

CONTRASTING STYLES OF CHEMICAL COMPACTION IN THE UPPER PENNSYLVANIAN DENNIS LIMESTONE IN THE MIDCONTINENT REGION, U.S.A.¹

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ABSTRACT: The Upper Pennsylvanian Dennis Limestone of the U.S. Midcontinent region displays a wide variety of pressure dissolution features, the distribution of which was governed largely by patterns of earlier diagenesis. Intergranular compaction in grainstones, measured quantitatively using a newly-developed compaction index, is generally restricted to diagenetic facies that received little cement. Compaction is greatest in the most seaward and stratigraphically lowest facies, where degrading neomorphism, an apparent result of compaction, also occurred despite shallow burial and absence of tectonic deformation. The least compaction in grainstones occurred in the two well-cemented diagenetic facies that are geographically and stratigraphically in the middle of the formation. Intermediate degrees of compaction occurred in the uppermost, most landward diagenetic facies, which was less extensively cemented than the facies just below. Patterns of earlier cementation, and perhaps dissolution of early cement, were more important than porewater chemistry during burial in determining the extent of pressure dissolution in Dennis Formation limestones.

Stylolites occur in the most extensively cemented diagenetic facies, where petrographic relationships suggest that stylolitization was initiated by destruction of large voids generated by dissolution of phylloid algae. Dissolution seams, but not stylolites, occur in clay-rich fine-grained dolomitic limestones. Stylolitization, intergranular pressure dissolution, and degrading neomorphism occurred at burial depths of no more than 800 to 1500 m, but these processes may have been promoted by burial temperatures up to 160°C.

The style and extent of chemical compaction vary greatly in the Dennis Formation across as little as 5 m vertically and 40 km laterally. This extreme variability within one formation illustrates that compaction and porosity destruction in limestones may often be unpredictable from a purely geographic or stratigraphic perspective, and emphasizes the necessity of understanding earlier diagenetic and hydrologic systems within any given formation.

INTRODUCTION

Chemical compaction varies greatly in both style and extent in the limestone members of the cyclothemic Upper Pennsylvanian Dennis Formation in the Midcontinent region of the United States. Intergranular compaction in grainstones varies from negligible to intense between different diagenetic facies that were previously delineated by variation in degree of grain preservation. Similarly, stylolites and dissolution seams are common in some diagenetic and depositional facies but absent elsewhere, and degrading neomorphism, an apparent result of compaction, occurs in one diagenetic facies.

This paper documents the differing styles of chemical compaction that occur in the Dennis Formation and examines their distribution with respect to the previously recognized diagenetic facies and patterns of cementation. In doing so, it employs a new compaction index for the quantitative petrographic analysis of grain-to-grain compaction in grainstones. The significance of chemical compaction lies both in its modification of limestone fabrics and in its double destruction of porosity, once at the site of compaction and later where the resulting dissolved constituents are precipitated as cements (e.g., Finkel and Wilkinson 1990). Other regional studies on the variability of compaction in carbonates include work by Meyers (1980) and Meyers and Hill (1983) on intergranular compaction, but not stylolitization, in Mississippian skeletal limestones across a 300 km outcrop belt in New Mexico. In contrast, the present study examines regional variation in both intergranular compaction and stylolitization in more lithologically diverse Pennsylvanian carbonates across a 500 km outcrop belt.

SUMMARY OF DENNIS FORMATION DEPOSITIONAL AND DIAGENETIC FACIES

Deposition

The Missourian or Stephanian Dennis Formation in Iowa, Missouri, and Kansas consists of three members: the Canville Limestone Member (a thin limestone present only in southeastern Kansas), the overlying and widespread Stark Shale Member, and the similarly widespread Winterset Limestone Member (Figs. 1, 2). The Canville Limestone consists largely of skeletal grainstones and packstones. The Stark Shale is a locally phosphatic black shale. The Winterset Limestone consists of algal and skeletal wackestones and packstones with local skeletal and oolitic grainstones that were deposited over a broad shelf with shorelines to the northeast of the study area (Fig. 1; for a cross-section showing depositional facies, see Railsback 1984). However, the Winterset thickens into mounds or banks of phylloid algal packstones and boundstones punctuated by oolitic grainstone beds in southeastern Kansas. South of that algal mound belt lies a much thinner sequence of sandy skeletal packstones and grainstones. Above and below the Dennis Formation, grey shales separate the formation from similar underlying and overlying limestone-black shale-limestone cycles. The Dennis Formation extends west into the subsurface of central and western Kansas and Nebraska, where Dubois (1979) studied the development of porosity in a unit correlative with the Winterset Limestone.

Midcontinent Pennsylvanian cyclic sequences like the Dennis Formation were interpreted by Heckel (1977) to have been deposited during transgressive-regressive cycles related to Gondwanan glacial fluctuations. In this scheme, lower limestone members like the Canville Limestone represent transgressive phases, "core" black

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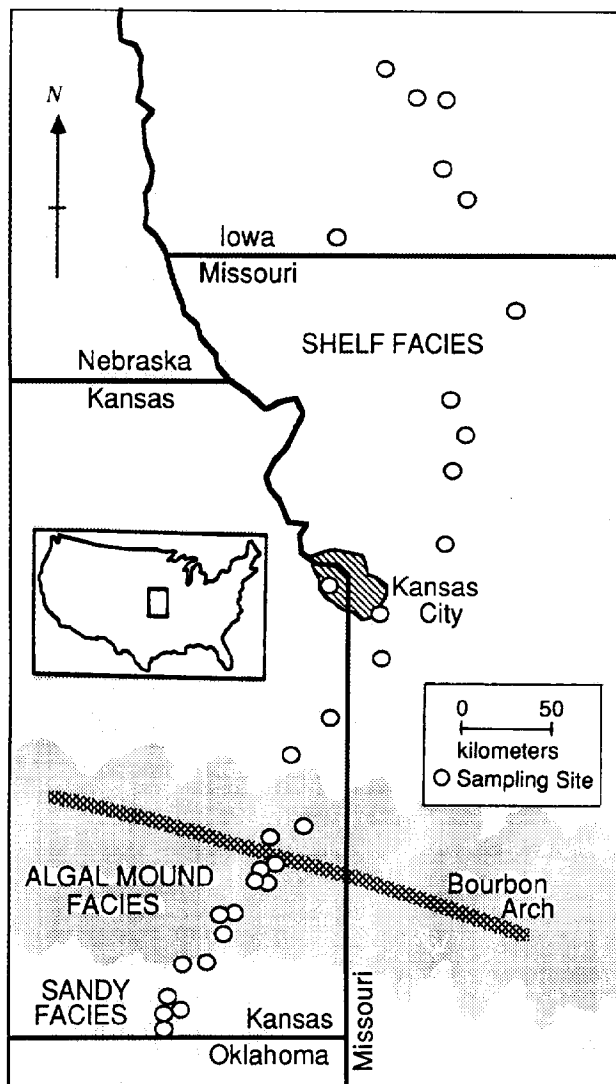


FIG. 1.—Map of Midcontinent region showing sampling sites and depositional facies of the Dennis Formation. Depositional facies are modified from Heckel (1978). Bourbon Arch separates Forest City Basin (to northeast) from Cherokee Basin (to southwest). For more information on sampling sites, see Railsback (1984).

shales like the Stark Shale represent sea-level highstands, and upper limestone members like the Winterset Limestone represent regressive deposition. The extent of these transgressions and regressions can be seen in Heckel's (1980) paleogeographic reconstructions, which indicate that during highstands, seas extended eastward to the Appalachians, 1400 km to the northeast; during lowstands they withdrew into western Oklahoma and Texas, 500 km to the west.

Eogenetic Diagenesis

Railsback (1984) used the degree of preservation of originally unstable (aragonite or high-Mg calcite) grains to delineate regional diagenetic facies within the lime-

stone members of the Dennis Formation (Fig. 2). The uppermost and northernmost, and thus most shoreward, diagenetic facies (Facies A) consists of limestones in which molluscan and phylloid algal grains that were originally aragonitic retain no trace of their original microstructure (Railsback 1984). For example, molluscan and phylloid algal grains in this facies consist only of mosaics of clear calcite, presumably resulting from complete dissolution of aragonite and subsequent filling with calcite cement. Facies B, which is stratigraphically lower, consists of limestones in which at least a few of these originally unstable grains have some remnants of their original microstructure, such as traces of organic matter, lines of inclusions parallel to shell walls, and "dust patches" (see Bathurst 1983). Facies C consists of limestones in which most but not all of the originally unstable grains have remnants of their microstructure, suggesting that they have undergone neomorphic alteration, and some grains retain much of that original microstructure. Facies D, the southernmost and lowermost facies, consists of limestones in which all grains of unstable original mineralogies preserve some relic of their original microstructure. Facies E, which occurs at the base of the Winterset Limestone, consists of dolomitic mudstones and wackestones and, unlike the other diagenetic facies, is not defined by a particular style of preservation of unstable grains.

Railsback (1984) interpreted the differing degrees of preservation of unstable grains in Facies A through D as the result of differing degrees of invasion by meteoric waters during the regression after Winterset deposition (during the "eogenetic diagenesis" of Choquette and Pray 1970). Petrographic evidence indicates that waters undersaturated with respect to CaCO_3 caused extensive dissolution of unstable grains in Facies A, which was nearest recharge areas, and dissolution of CaCO_3 resulted in development of a saturated meteoric zone that allowed progressively less dissolution in Facies B, C, and D. As regression progressed, the migration of these waters through the formation was promoted by more permeable beds (e.g., grainstones) and retarded by relatively impermeable beds like the Stark Shale.

Railsback (1984) also interpreted the distribution of cement in Dennis Formation limestones (discussed below) as largely the result of transfer of CaCO_3 from Facies A to cements in Facies B and C, whereas Facies D received little cement. Most cements in the Dennis Formation are sparry calcite, but ferroan saddle dolomite also occurs as a late cement in all diagenetic facies.

Because they cut across depositional facies, Diagenetic Facies A, B, C, and D generally have similar carbonate grain types and non-carbonate mineralogies. Minor exceptions are that quartz sand and silt occur only in Facies D, and Facies C contains fewer foraminiferal grains and intraclasts than the other facies. Facies A through D thus are sufficiently similar to allow comparison of compaction independent of effects caused by variation in mineralogies and depositional fabrics. Diagenetic Facies E contains few grains to measure intergranular compaction and has been dolomitized; it therefore receives little attention in this paper.

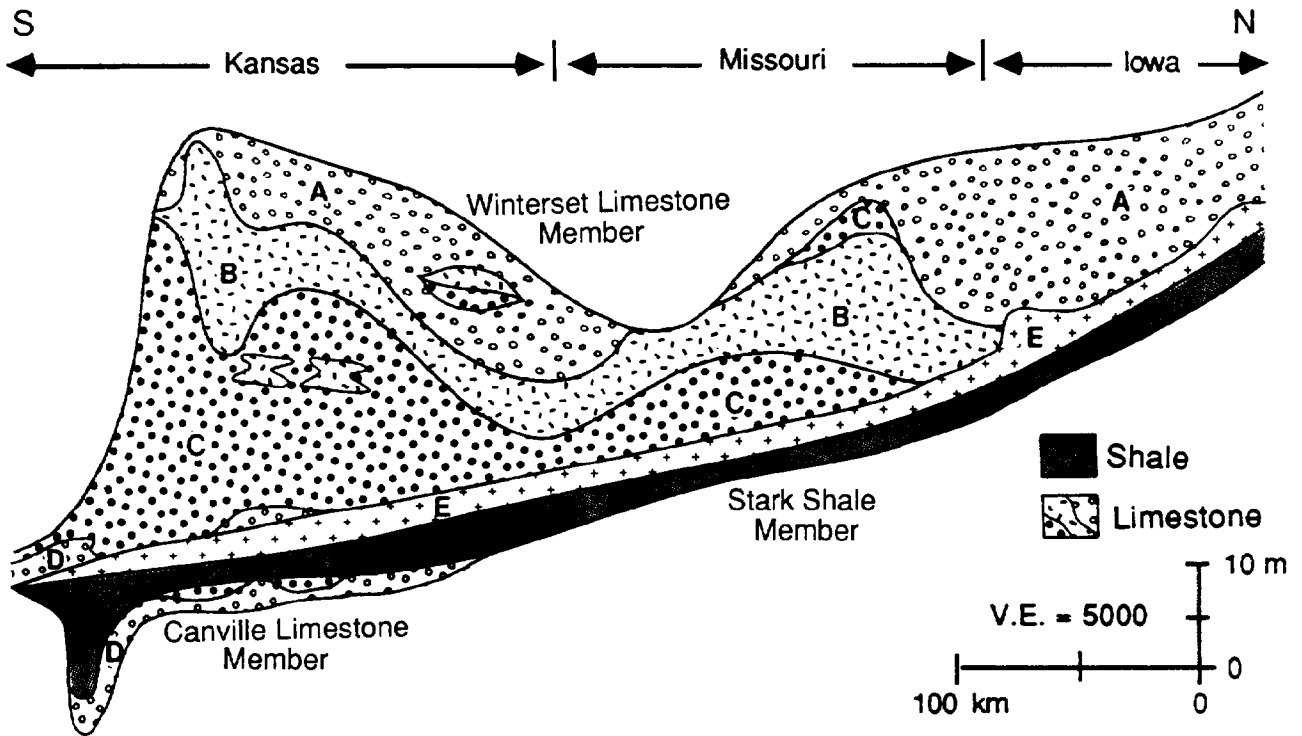


FIG. 2.—Stratigraphic cross-section showing regional diagenetic facies in Dennis Formation. Diagenetic Facies A through D represent increasing preservation of originally unstable grains (i.e., grains of aragonite or high-Mg calcite); Facies E contains fine-grained dolomite. Cross-section constructed to reflect topography shortly after deposition; see text and Railsback (1984) for inferred paleohydrology and movement of diagenetic environments.

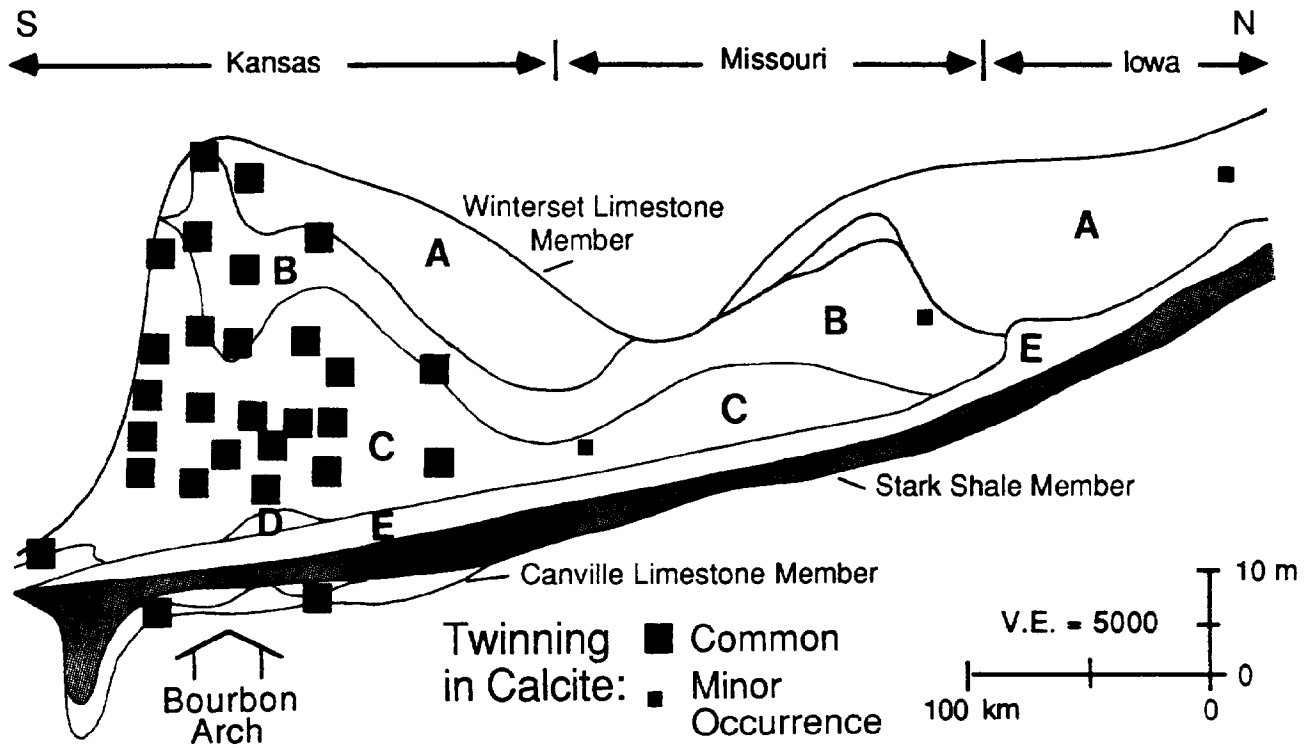


FIG. 3.—Distribution of twinned calcite in Dennis Formation. "Common occurrence" denotes samples in which roughly one fourth of large calcite crystals are twinned. Twinned calcites (interpreted to be indicators of tectonic stress) are generally restricted to area of Bourbon Arch in southeastern Kansas and show no relation to distribution of compactional effects.

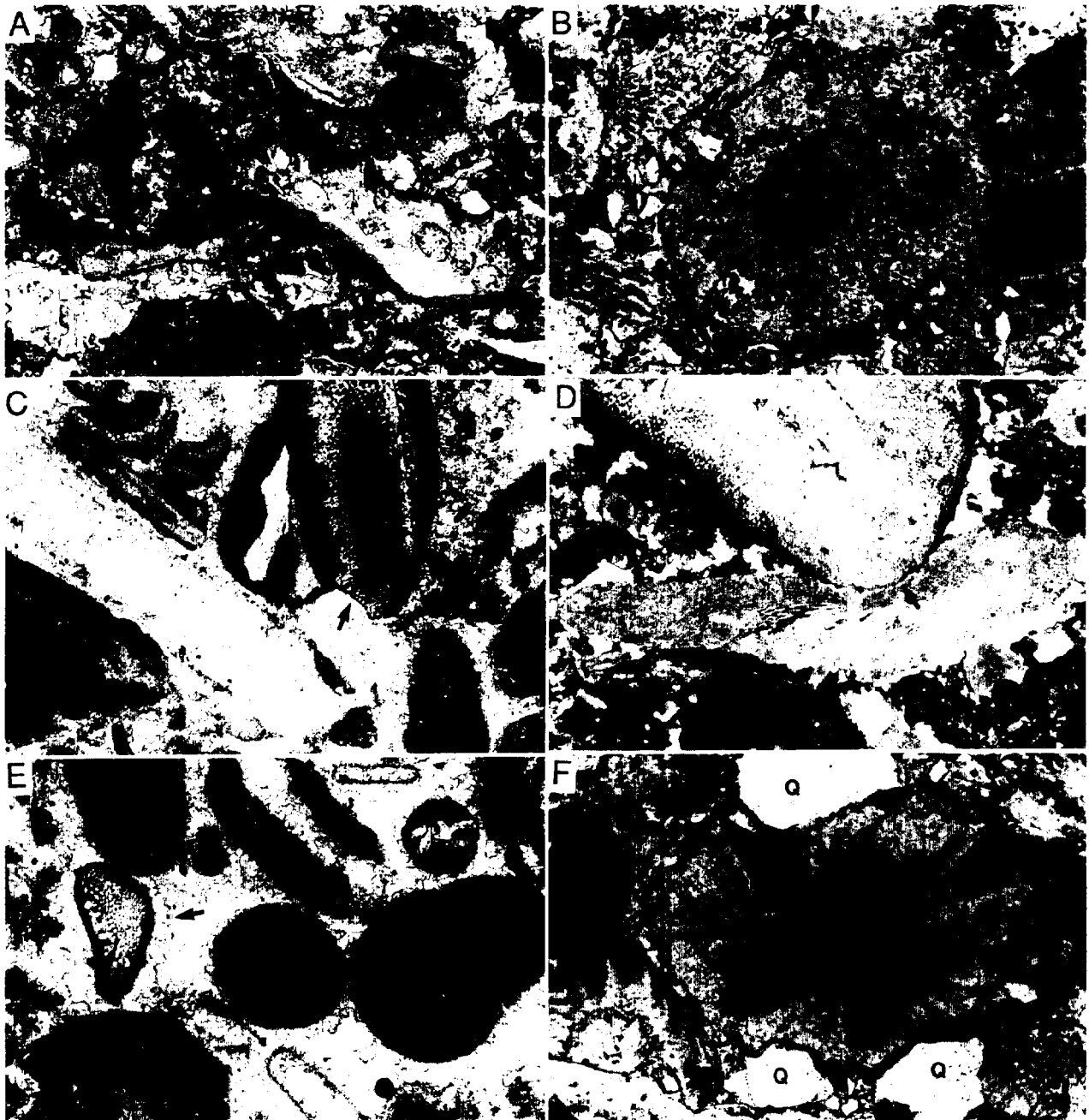


FIG. 4. — Thin-section photomicrographs illustrating intergranular contacts and compaction. Stratigraphic up as shown in all photos. **A)** Pressure-dissolved and flattened contacts between bryozoan and other skeletal grains in Facies D; white areas are quartz grains. Fabric resembles "fitted fabric" of Logan and Semeniuk (1976) and Buxton and Sibley (1981). Width = 2.4 mm. **B)** Echinoderm grain in Facies D consisting of at least 4 areas with different optic orientations due to degrading neomorphism or crystal diminution. Photographed with polarizer and analyzer at ca. 20°; numbers label regions with different optic orientations. Width = 0.85 mm. **C)** Evidence of compaction in Facies A. Note concave pressure dissolution contact between two echinoderm grains (arrow) and close packing of grains. Width = 2.2 mm. **D)** Broken and pressure-dissolved (arrow) echinoderm grain in Facies A. Width = 2.90 mm. **E)** Oolitic grainstone in Facies C showing loose packing of grains and little evidence of pressure dissolution. Note possible early marine cements (arrow). Width = 2.2 mm. **F)** Bryozoan fragment in Facies D pressure-dissolved at contacts with quartz grains (Q) above and below, which show little evidence of pressure-dissolution. Width = 0.90 mm.

Tectonic Effects and Compaction

Twinned calcites are rare in the Dennis Formation except in southeastern Kansas (Fig. 3), where twinning is present in roughly one quarter of all large calcite crystals.

In this area, twinning occurs in both eogenetic and burial cements, suggesting that it resulted not from growth but from stress. This stress apparently resulted from movement of the Bourbon Arch, a subtle structural feature in

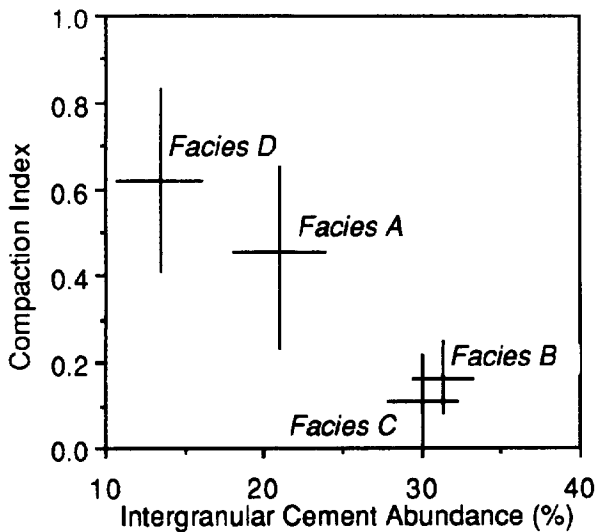


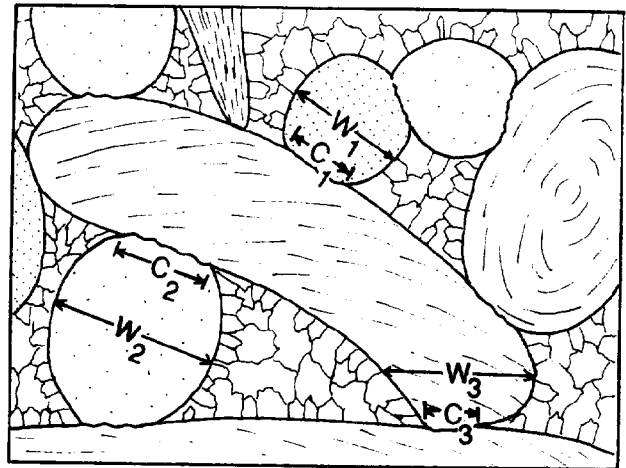
FIG. 5.—Variation of compaction index measurements and cement abundances in grainstones in four diagenetic facies discussed in text. Lines indicate one standard deviation each way from means (at intersections of lines). Compaction index is defined in Figure 6. Linear regression line through means ($R = 0.98$) extrapolates to compaction index of 1.02 at 0% cement and to 35.8% cement at a compaction index of 0.0.

southeastern Kansas that separates the Forest City Basin to the northeast from the Cherokee Basin to the southwest (Fig. 1). Twinned calcites in the vicinity of the Bourbon Arch occur in all diagenetic facies and in both limestone members of the Dennis Formation (Fig. 3), and their distribution shows no relation to that of compactional effects. The distribution of calcite twinning, which is commonly used as an indicator of stress (e.g., Friedman et al. 1976), thus suggests that compactional fabrics within the Dennis Formation are not the result of tectonic stress.

INTERGRANULAR COMPACTION IN DENNIS FORMATION GRAINSTONES

In grainstones in Facies D, results of intergranular chemical compaction, such as interpenetrating grain contacts and elongate or concavo-convex contacts, are very common (for quantification, see section on compaction index measurements, below) (Fig. 4A). Considerable mechanical compaction has also occurred in Facies D, and elongate grains such as brachiopod fragments are generally broken or deformed. This compaction has resulted in development of a texture approaching "fitted fabric" (cf. Buxton and Sibley 1981; Bathurst 1991). According to Buxton and Sibley (1981) and Logan and Semeniuk (1976), part of the definition of "fitted fabric" is that each grain be completely in contact with surrounding grains, with no cement. Intergranular cement is only 11 to 15% of total rock volume in Facies D, much lower than that in the other facies (Fig. 5).

Evidence of pressure dissolution between grains also occurs in grainstones in Facies A but much less commonly than in Facies D (Fig. 4C, D). Fabrics in Facies A in no way resemble fitted fabric, and sutured contacts are not



$$\text{Compaction Index} = \frac{C_i}{W_i}$$

C_i = Length of grain contact
 W_i = Width of smaller grain (parallel to contact)

FIG. 6.—Definition of compaction index used in this paper. Index measures intergranular compaction at specific grain contact as ratio of length of contact to width of smaller grain, measured parallel to the contact, to reflect extent to which original "point" contact has been lengthened.

common. However, linear contacts seem anomalously abundant between typically rounded to well-rounded grains, and some contacts are interpenetrative (Fig. 4C, D). Intergranular cement is about 17 to 24%, more than in Facies D but less than in Facies B and C (Fig. 5). Effects of mechanical compaction such as grain breakage are also found (Fig. 4D) but are not as common as in Facies D.

Grainstones in Facies B and C show little readily visible evidence of intergranular compaction; most grains show no contacts or only "point" contacts in thin section (Fig. 4E). Intergranular cement is 30 to 33% of rock volume in Facies B and 28 to 32% in Facies C, higher than in the other diagenetic facies (Fig. 5).

COMPACTION INDEX MEASUREMENTS

Definition and Assessment

In order to quantify the varying degrees of intergranular chemical compaction in the diagenetic facies of the Dennis Formation, a "compaction index" was developed for evaluation of grainstones. This index should not be confused with the compaction index which Bathurst (1987) used to measure mechanical compaction and grain rotation. For any given grain-to-grain contact, the compaction index used herein is defined as the length of the grain contact divided by the maximum width of the smaller grain, where the latter is measured parallel to the contact in question (Fig. 6). The index thus measures the extent to which an original "point" contact has been

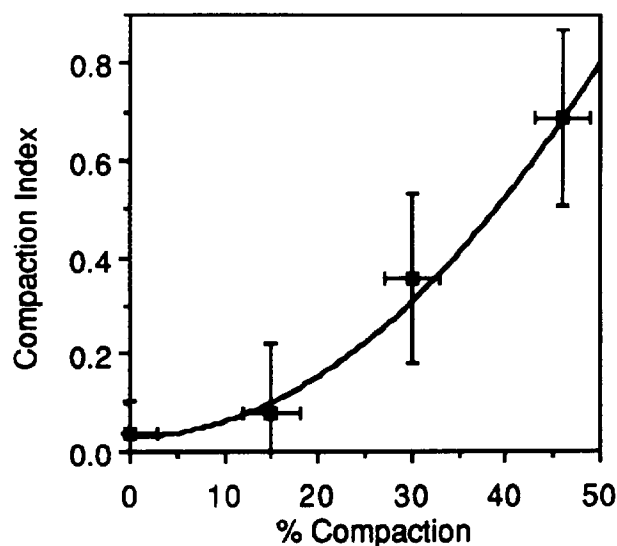


FIG. 7.—Results of experimental compaction of kaolin spheres to test validity of compaction index. Compaction index is defined in Figure 6; % compaction is $100 \cdot (1 - V_f/V_o)$, where V_f and V_o are the final and original volumes of grains plus porosity. Vertical bars indicate one standard deviation each way from mean compaction index; horizontal bars show uncertainty of experimenter's determination of volume reduction. Equation for regression curve is $y = 0.00031x^2 + 0.03$; $R = 1.00$. Results indicate that compaction index is a useful measure of compaction; however, results cannot be used to calibrate index to volume reduction in carbonate rocks because of the plasticity of the experimental material and because the experiments did not involve dissolution and mass transfer.

lengthened by compactive processes to its theoretical maximum (the width of the smaller of the two grains). A grain in contact with no other grain yields a compaction index of zero. The only complication arises in the case of long or irregularly shaped grains in bridging positions that preclude compaction to the maximum width of the smaller grain (case 3 of Fig. 6). In this case the petrographer is left to decide the most appropriate position to measure the effective maximum width of the smaller grain.

The validity of this index as a measure of compaction was tested in two ways. As an experimental test, spheres of kaolin 8 mm in diameter were compacted to varying degrees in cylinders 31 mm in diameter and 50 mm long, and then impregnated and sectioned. Compaction index measurements from these experiments are correlative with volume reduction, supporting the use of the index as a measure of compaction (Fig. 7). As a petrographic test, compaction index measurements in Dennis Formation grainstones were compared with measurements of contacts per grain, a parameter that was used, for example, to measure compaction in sandstones by Taylor (1950). The correlation of compaction index measurements with contacts per grain (Fig. 8), combined with the experimental results, suggests that the compaction index is a valid petrographic measure of compaction.

The compaction index is designed to minimize the effects of unusual grain shapes and to avoid inconsistencies caused by variations in grain size (discussed below). This index also does not require subjective estimation of orig-

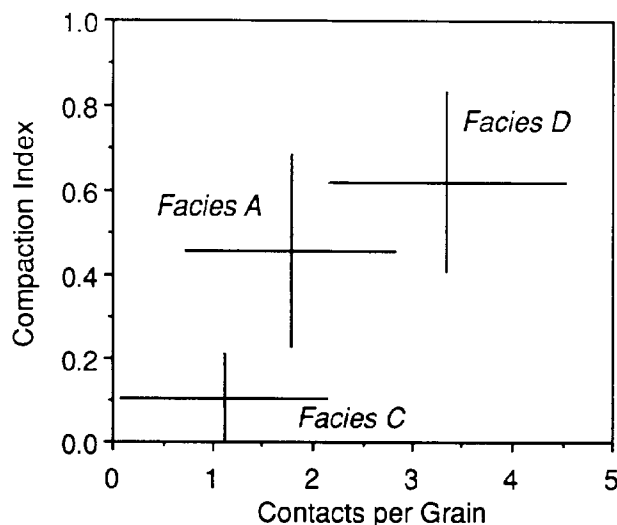


FIG. 8.—Variation of compaction index measurements and contacts per grain in Facies A, C, and D. Lines indicate one standard deviation each way from means (at intersections of lines). "Contacts per grain" were measured only on grains no more than 5 times as long in any direction than average grain size to avoid anomalies caused by poor sorting, as discussed in text (five times is an arbitrary limit). Plot shows that compaction index is correlative with contacts per grain and is at least as effective in distinguishing between different facies as contacts per grain.

inal grain shapes, in contrast to measurement of grain overlap (e.g., Sibley and Blatt 1976; McBride et al. 1991). The compaction index has advantages over counts of contacts per grain, because the latter measure of compaction increases (or its range increases) with poorer sorting. For example, a brachiopod fragment 18 mm long in the silty portion of Facies D was in contact with 48 other carbonate and quartz grains. Another shortfall in measuring contacts per grain is that it does not allow study of the effect of specific grain types, because "contacts per grain" depends on both the grain in question (for example, an ooid) and its potentially diverse neighbors (perhaps including a quartz grain and an oncolite). Because the compaction index described above applies to specific grain-to-grain contacts and not to entire fabrics, it can be used to analyze contacts between grains of a particular type. As an example, contacts between grains of differing mineralogies are discussed below.

Comparison of Grainstones in Different Diagenetic Facies

Histograms of measurements of the compaction index for contacts between calcite grains in grainstones in Facies A, B, C, and D are shown in Figure 9. The histograms summarize 349 measurements made on 16 samples from 12 localities across the 4 facies; those samples represent all the grainstones from over 300 samples of the Dennis Formation. The data were compiled from measurements of the contacts of grains determined by traverses across thin sections with a mechanical stage that generated a point distance equal to 4 to 8 grain diameters, which

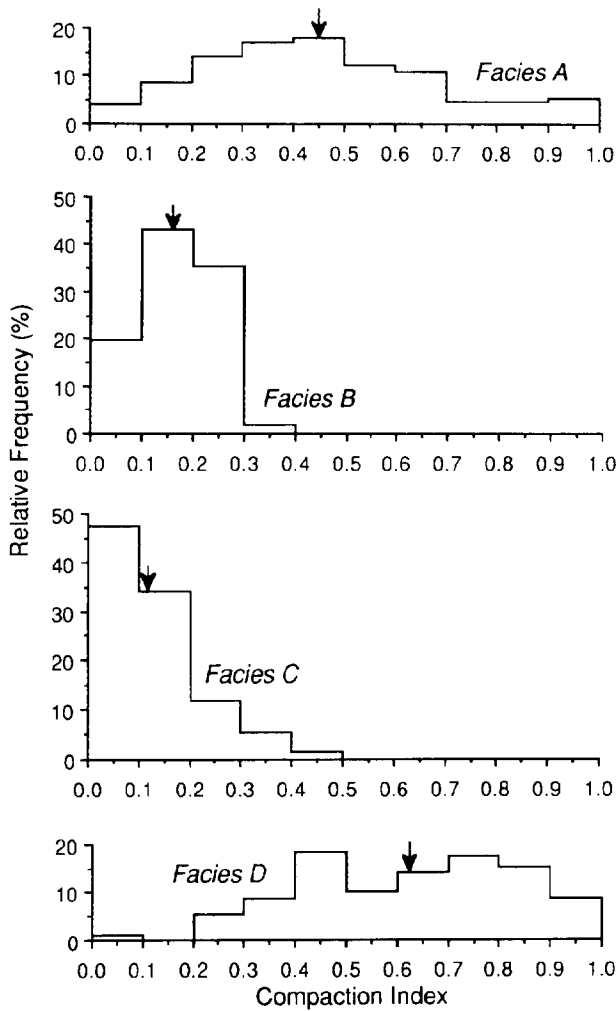


FIG. 9.—Histograms of compaction index measurements in grainstones in Diagenetic Facies A, B, C, and D. Mean measurements for all four facies (arrows) differ with statistical significance from each other ($P < 0.001$). Data support the qualitative observation that intergranular compaction is greatest in Facies D and least in Facies B and C (see Fig. 4).

presumably generated a random sample of grains (see Van der Plas and Tobi 1965). However, measurements were made only for contacts between rigid calcitic grains (e.g., between brachiopod, echinoderm, bryozoan, and foraminiferal grains, ooids, and intraclasts). These different grain types occur in all 4 diagenetic facies, but foraminiferal grains and intraclasts are less abundant in Facies C than in other facies.

Mean compaction index measurements for grainstones in all 4 facies differ with statistical significance from each other ($P < 0.001$) and are greatest in Facies D (mean of 0.62) and least in Facies B and C (means of 0.17 and 0.11, respectively). These results lend support to the distinction of the 4 diagenetic facies, which were originally recognized by other diagenetic indicators. More importantly for this study, these quantitative results support the qualitative petrographic observations described above that intergranular compaction is greatest in Facies D

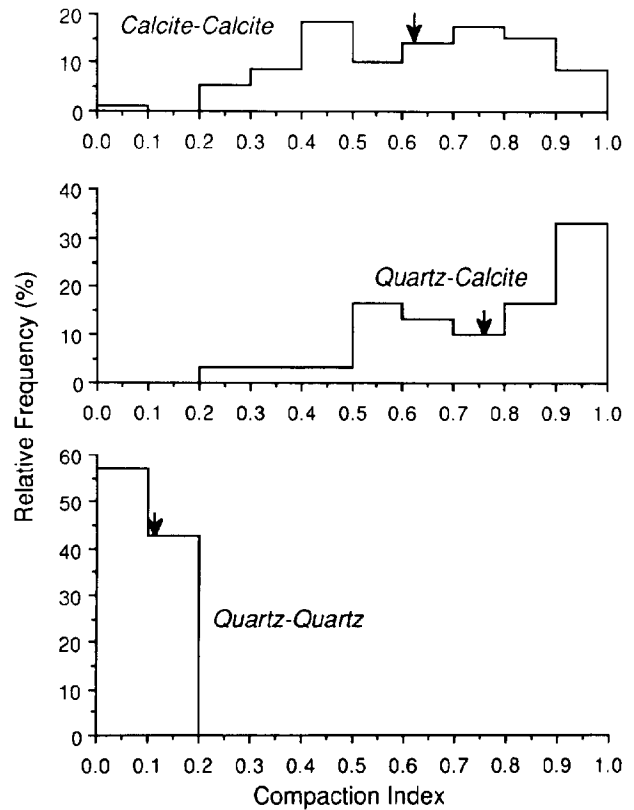


FIG. 10.—Histograms of compaction index measurements for contacts between grains of differing mineralogies in Facies D. Arrows indicate means.

grainstones, significant in Facies A, and least in Facies B and C.

Comparison of Grain Types within Facies

To examine the control of mineralogy on intergranular compaction, measurements of compaction index were also made in Facies D at quartz-calcite intergranular contacts and quartz-quartz intergranular contacts (Fig. 10). The compaction index is greatest for quartz-calcite contacts (mean = 0.76) and is much less for quartz-quartz contacts (mean = 0.11) than for calcite-calcite contacts (mean = 0.62). These results presumably reflect the greater susceptibility of calcite to pressure dissolution than quartz (see Fig. 4F) and support Trurnit's (1968) ranking of minerals according to susceptibility to pressure dissolution, in which calcite was ranked second and quartz was ninth of sixteen.

DEGRADING NEOMORPHISM IN DIAGENETIC FACIES D

Another apparent result of compaction in Facies D is degrading neomorphism (Folk 1965), which is also known as grain diminution (Orme and Brown 1963) or crystal diminution (Dixon and Wright 1983). Echinoderm grains in Facies D commonly do not display unit extinction but



FIG. 11.—Photonegative print showing stylolite (dark arrows) in Facies C. At right is encrusting bryozoan around calcite and dolomite cements (C) that fill space left by dissolution of encrusted phylloid algal blade. Stylolite extends from that cement to spar (S) after phylloid algae at left of photo. Width of photograph = 16 mm.

instead consist of up to 6 optically distinct regions (Fig. 4B). For example, of 105 echinoderm grains examined during mechanical traverses in a thin section from Facies D, 52 consisted of two or more optically distinct regions in cross-polarized light. The crystal boundaries of these regions are irregular rather than planar, and the rectilinear patterns of inclusions typical of echinoderm grains are preserved within the various regions, suggesting a neomorphic rather than precipitative origin. Voll (1960) documented similar but more finely crystalline fabrics in crinoid ossicles and calcite cements, and he attributed those fabrics to low-temperature recrystallization caused by "light straining". Folk (1965) called the development of this fabric "degrading strain recrystallization", which he considered a form of degrading neomorphism. The presence of echinoderm grains recrystallized due to strain in Facies D but not in other diagenetic facies suggests that recrystallization resulted from the same stresses that caused the mechanical and chemical compaction in Facies D noted above.

STYLOLITES AND DISSOLUTION SEAMS IN THE DENNIS FORMATION

Through-going pressure-dissolution features, as opposed to sutured or microstylolitic contacts between individual grains, occur in the Dennis Formation as dissolution seams and stylolites (using the classification of Buxton and Sibley 1981). Dissolution seams occur in clay-rich lime mudstones and wackestones in Diagenetic Facies B and C of the Winterset Limestone and are particularly common at margins of limestones interbedded with shales. The rocks in which these seams occur also contain dolomite rhombs 0.01 to 0.10 mm in size that make up 15 to 35% of the rock volume. (However, dissolution

seams do not occur in Diagenetic Facies E, where dolomite abundances are 80 to 100%). The argillaceous seam-bearing carbonates are probably comparable to the "clay-rich zones" in younger Midcontinent Pennsylvanian carbonates in which McNeice (1988) found microstylolites and truncated grains as evidence of chemical compaction.

Stylolites occur only where Diagenetic Facies B and C overprint the algal mound depositional facies. These stylolites (Types 1A and 1B of Buxton and Sibley 1981) occur in or near areas of void-filling calcite and ferroan dolomite cement (Fig. 11) in algal wackestones and packstones. These cements fill shelter porosity beneath phylloid algae or other grains, or secondary porosity resulting from dissolution of phylloid algae. Stylolites occur at the margins of these cements, and some but not all transect the cements as well. The stylolites extend laterally for no more than about 8 cm, and their amplitudes are less than 4 mm.

DISCUSSION

Intergranular Compaction and Cementation

The differences in intergranular compaction in grainstones discussed above can be readily interpreted in terms of prior cementation. Figure 5 shows that compaction index is inversely correlative with abundance of intergranular cement. Facies D, the facies furthest from recharge sources in early diagenesis, underwent little cementation, leaving uncemented grains to support an increasing overburden that generated localized stresses sufficient for intergranular pressure dissolution. In contrast, extensive cementation in Facies B and C (strati-

graphically in the middle of the formation) apparently stabilized the grain framework, so that little intergranular compaction occurred. Facies A, nearest recharge sources, underwent less extensive cementation than Facies B and C, which allowed more compaction than in Facies B and C but much less than in Facies D.

The increase in intergranular chemical compaction in grainstones from Facies B and C to Facies A to Facies D supports the widely held view that intergranular compaction increases with decreasing early cementation (e.g., Purser 1978; Bathurst 1983; Hird and Tucker 1988). Meyers and Hill (1983) likewise attributed regional variations in compaction to variation in the extent of early diagenesis, in that areas of minimal early cement did not have an adequate rock framework to support burial stress. However, the Dennis Formation presents a more complex example. In a study by Meyers and Hill (1983), variation in compaction was essentially geographic, although it was controlled by diagenesis and pore water chemistry. In the Dennis Formation, because variation in compaction is dependent on diagenetic facies, similar lithologies from the same geographic locality (e.g., grainstones 5 m apart in Diagenetic Facies C and D in southeastern Kansas) may exhibit greatly differing degrees of compaction.

Although lesser cementation probably accounts for the enhanced compaction in Facies A, it is also possible that loss of early cements allowed compaction. Railsback (1984) noted that inclusion-rich prismatic cements, interpreted as marine cements, were rare in Facies A but occurred in stratigraphically lower diagenetic facies. This difference was attributed to early dissolution of marine cements rather than to their original distribution (Railsback 1984). Thus, loss of stabilizing marine cements may in part account for the unequal distribution of overburden pressure that caused more pressure dissolution in Facies A than in Facies B and C.

Scholle and Halley (1985) generalized from several studies that water chemistry is commonly a major control on rates of pressure dissolution, but this is probably not the case in Dennis Formation limestones. Preservation of originally unstable grains suggests that Facies D probably entered the burial environment with pore fluids that were saturated with respect to CaCO_3 . Such pore fluids would not have been more conducive to pressure dissolution than pore waters in other facies as they entered the burial environment, yet Facies D underwent the most extensive intergranular compaction. Little is known about the chemistry of burial porewaters in the Dennis Formation, but the similarity of burial cements, especially ferroan dolomites, along the outcrop belt suggests little variation in the chemistry of burial fluids. Clearly patterns of cementation were more important than porewater chemistry in determining the extent of pressure dissolution in the Dennis Formation.

Controls on Stylolitization

The occurrence of stylolites in Facies B and C suggests that stylolitization was limited to those portions of the

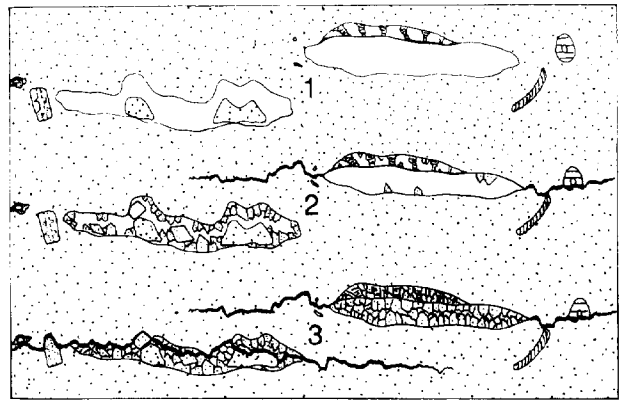


FIG. 12.—Sequence of sketches showing interpreted generation of stylolites near voids derived from dissolution of phylloid algae in Diagenetic Facies B and C. Step 1: Voids in wackestone left by dissolution of phylloid algal blades. Roof of void at left has partially collapsed; features like encrusting bryozoan at right demonstrate that these voids represent prior grains rather than vugs. Step 2: Partial cementation of void at left; partial reduction of void at right requires removal of rock at margins by pressure dissolution and stylolite development. Step 3: Reduction of remaining void at left causes stylolite through both cement and surrounding rock; complete cementation of remaining void at right. Sketches in Step 3 schematically represent stylolites found 3 m apart vertically.

Dennis Formation that were cemented early to generate a rigid rock framework. This is similar to trends found in other limestones by Garrison and Kennedy (1977) and Buxton and Sibley (1981), who concluded that stylolite development is most common in well-cemented limestones. Wanless (1979) similarly concluded that stylolites often form in limestones “having structural resistance to stress”. However, some porosity must have remained within the stylolitized rock mass to allow solution transfer (Bathurst 1990).

Limitation of stylolites to rocks rich in phylloid algae in Facies B and C suggests that stylolite development was limited to areas of large voids that concentrated stress on areas between or around voids. Stylolites at the margins of phylloid algal voids probably reflect the removal of surrounding rock by pressure dissolution to accommodate the collapse of those voids (Figs. 11 and 12). The fact that some stylolites transect the void-filling cements but many do not suggests that stylolitization occurred both before and after burial cementation. Dependence of stylolitization in the Dennis Formation on large algal voids may also explain the lateral discontinuity of these stylolites.

Stylolites in the Dennis Formation did not preferentially form at lithologic contacts, in contrast to the findings of Buxton and Sibley (1981) in the Devonian Alpena Limestone of Michigan. Some stylolites in the Dennis Formation are localized at the lower margins of sparry areas. These may have resulted from the contrast in rock fabric, at microscopic scale, between spar and micrite in the manner described by Buxton and Sibley (1981) for larger-scale stylolitic contacts, but they may also be the by-product of physical destruction of shelter porosity beneath phylloid algae, as discussed above.

*Burial Depths, Pressure Dissolution, and
Degrading Neomorphism*

The only estimates of former burial depths of the Dennis Limestone come from thicknesses of overlying strata to the west in central Kansas. Addition of thicknesses of overlying Pennsylvanian and Permian units listed in the stratigraphic compilation of Zeller (1968) yields a maximum of 1380 m of Paleozoic overburden, most of which is limestone. However, the cross-sections by Merriam (1963) suggest a total Paleozoic overburden of only about 850 m, as measured in Dickinson and Graham Counties in northern Kansas, 220 to 420 km west of the Dennis Formation's outcrop belt. Merriam's (1963) isopach maps suggest a Cretaceous thickness of no more than 150 m in eastern Kansas and even less for Jurassic or Tertiary strata. Maximum possible burial depths for the Dennis Formation thus seem to be about 1550 m, but more reasonable estimates may be nearer 1000 m. For example, Coveney et al. (1987) estimated maximum burial of the Dennis Formation near Kansas City at 300 to 650 m, and McNeice (1988) cited recent estimates as low as 800 m further west.

Burial depths of approximately 1000 m are somewhat shallow for pressure dissolution and stylolitization, although examples at lesser depths exist. Dunnington (1967) indicated that extensive stylolites developed in Cretaceous limestones in Qatar under as little as 615 to 925 m of overlying sediments. Finkel and Wilkinson (1990) described stylolites in Mississippian limestones buried to about 700 m. Buxton and Sibley (1981) observed a variety of pressure dissolution features in Devonian limestones buried less than 1500 m. Intergranular dissolution and development of dissolution seams may require even less burial. Extreme examples include Bathurst's (1975) report that microstylolites developed in a clay-rich limestone in Guam under as little as 90 m of overburden and James and Bone's (1989) documentation of sutured grain contacts in Tertiary limestones buried no more than 100 m.

Degrading strain recrystallization is certainly not expected at the shallow burial depths estimated for the Dennis Formation. Commonly cited examples of degrading strain recrystallization come from regions of low-grade metamorphism (Tucker and Kendall 1973), regional compression and folding (Wardlaw 1962), or very deep burial (Borak and Friedman 1981). Even Dixon and Wright's (1983) example from South Wales, which was intended to illustrate that degrading strain recrystallization can be diagenetic rather than metamorphic, came from an area used by Gill et al. (1977) to define transitional zones from diagenesis to metamorphism. Degrading neomorphism under 800 to 1500 m overburden with no tectonic deformation thus appears anomalous.

Compactive effects at shallow burial depths in Dennis Formation limestones may, however, have been thermally enhanced. The experimental work of Ehardt (1968) and field observations of Dunham and Larter (1981) suggest that increased temperatures, as well as pressures, may promote pressure dissolution. Similarly, Wetzel (1989)

found that higher temperatures accelerated cementation of pelagic oozes, which he attributed to increased pressure dissolution, and Schmoker (1984) found that limestone porosity decreases with increasing time-temperature index, presumably in part because of compaction.

If one assumes a past geothermal gradient like the 26 to 29°C/km gradient presently found in southeast Kansas (Kron and Stix 1982), maximum burial temperatures expected in the Dennis Formation in the study area would have been 45 to 60°C. However, in Winterset Limestone samples from the Kansas City area, Coveney et al. (1987) found that primary and secondary fluid inclusions yielded homogenization temperatures ranging from 50 to 140°C. Similarly, Walton et al. (1990) reported that fluid inclusions in Pennsylvanian rocks from southeastern Kansas indicated burial temperatures from 100 to 160°C, and they suggested that these anomalously high temperatures were caused by hot brines flushed from the Arkoma Basin. Within the Dennis Formation, anomalously high temperatures are also suggested by the presence of saddle dolomite, the formation of which requires temperatures of 60 to 150°C (Radke and Mathis 1980). Thus, pressure dissolution and degrading neomorphism at shallow burial depths in the Dennis Formation may have been enhanced by high temperatures.

CONCLUSIONS

- 1) The extent of intergranular compaction in Dennis Formation grainstones varies between previously defined diagenetic facies and increases with decreasing cement abundance, suggesting that intergranular compaction in Dennis grainstones was controlled by patterns of earlier cementation. Patterns of intergranular compaction cannot be explained by strictly geographic, stratigraphic, or tectonic parameters, nor by variation in burial pore fluids.
- 2) Degrading neomorphism (or grain diminution) occurred in echinoderm grains in the most compacted diagenetic facies as the result of the same stresses that caused intergranular compaction.
- 3) The compaction index used to measure variation between diagenetic facies noted above is also useful in comparing susceptibilities of various grain types to pressure dissolution, because it is applied to individual grain contacts, not entire fabrics.
- 4) Stylolites formed in well-cemented diagenetic facies, largely as a response to physical destruction of voids beneath, or generated by the dissolution of, phylloid algae. Dissolution seams formed in clay-rich dolomitic rocks.
- 5) Stylolitization, intergranular pressure dissolution, and degrading neomorphism occurred at burial depths of no more than 1500 m and probably less than 1000 m, but these processes may have been promoted by burial temperatures of up to 160°C.
- 6) The style and extent of chemical compaction in the Dennis Formation vary greatly over short horizontal and vertical distances. The distribution of compaction effects thus presents a confusing array when viewed

in terms of strictly geographic or stratigraphic parameters, but it can be predicted from an understanding of the diagenetic facies present and their pore fluid histories. Similar diversity is likely to occur in any limestone unit in which multiple diagenetic environments have migrated through rocks of contrasting original permeabilities, as was the case in the Dennis Formation.

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