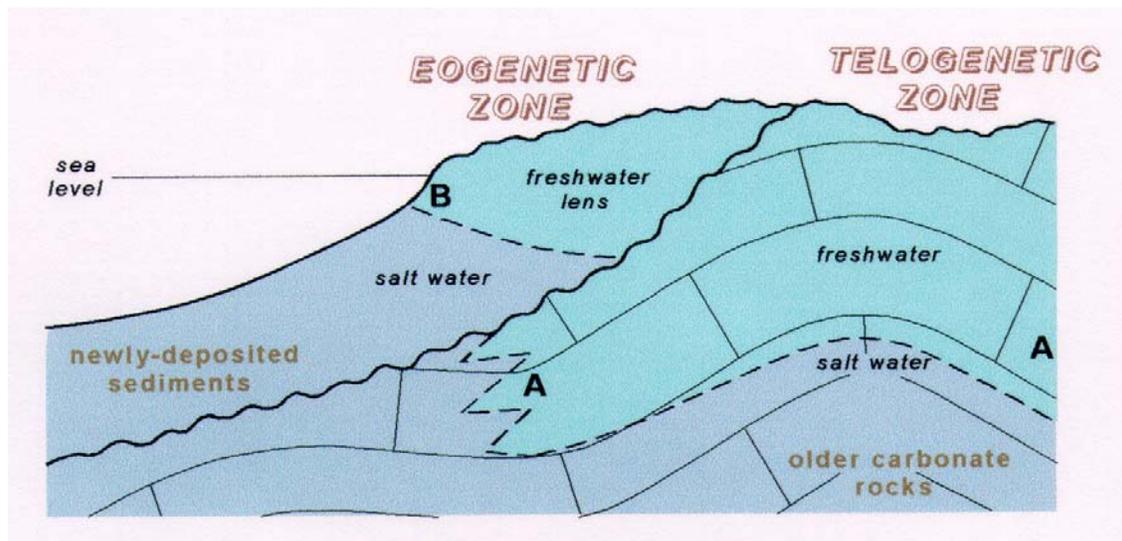
discussed above, composed of mixtures of aragonite, high-magnesium and low-magnesium calcite), results in the formation of cemented limestones, of varying porosity, of stable low-magnesium calcite composition (notwithstanding dolomitization). In contrast, late eogenetic or telogenetic freshwater exposure of older limestones that have already been mineralogically stabilized and cemented is not driven by such differences in the relative solubility of aragonite, high-magnesium and low-magnesium calcite because the rocks already are mineralogically stabilized to low-magnesium calcite, and further dissolution can occur only if the fluids are quite undersaturated with respect to calcite (the least soluble of the aforementioned carbonate minerals). Typically, such dissolution forms vugs and caverns, which can also form in polymineralic carbonate sediments. Dissolution of already stabilized limestones can also result in the formation of particle-selective pores when certain particles in the rocks are slightly more soluble than other particles because of differences in particle size or their micro-architectural arrangement of component calcite crystals. For example, crinoid fragments in older rocks are composed of single, relatively large crystals of low-magnesium calcite, which have relatively low solubility. It is for this reason that crinoid-rich Mississippian limestones, for example, typically have low porosities. In contrast, foram shells are composed of myriads of small calcite crystals, which have relatively higher solubility, and they usually are more readily dissolved than crinoid fragments.

In either case, it is important to note that carbonates can be affected by meteoric dissolution not only directly beneath unconformities on land, but also for some distance down-dip into the subsurface ("A" in Figure 2) and some distance in a seaward direction below sea level, depending on the extent of freshwater lenses ("B" in Figure 2). Porosity generation by dissolution eventually ceases, generally in a down-dip direction within phreatic zones when that water becomes saturated with respect to dissolved calcium carbonate. At that point, porosity can be maintained, or if the water becomes even more saturated with respect to dissolved calcium carbonate, it can begin to be occluded by carbonate cement (and other cements as well, such as gypsum/anhydrite or silica).

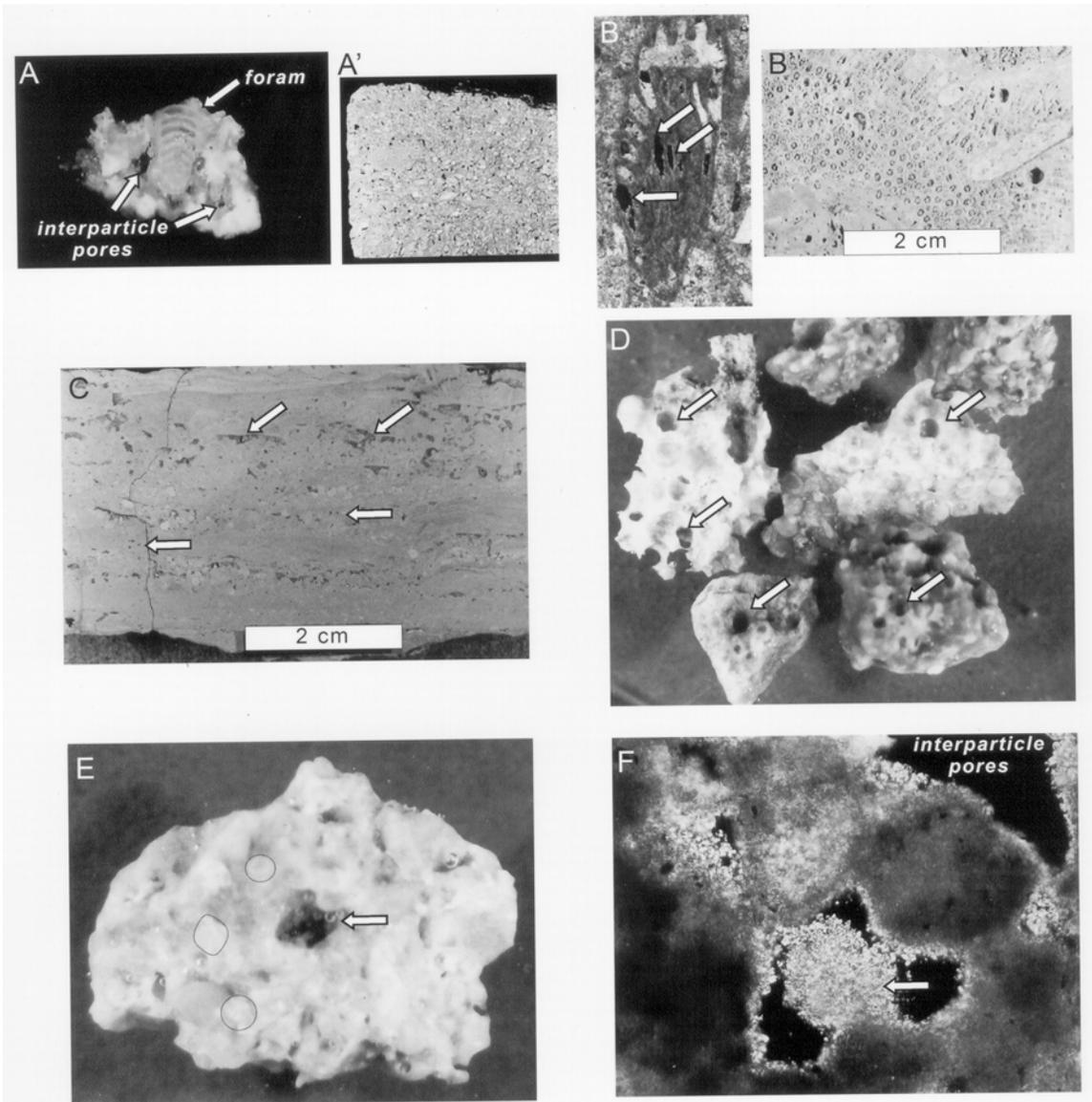
### **Meteoric Porosity in Limestones**

In limestones, common secondary pore types formed as a result of post-depositional dissolution variously include exhumed interparticle, intraparticle, fenestral, shelter, and growth-framework pores, all of which are considered to be fabric-selective pores; and also not fabric-selective vugs (Figure 3E) and dissolution-enlarged fractures. The size of vugs (Figure 4) varies from small (but larger than component particles in the rocks) to caverns or cavernous porosity. Vugs may originate either by wholesale dissolution of parts of the rock or by dissolutional enlargement of fabric-selective pores (Figure 3E). In many cases there is coincidence between the types of fabric-selective pores present in the rocks and the depositional environment of the rocks, which serves as an important guide in evaluating permeability and potential recoverable reserves from the reservoir, and in deciding on what stimulation procedures to use but only if one knows the depositional environment of the rocks from study of subsurface samples. For example, carbonate sands (lime grainstones), deposited in high-energy environments such as oolite shoals or skeletal sand shoals, commonly have high interparticle porosity (Figure 3A) and

attending relatively high permeabilities. On the other hand, however, high porosity but low permeability may characterize a carbonate sand (limestone) reservoir wherein only the particles have been dissolved (for example, in cases where the reservoir contains only molds of oolites referred to as oomoldic porosity or by the older term oocastic porosity: Figure 3D). In such cases there may be ample hydrocarbon storage volume in the pores in the rocks, but in the absence of fractures, there is little interconnected porosity. Notwithstanding porosity associated with dolomitization (discussed later), limestones deposited in tidal-flat environments commonly contain a specific type of vuggy porosity referred to as fenestral pores (for example, birdseye pores, which is one type of pinpoint porosity: Figure 3C), and unless fractured, such rocks may have decent porosity but limited permeability. Skeletal sands shoals wherein the particles mainly are foraminifera, which are common in Midcontinent Pennsylvanian limestone reservoirs (Wilhite and Mazzullo, 2000), may be highly porous and permeable because of the presence of interparticle pores, and within the forams, of intraparticle pores as well (Figure 3B). On the other hand, if intraparticle pores are the only pore types present, then porosity (and hydrocarbon storage volume) might be high, but permeability would be low (notwithstanding fracturing). As a corollary, variations in porosity and permeability from well-to-well within a given zone may be a consequence of different depositional environments in that zone and/or from differing extents of porosity generation versus occlusion between wells. Only study of cores/cuttings and thin-sections can resolve the possible reasons for such variations between wells.



**Figure 2.** Meteoric subaerial exposure of newly-deposited sediments in the eogenetic zone and of older carbonate rocks in the telogenetic zone. Note that the freshwater lens in the eogenetic zone can extend some distance below sea level ("B"), and that freshwater may extend down -dip for a considerable distance in the telogenetic zone ("A").

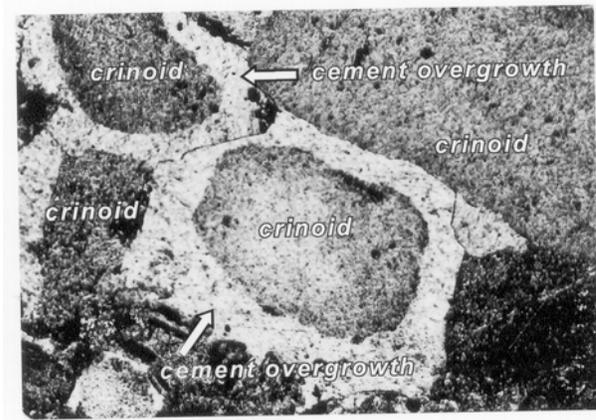


**Figure 3. Typical secondary dissolution pore types in carbonate rocks that are readily identifiable in cuttings and core samples. A - interparticle pores in a grainstone (cuttings sample) and A' core showing interparticle porosity in a carbonate grainstone. B - thin-section photomicrograph of intraparticle porosity (arrows) within a fusulinid and B' core showing intraparticle porosity within a coral. C - Fenestral porosity in a tidal-flat dolomite. Tilted arrows point to planar (laminar) fenestral pores, and horizontal arrows point to smaller "birdseye" pores. D - Cuttings samples with oomoldic pores (arrows) in an oolite grainstone. E - Carbonate grainstone with identifiable skeletal particles (circled) and larger vug (arrow) that formed from the initial dissolution of a particle and then further dissolution of the matrix around it (cuttings sample). F - Thin-section photomicrograph of interparticle porosity in a carbonate sand wherein remnant cement (arrow) restricts pore throats and reduces permeability.**



**Figure 4. Core slab with secondary vugs (arrows) resulting from the partial dissolution of earlier dolomite cement (white).**

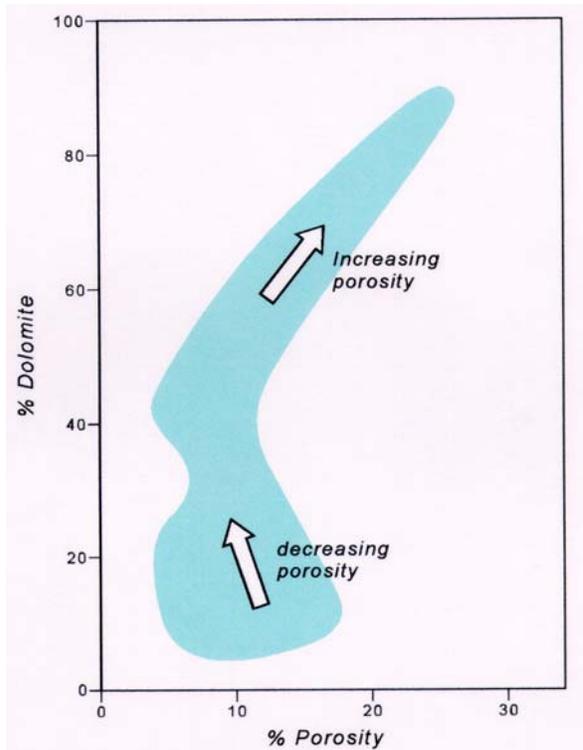
Over-riding such generalizations about the relationships among pore types, permeability, and depositional environments of the limestones is the importance of pore throats in the rocks (Wardlaw, 1976). In limestones, particularly in grainstones, for example, the nature of pore throats and their effect on permeability is controlled by the size of the particles in the rocks, and more importantly, by the distribution of any remaining earlier-precipitated cement in the pores that was not dissolved (Figure 3F). Calcite cement overgrowths on crinoid fragments can significantly restrict pore throats as well (Figure 5), which is why many crinoid-rich Mississippian limestones are of low-permeability nature. The best way to determine the extent of pore-throat restriction in the rocks under consideration is by examining the rocks petrographically in thin-section. Clay-mineral cements are extremely rare in carbonate reservoir rocks, and therefore, need not be considered here.



**Figure 5. Thin-section photomicrograph of calcite cement overgrowths on crinoid fragments that occlude interparticle porosity.**

### **Meteoric Porosity in Dolomites**

In contrast to earlier postulates on the subject (e.g., Murray, 1960; Weyl, 1960), the process of dolomitization of a pre-existing limestone does *not* automatically create secondary porosity. Whereas it is true that porosity tends to increase as amount of dolomite increases (Figure 6), it generally does so for the following reasons. In partly dolomitized limestones exposed to telogenetic meteoric fluids, for example, any remaining calcite (which may represent particles and/or carbonate mud matrix) inherently is more susceptible to dissolution by freshwater because it is more soluble than dolomite. Hence, subaerial exposure of a partly dolomitized limestone can result in the generation of the same types of pores as described above by dissolution of remaining calcite, depending on the original texture of the rock (mudstone, wackestone, packstone, or grainstone), its depositional environment, and degree of replacement by dolomite (Figure 7). Likewise, remaining evaporites in the rocks can also be dissolved. In more pervasively dolomitized rock exposed to telogenetic meteoric fluids, remaining calcite (or evaporite minerals) between dolomite crystals can be dissolved during subaerial exposure to produce intercrystalline pores between dolomite crystals. In completely dolomitized rocks, vugs (and sometimes dissolution-enlarged fractures) are common pore types present if the meteoric fluids were highly acidic or acted on the rocks over long periods of time. Selective dissolution of small dolomite crystals (because solubility increases as crystal size decreases), or of more soluble dolomite phases in the rocks, can result in the development of vugs and intercrystalline pores. All such processes and resulting pore types can be represented in a given reservoir. As in limestones, the nature of pore throats in dolomites affects permeability, and as a general rule, intercrystalline pore throat sizes decrease with decreasing crystal size and extent of dolomitization (Figure 8).



**Figure 6. Relationship of porosity to percent dolomite in carbonate rocks (after Murray, 1960).**

### **Cavernous Porosity in Carbonate Rocks**

Cavernous and associated vuggy porosity are major attributes of hydrocarbon production from reservoirs such as the Arbuckle Group in Kansas (Walters, 1946; Newell et al., 1987) and Oklahoma (Gatewood, 1970), and from its stratigraphic correlative, the Ellenburger Group, in west Texas and New Mexico (Holtz and Kerans, 1992). Additional examples of hydrocarbon reservoirs in paleo-caverns are given in Mazzullo and Chilingarian (1996). Only rarely are completely fluid-filled caverns encountered in the subsurface. Rather, paleo-caverns usually are filled by porous (or, unfortunately, sometimes tight) cave-roof collapse breccia and associated sediments and/or by overlying, younger rocks (Figure 9). Rather than being single zones, paleo-caverns typically are labyrinthine systems characterized by extreme lateral and vertical reservoir compartmentalization (Figure 9). Cavernous porosity undoubtedly also locally contributes to hydrocarbon production from some Mississippian reservoirs in Kansas. I have encountered a number of instances in Kansas, for example, where wellsite geologists' reports picked the top of the Mississippian at a certain depth, and then the limestone or dolomite directly below seemingly was underlain by a section of sand (which I presume to be Pennsylvanian-age siliciclastic sand) that is, in turn, underlain by more carbonate rock. Such occurrences may indicate that the wells penetrated sand-filled paleo-caverns (Figure 10).

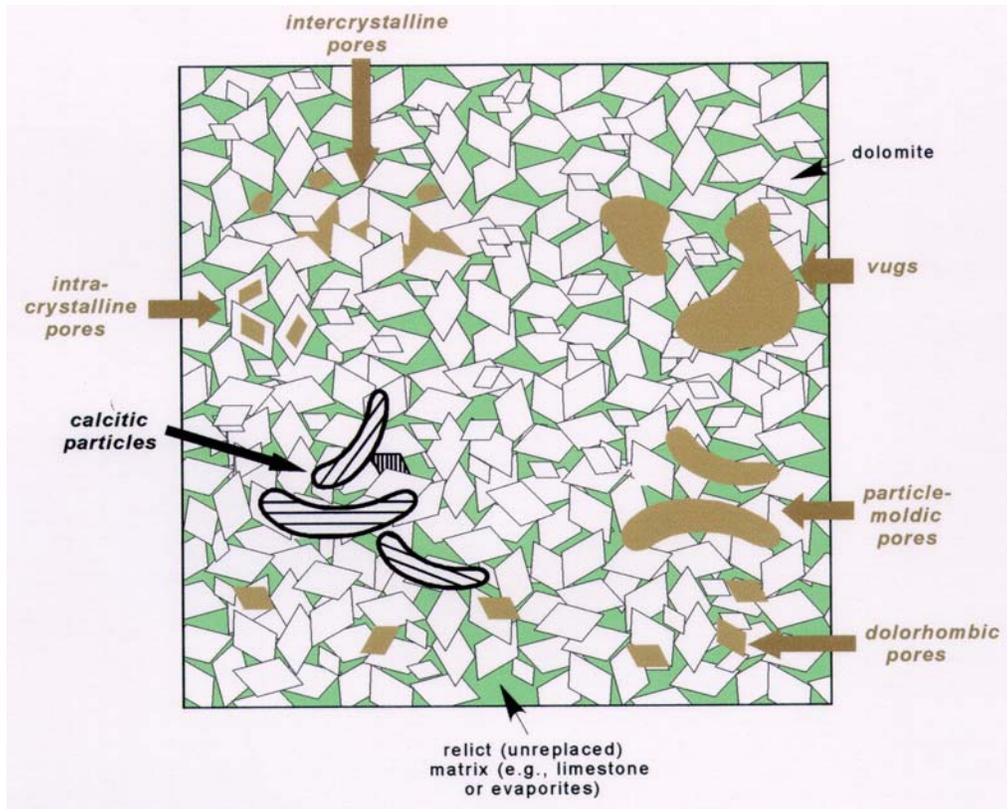


Figure 7. Origin of secondary dissolution porosity in dolomites.

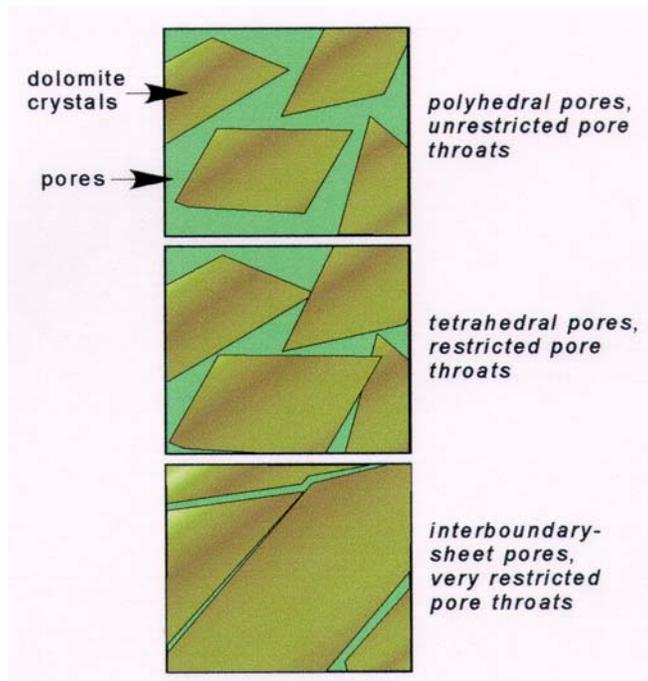
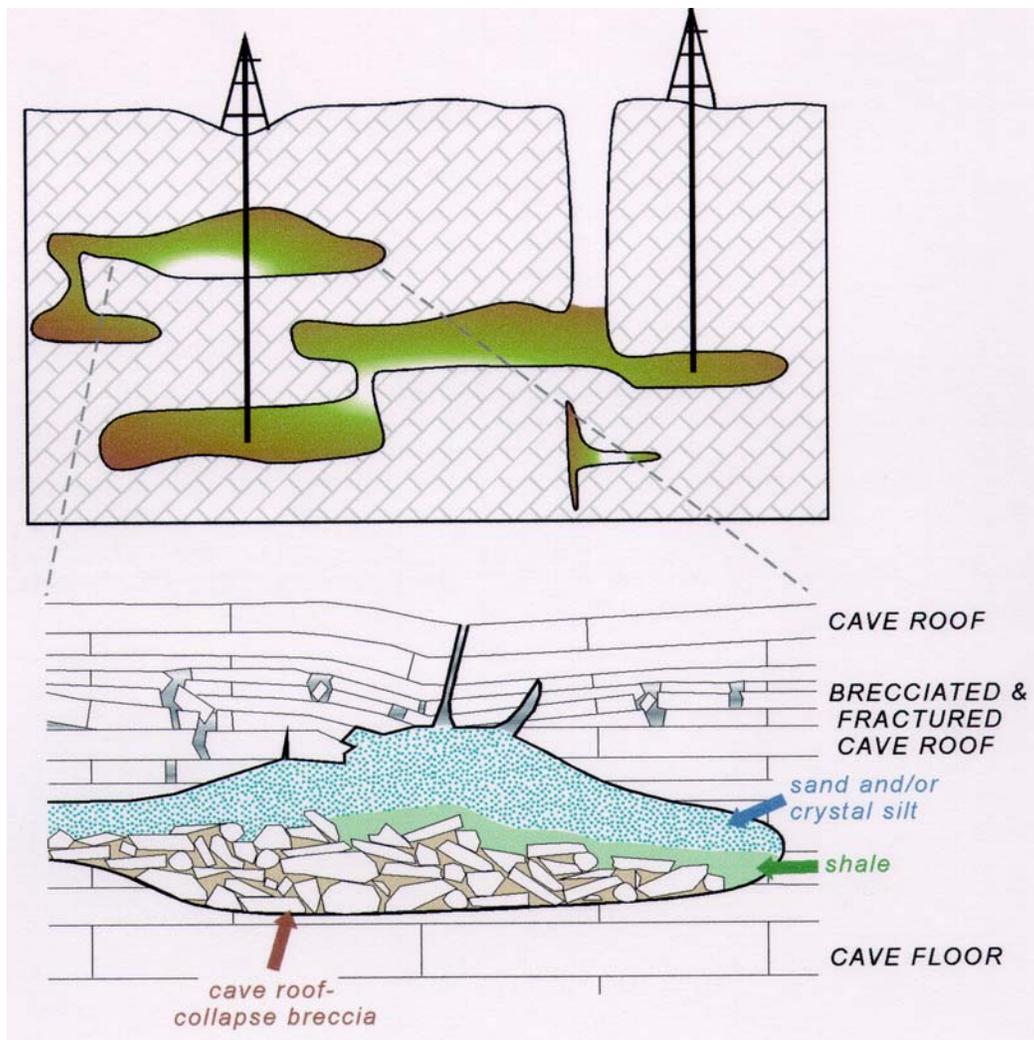


Figure 8. Intercrystalline pores and pore throats in dolomites. Relative size of pores and pore throats is not necessarily correlative to dolomite crystal size because porosity is a percentage of total rock volume. Pore throat characteristics, however, do reflect the degree (extent) of dolomitization in rocks. Dolomites with polyhedral pores generally are referred to as "sucrosic."

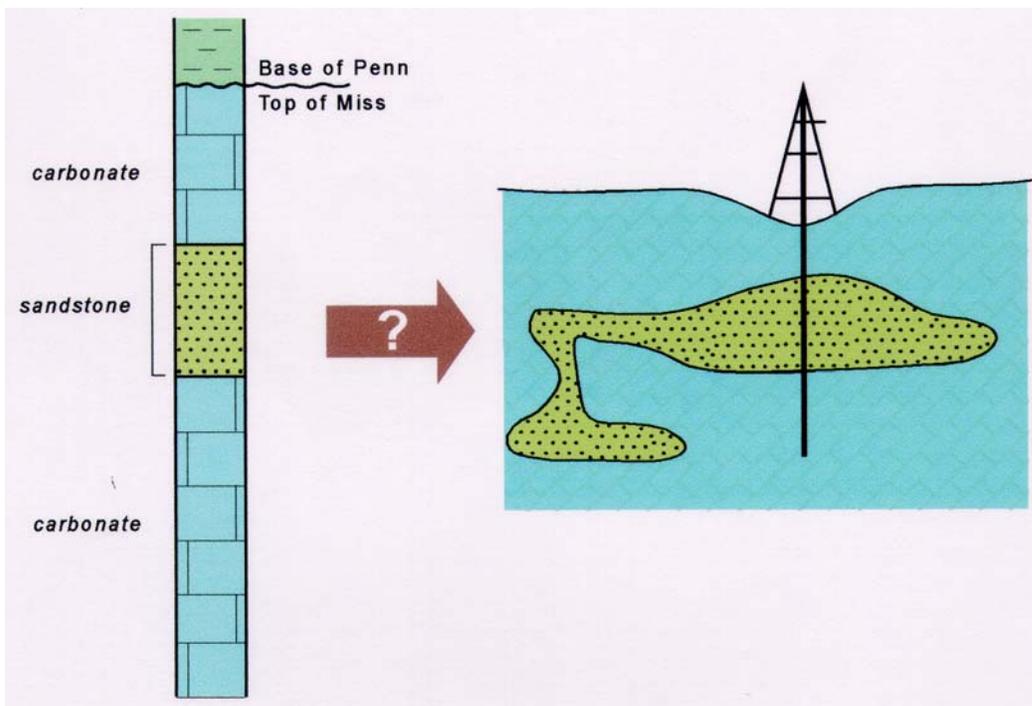


**Figure 9. Top - Typical lateral and vertical compartmentalization of reservoir zones in cavernous (karsted) carbonate rocks. Bottom - Typical porosity types and fills of cavernous reservoirs. Cave roof rocks become progressively more brecciated downward, with attending fracture, dissolution-enlarged fracture, and commonly, vuggy porosity. Cave-fill deposits variously can be: (1) cave roof-collapse breccia, which can have inter-clast porosity as well as intra-clast vuggy and fracture porosity. Conversely, original inter-clast porosity can be filled with cements and/or shale, or can be filled with porous sand; (2) impermeable shale infiltrated into the cavern from above; or (3) porous or tight sand infiltrated into the cavern from above. Cave roof collapse and infiltrating of sand and/or shale can occur soon after karstification or later.**

### **Alternative Origin Of Secondary Porosity: The Mesogenetic Model**

Since the late 1970s-early 1980s, geologists began to suspect that not all secondary dissolution porosity in carbonate rocks forms or formed solely beneath unconformities by freshwater dissolution in either the eogenetic or telogenetic environments (e.g., Bathurst, 1980; Scholle and Halley, 1985; Choquette and James, 1987; Moore, 1989). Rather, there was growing realization of the significance of, and processes controlling, secondary dissolution porosity formation (and porosity occlusion) in the deep-burial environment **B** which is what Choquette and Pray (1970) referred to as the *mesogenetic environment*.

Two important points in this regard are the facts that: (1) not all porous carbonates are associated with unconformities; and (2) specifically, there are a number of examples of porous and permeable carbonate rocks deposited in deep-water settings and which later were deeply buried and never subaerially exposed. Hence, meteoric exposure at any time after deposition has been ruled out for such rocks (e.g., Mazzullo and Harris, 1991; Mazzullo, 1994). Therefore, post-depositional diagenesis and the formation of secondary dissolution porosity in such rocks must have occurred in an environment *other than* the meteoric eogenetic or telogenetic environment. Furthermore, if we are exploring for hydrocarbon reservoirs in the subsurface, then those reservoirs must have resided in the subsurface, variously shallowly or deeply buried, for long periods of time. Insofar as carbonate diagenesis never ceases, any diagenesis that occurs in the mesogenetic environment overprints earlier diagenesis, including that which may have occurred, for example, in subunconformity, meteoric eogenetic or telogenetic environments. Geologists have since come to realize that deep-burial diagenesis has significantly contributed to secondary dissolution porosity and permeability evolution in many carbonate hydrocarbon reservoirs (e.g., Mazzullo and Harris, 1992).



**Figure 10.** On the left is a partial stratigraphic column of what I have sometimes encountered in the Mississippian in Kansas a few feet of carbonate below the top of the Mississippian, underlain by sandstones (which are not laterally correlative for any distance), in turn underlain by more carbonate. On the right is a possible interpretation of such a stratigraphy that the wells in question encountered sand-filled caverns below the top of the Mississippian.

As discussed above, in order for dissolution of any carbonate rocks to proceed, they must be exposed to fluids that are undersaturated with respect to calcium carbonate. That is easy enough to do in the subunconformity meteoric environment because rain water, the ultimate source of near-surface freshwater, is undersaturated. In the mesogenetic environment, however, most connate fluids are brines that typically are saturated or even supersaturated with respect to calcium carbonate, which means they are not capable of dissolving carbonate rocks and creating secondary porosity. Rather, such fluids tend to precipitate carbonates in the form of calcite or dolomite cement, and in some cases, they may be capable of dolomitization. How, then, can carbonate dissolution and porosity formation proceed in the deep-burial environment? In other words, how are fluids undersaturated with respect to calcium carbonate generated in the mesogenetic environment?

Studies of porosity evolution in sandstones, combined with studies of organic matter maturation and hydrocarbon generation in source rocks (e.g., Foscolos, 1984; Surdam et al., 1984; Kharaka et al., 1986; Lundegard and Land, 1986; Meshri, 1986; Sassen and Moore, 1988), have provided answers to these questions which have been applied to carbonate reservoir rocks, both limestone and dolomite, around the world (e.g., Druckman and Moore, 1985; Heydari and Moore, 1989; Mazzullo and Harris, 1992). It is known that carbon dioxide, hydrogen sulfide, and great quantities of organic acids are generated during the maturation of organic matter to hydrocarbons in buried source rocks (Figure 11). As these gases and organic acids are expelled from the source rock, the evolved CO<sub>2</sub> combines with subsurface water to produce carbonic acid and the H<sub>2</sub>S similarly combines with water to produce sulfuric acid. Together, these acids and associated organic acids can migrate great distances laterally as well as vertically (Figure 12; e.g., Hanor, 1987) to dissolve buried carbonates just ahead of migrating hydrocarbons. Likewise, once the acids are spent, subsurface fluids can then precipitate carbonate cements, which is why many examples of such cements contain hydrocarbon inclusions (e.g., Burruss et al., 1985). In given rocks, secondary dissolution porosity formation can alternate with cementation many times to result in complex diagenetic histories of reservoirs (e.g., Moore and Druckman, 1981; Mazzullo and Harris, 1989; Moore, 1989). Because such evolved subsurface fluids can migrate great distances both laterally and vertically, they can affect carbonates that were or are not deeply buried. I have come across several wells in Ness County, Kansas, for example, where there is a great concentration of pyrite and minor sphalerite at the top of the Mississippian, which is at only about 4100 feet below the surface, clearly not in the deep-burial environment (at least, not today). These minerals may have been emplaced along the pre-Pennsylvanian unconformity by fluids that evolved within and subsequently migrated out of the Anadarko Basin.

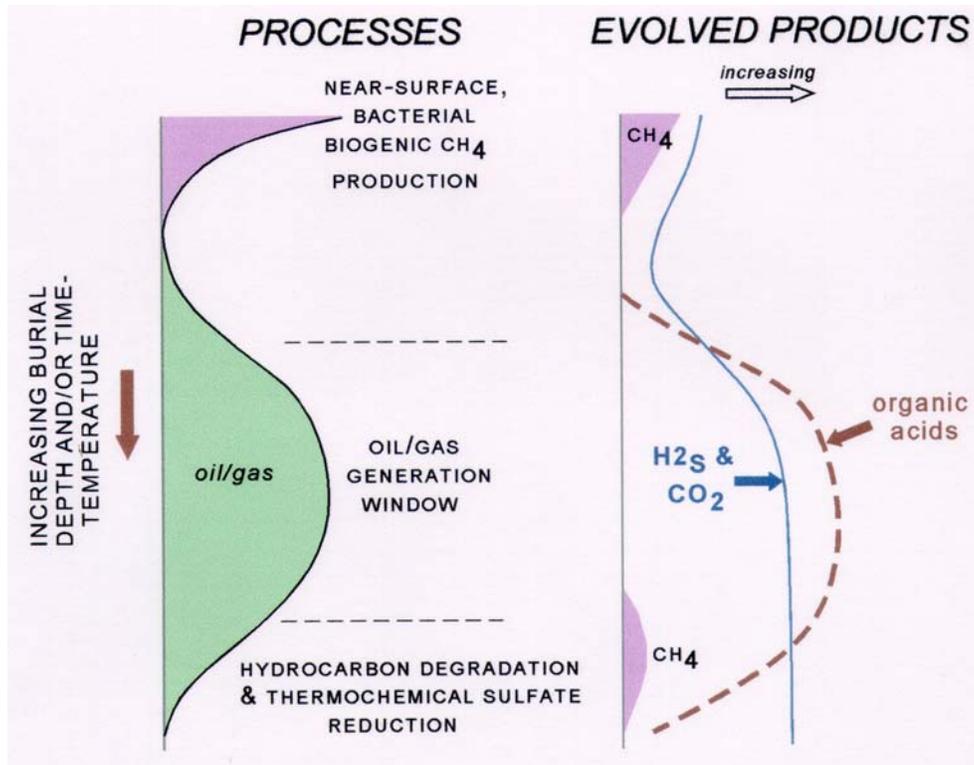


Figure 11. Processes and products of hydrocarbon generation. Relatively minor amounts of methane ( $\text{CH}_4$ ) are generated during shallow burial by bacterial processes. As organic matter in source rocks matures and generates oil or gas, copious amounts of hydrogen sulfide ( $\text{H}_2\text{S}$ ) and carbonate dioxide ( $\text{CO}_2$ ) are generated, along with organic acids.

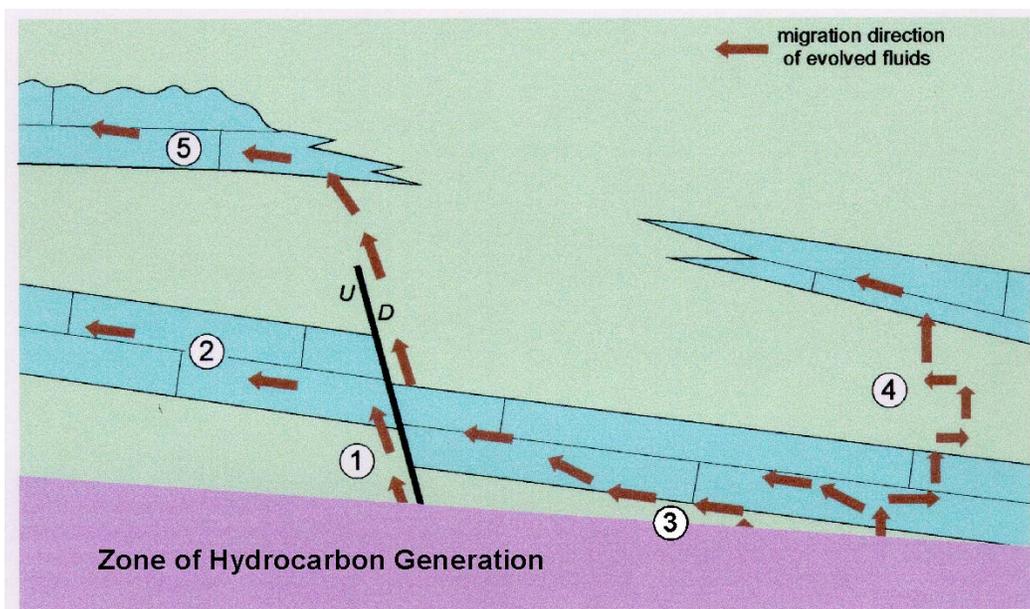
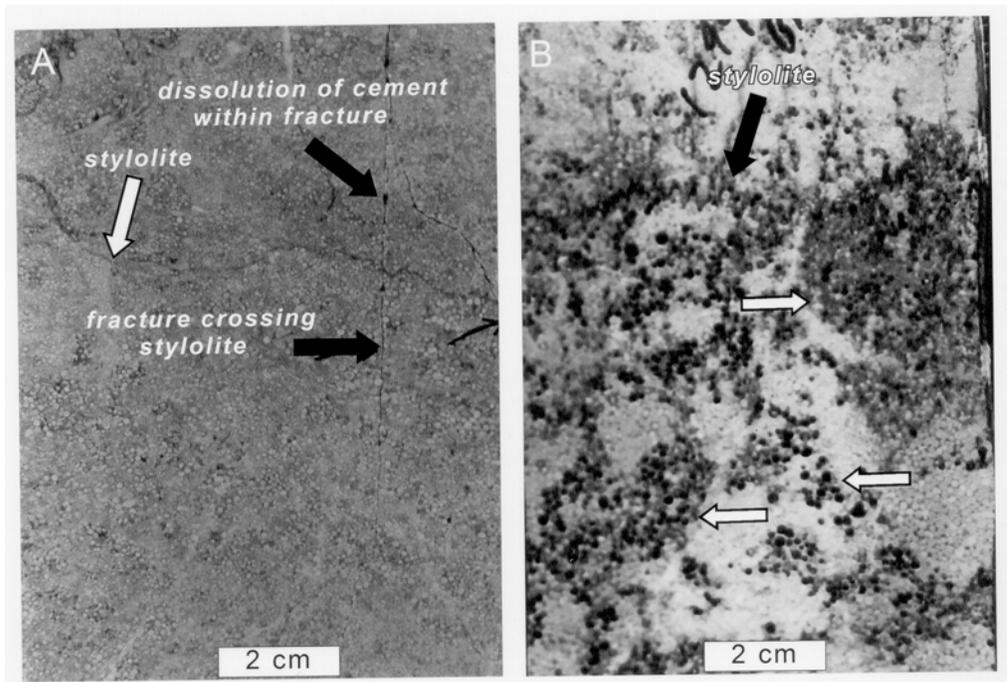


Figure 12. Carbon dioxide, hydrogen sulfide, and organic acids evolved during hydrocarbon generation mix with connate waters in the burial environment to produce acids. These acidic fluids can migrate both vertically and laterally for great distances, even affecting carbonate rocks that are not deeply buried (5). Acidic fluids can migrate along faults (1), along bedding planes or porous strata within beds (2), along formational boundaries (3), or upward along fractures through otherwise non-permeable beds ("cross-formational flow" (4).

Porosity formed in the mesogenetic environment is represented by the same types of pores that can form in the eogenetic and telogenetic freshwater environment (Mazzullo and Harris, 1992), including even cavernous porosity (e.g., Hall, 1990; Hill, 1992), which otherwise is generally known as **A** burial karst. Therefore, an important point to remember in this regard is that the diagenetic environment in which porosity formation occurred cannot be determined on the basis of the pore types present in a reservoir! How, then, does one recognize porosity formed in the mesogenetic environment? The answer to this question is "*With careful, detailed petrographic study of thin-sections, often combined with analysis of carbon and oxygen isotopic values of carbonate cements associated with suspected mesogenetic porosity.*" For details, refer to the papers by Moore and Druckman (1981), Druckman and Moore (1985), Heydari and Moore (1989), Moore (1989) and Mazzullo and Harris (1991, 1992), and the many papers cited therein. There are, however, some readily visible clues to the mesogenetic origin of porosity in carbonate rocks that can be identified in cuttings and core samples. For example: (1) porosity along and associated with stylolites or which cuts across stylolites (Figure 13).



**Figure 13. Mesogenetic porosity. A - Fractures, which cross and therefore post-date stylolites, with porosity from the dissolution of post-fracture cement. Adjoining this area is selective dissolution of oolites. B - Oomoldic porosity adjoining and post-dating the formation of a stylolite.**

As pressure-dissolution features, stylolites form most commonly in the deep-burial environment. So, if the timing of porosity formation can be determined by the law of cross-cutting relationships to be *younger* than the stylolites, and the rocks are still buried, then porosity formation must be the result of mesogenetic dissolution; (2) pores that cut across cements that contain hydrocarbon inclusions; (3) pores that cut across or which are intimately associated with fluorite, metal sulfides such as galena and sphalerite, and

sometimes pyrite/marcasite, which are common Mississippian Valley-type precipitates from migrating deep-basinal fluids (e.g., Cathles and Smith, 1983; Clendenin and Duane, 1990); and (4) pores that cut across saddle dolomite (Figure 14). Saddle dolomite, otherwise known as baroque dolomite or "pearl spar," is a cement or replacive mineral that is readily identified by its characteristic curved crystal faces, common opaque-white color, and relatively coarse crystal size, that commonly forms in the deep-burial environment (Radke and Mathis, 1980). Hence, if pores cut across saddle dolomite, then porosity formation must be mesogenetic if the rocks are still buried. Are there examples of mesogenetic porosity in carbonate reservoirs in Kansas? Undoubtedly, but specific published instances where mesogenetic porosity is present are not available.



**Figure 14. Mesogenetic porosity. A - Thin-section photomicrograph of curved saddle dolomite (with typical undulose extinction) filling a pore. B - Core slab showing coarse crystalline saddle dolomite that has been partially dissolved to exhume (vuggy) porosity**

### **Microporosity, Pinpoint Porosity, and Chalky Porosity**

The term *microporosity* refers to any very small pores that can be recognized only with the aid of a high-powered binocular microscope or thin-section (Choquette and Pray, 1970; Pittman, 1971). Micropores, otherwise known as pinpoint pores, may variously represent: (1) birdseye pores in tidal flat deposits; (2) intraparticle pores within small particles; (3) intercrystalline pores between dolomite crystals or between calcite cement crystals; (4) intercrystalline pores within the nuclei or cortices of oolites; or (5)

intracrystalline pores within individual dolomite or calcite cement crystals. In some cases, whereas matrix microporosity/pinpoint porosity may not be very permeable for oil, it very well may be permeable enough for natural gas (e.g., Roehl, 1985; Ruzyla and Friedman, 1985). *Chalky porosity* is a term that refers to microporosity that commonly forms in highly weathered or otherwise highly diagenetically altered carbonate rocks (Pittman, 1970; Reeckmann and Friedman, 1982) that, as a consequence of being strongly altered, are very soft. Although true chalks (i.e., those limestones of Cretaceous to Paleogene age that are composed dominantly of coccoliths) commonly contain microporosity, the presence of chalky porosity does not necessarily indicate that the rocks under consideration are chalks. In fact, tripolitic cherts commonly have microporosity, and because the rocks are relatively soft, such porosity can be referred to as chalky. I point out that microporosity, and specifically chalky porosity, can form in carbonate rocks (and cherts) beneath unconformities as well as in deeply buried rocks, and as such, their presence does not necessarily imply the existence of a stratigraphically-nearby unconformity.

### Conclusions

Secondary porosity in carbonate rocks, in both limestone and dolomite, can be formed by: (1) freshwater dissolution either in the subunconformity meteoric, eogenetic or telogenetic environment, or; (2) by dissolution by chemically aggressive subsurface fluids, generated during maturation of organic matter in source rocks, in the deep-burial (mesogenetic) environment. Although pore types formed in these environments are similar, their origin often can be determined by careful observation, thin-section petrography, and stable carbon-oxygen isotope analysis.

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