LITHOLOGIC CONTROLS ON MORPHOLOGY OF PRESSURE-DISSOLUTION SURFACES
(STYLOLITES AND DISSOLUTION SEAMS) IN PALEOZOIC CARBONATE ROCKS
FROM THE MIDWESTERN UNITED STATES

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ABSTRACT: Examination of pressure-dissolution surfaces (styolites and
dissolution seams) in carbonate rocks from the midwestern U.S. demonstrates
that the morphology of these surfaces varies with rock fabric. One can
quantify the morphology of these pressure-dissolution surfaces using four
parameters: thickness T of the insoluble residue defining the surface,
amplitude of irregularity I of the surface between vertical offsets, amplitude
of vertical offsets O of the surface, and frequency F of vertical offsets along
the surface.

I and O increase with T, reflecting progressive development of styolites,
whereas F is proportional to 1/T, apparently because the physical strength
of a styolite inhibits the formation of offsets. Mean values of T, O, and F
are greater in grainstones and packstones than in mudstones and wacke-
stones, and the abundance of dissolution seams relative to that of styolites
decreases through the lithologic spectrum from mudstones to grainstones.
These relationships suggest that development of pressure-dissolution
surfaces to form large styolites depends on the heterogeneity of the limestone
fabric.

Maximum values of T and O in grainstones correlate with cement abund-
ance. T and O are low in grain-rich, cement-poor grainstones, even though
grainstones are favored as leading or penetrating elements of styolite columns.
Dissolution seams (surfaces with T and F near zero) are found only in
rocks containing fine-grained dolomite. T and I of those seams decrease
with increasing dolomite abundance in dolomitic limestones, and T and O
are lower in dolostones than in limestones. Taken together, these relations-
ships suggest that extreme abundance of pressure-resistant compo-
nents, such as carbonate grains and dolomite, inhibit development of pres-
sure-dissolution surfaces. In contrast, styolites are most extensively
developed (i.e., T, O, and F are greatest) in carbonate rocks in which these
pressure-resistant elements are mixed with less resistant components, such
as cement and carbonate mud. The latter rocks, which include packstones,
well cemented grainstones, and slightly dolomitic limestones, may undergo
preferential removal from the stratigraphic record by pressure dissolution.

INTRODUCTION

Portions of a crystalline solid under stress are more soluble than less
stressed or unstrusted portions (DeBoer 1977a, 1977b). Dissolution of
these stressed domains leads to mass transfer through an aqueous phase,
either by diffusion or bulk flow, and can result in precipitation in less
stressed domains (DeBoer 1977a, 1977b; Rutter 1983; McClay 1977; Geis-
er and Sansone 1981). This process, termed pressure dissolution (Bat-
thurst 1987), usually results from gravitational loading by overburden or from
tectonic forces (Durney 1972; Engelder et al. 1981). Different minerals
possess differing susceptibilities to pressure dissolution (Trurnit 1968),
and so residues of less soluble minerals may remain after pressure dis-
solution of more soluble ones (e.g., Schwander et al. 1981). Calcite is
among the most pressure-susceptible minerals, and so carbonate rocks
undergo pressure dissolution in situations ranging from shallow burial
(e.g., Buxton and Sibbey 1981; Railsback 1993) to intense structural de-

In limestones, pressure dissolution causes suturing of intergranular con-
tacts at the microscopic scale and formation of macroscopic pressure-
dissolution surfaces (styolites and dissolution seams). Pressure dissolution
thus modifies limestone fabric, and it contributes to destruction of porosity
in two ways, by eliminating intergranular space at sites of dissolution and
by providing cement to fill pores elsewhere (e.g., Wong and Oldershaw
1981; Finkl and Wilkinson 1990). In addition to these petrologic effects,
pressure-dissolution surfaces (1) can generate bedding planes (Bat-
thurst 1987), (2) can account for significant thicknesses of lost section in strat-
igraphic reconstructions (e.g., Glover 1968; Mossop 1972), and (3) can be
of economic significance by modifying reservoir properties, by serving as
barriers to petroleum migration, and by acting as seals for reservoirs (e.g.,
Dunnington 1967).

Despite the petrologic, stratigraphic, and economic importance of pres-
sure dissolution, controls on the origin and morphology of pressure-di-
solution surfaces are not well understood. This paper examines lithologic
controls on morphology of styolites and dissolution seams in relatively
undeformed carbonate rocks. It thus addresses many of the same problems
on which Buxton and Sibbey (1981) focused, but it quantifies, rather than
categorizes, morphology of pressure-dissolution surfaces, and examines a
more general sample suite from the midwestern United States (Fig. 1).

Much previous work has shown that pressure dissolution varies with
gross lithology, as in Alvarez et al.'s (1978) comparison of marls vs. lime-
stones, Marshak and Engelder's (1985) comparison of pure limestones and
clay-rich limestones, and Tucker's (1991, p. 34) general observation that
styolites are more common in limestones than in sandstones. The present
study, however, focuses on only carbonate rocks and shows that thickness
and shape of styolites and dissolution seams correlate with individual
limestone fabric or lithology (as defined by Dunham 1962). Pressure-
dissolution surfaces also vary with abundance of particular carbonate lith-
ologic components, and their variability depends largely on the degree of
heterogeneity of their host carbonate fabrics.

WHAT IS A STYOLITE?

The meaning of "styolite" has evolved from its original use for the
supposed fossil Styloithes sulcatus (Klönne 1828), and important change
in its meaning has continued in recent years. In a much cited monograph,
Stockdale (1922) defined a styolite as "each individual, penetrating col-
mun. Thus . . . a styolite seam is made up of many styolites whose
direction of penetration, with few exceptions, is at right angles to it" (St-
ockdale 1922, p. 10–12; Stockdale 1943, his fig. 2, and similarly Wagner
1913, p. 16). Thus for Stockdale Figure 2A of this paper shows not one
but three or perhaps five complete styolites in a "styolite seam" or "sty-
olite surface".

In subsequent usage, "styolite" came to refer to the entire surface rather
than the individual column, and Conybeare's (1949) and Dunnington's
(1954) papers may represent the turning point. Bathurst (1975) defined a
styolite as "a complex interface between two rock masses, each mass
having a number of roughly columnar or finger-like extensions which fit
in between the opposed columnar extensions of the opposite mass." Thus
according to Bathurst's (1975) definition, Figures 2A and 2B would each
show one styolite and Figure 2C would show a debatable case, but Figure
2D would not show a styolite for lack of columns. Wanless (1979) main-
tained Bathurst's requirement of columns by writing that "styolites are
sutured (pillar and socket) interpenetrating surfaces".

A more general definition of "styolite", similar to that of Park and
Schot (1968), was provided by Buxton and Sibbey (1981), who stated that
"stylolites are serrated boundaries between [rock] units; the boundary usually has an accumulation of clay, oxides, and/or organic matter". Bathurst (1987) adapted Buxton and Sibley's definition not requiring columns in saying that "a stylolite is a serrated interface between two masses of rock; it has a sutured appearance in sections normal to the plane of the stylolite". Thus according to usage of carbonate petrologists in the 1980s, Figures 2A, 2B, 2C, and 2D would each display one stylolite. However, Groshong (1975, p. 1367), had already taken this trend a step further by defining a stylolite as "any zone of relatively insoluble residue against which the original fabric elements of the rock are truncated by removal rather than by offset"; and he specifically dropped the requirement of a "sutured" appearance. Thus by Groshong's (1975) definition, which was applied to a study of structural deformation and subsequently used by Geiser and Sansone (1981), Figures 2A, 2B, 2C, 2D, and 2E would each contain at least one stylolite.

Usage in the present paper follows the definitions of Buxton and Sibley (1981) and Bathurst (1987). "Dissolution seam" indicates an undulatory, nonserrate pressure-dissolution surface, also in keeping with Buxton and Sibley (1981) and Bathurst (1987).

**Fig. 2.**—Sketches of pressure-dissolution surfaces. A, B, C, and D are serrate surfaces, some or all of which are stylolites according to various definitions. E is an undulatory surface, which in one definition is a stylolite but in this paper is considered a dissolution seam. Scale is irrelevant.

**QUANTIFICATION OF PRESSURE DISSOLUTION SURFACES**

Many workers have proposed schemes to classify stylolites, but none have attempted a quantitative analysis of stylolite morphology. Trurnit (1968) and Park and Schot (1968) developed complex classifications of stylolites from theoretical perspectives, whereas Buxton and Sibley (1981), on the basis of field observations, divided stylolites into two types depending on the amplitude and frequency of vertical offsets. Powell (1979) classified cleavage surfaces and thus some stylolites quantitatively using the distance separating two or more surfaces, but he did not use quantitative parameters to characterize morphology of individual surfaces.

Any attempt at classification assumes that phenomena fall into discrete groups, but numerous workers have noted that pressure-dissolution surfaces are not easily categorized, except by scale. For example, Stockdale (1922) noted that druckstufen ("small, sharply intertoothed sutures") graded into "stylolite surfaces", and that contacts between pressure-dissolved pebbles have, under close examination, the morphology of stylolites. Similarly, Bathurst (1975, p. 468) observed that stylolites and sutured intergranular contacts differ only in scale, and Buxton and Sibley's (1981) classification was largely one of scale.

Because of the intergradational nature of pressure-dissolution surfaces and because of ambiguity in the definition of stylolites discussed above, quantification of morphology of pressure-dissolution surfaces is more effective than categorization. The pressure-dissolution surfaces collected for this study were quantified with four parameters (Fig. 3): T, thickness of the surface itself; I, amplitude of the irregularity of the surface (contrast with O below); O, amplitude of the (usually vertical) offset of the surface; and F, frequency of such offsets along the surface, in occurrences per unit distance. Frequency is thus the inverse of the average spacing of offsets.

In theory, T, I, O, and F can vary independently of each other (Fig. 2). Figure 2A shows a stylolite of large offset and thickness, Figure 2B shows one of large frequency, Figure 2C shows one of small offset and frequency, and Figure 2D shows one for which offset and frequency are zero. Druckstufen (see Wagner 1913) are stylolites of very low frequency and offset, whereas Buxton and Sibley's Type 1A stylolites have high offsets. This paper uses these four parameters, as well as the distinction between smooth,
undulatory surfaces and jagged, serrate surfaces (dissolution seams and stylolites as in Figure 2), to examine pressure dissolution in carbonates of diverse ages, locations, and depositional settings.

SAMPLES

Samples of pressure-dissolved Paleozoic carbonates were taken from quarries and fresh roadcuts in Tennessee, Kentucky, Indiana, Ohio, and Michigan (Table 1, Fig. 1). Ages of the samples range from Ordovician to Mississippian, and the samples come from the margins of the Illinois and Michigan Basins and the Nashville Dome and Cincinnati Arch. The samples thus span a considerable range of ages and settings, and so should allow general inferences about stylolites in undeformed carbonates independently of variation caused by possible changes in clay mineralogy and carbonate grains through time and space. Control samples of non-pressure-dissolved carbonates were also taken, but close examination in the laboratory revealed pressure-dissolution surfaces in many of these as well.

None of the samples have been buried deeply. Estimates of maximum burial depths are 1500 m in northern Michigan (Buxton and Sibley 1981), 800–1050 m in northern Indiana (Gray et al. 1987), 700 m in southern Indiana (Finkel and Wilkinson 1990), 700–1500 m in Kentucky (Shaver et al. 1984), and about 1200 m for Ordovician strata in Tennessee (Shaver et al. 1984).

As their shallow burial and cratonic setting would suggest, none of these limestones have been structurally deformed. Present dips are no more than 3°, none of the strata are folded or faulted in the vicinity of the sampling sites, and none of the limestones contain tectonically distorted fossils of the sort described by Engelder and Engelder (1977). Field observations by the author and by geologists he has consulted indicate that none of these limestones possess cleavage. None of these limestones contain recrystallized fibrous calcite along their stylolites, which Baghat and Marshak (1990) observed in solution cleavage in deformed limestones from eastern New York. Although some structural geologists consider stylolites a form of spaced disjunctive cleavage, stylolitic cleavage is a subset of disjunctive cleavage (Powell 1979, his fig. 3), and disjunctive cleavage is cleavage that cuts across bedding (Engelder and Marshak 1985; Borradale et al. 1982; Powell 1979). The stylolites examined for this project are all parallel to bedding, so they do not satisfy the definition of cleavage.

METHODS

Samples were sectioned perpendicular to pressure-dissolution surfaces, and the parameters introduced above were measured under a binocular microscope with 30x magnification. The value of O recorded was the average of the three largest offsets found, with care taken to not combine multiple offsets (like offsets 6 and 7 in Figure 3). F was measured across the most extensive lateral exposure of the stylolite. Point counts in grains used 350 to 800 points per thin section to determine abundances of (1) grains, intragranular cement, and intergranular porosity, (2) intergranular cement, and (3) intergranular porosity. Dolomite abundances were determined by visual estimation from thin sections. Insoluble-residue contents of selected samples were determined by dissolution of powdered subsamples in a buffer solution of acetic acid and sodium hydroxide with a pH of 5, prepared using a recipe in Long (1961). This solution is similar to that of Jackson (1979) but has a lower buffer capacity and was used at room temperature (see Railsback 1993).

The data presented here represent measurements from 169 pressure-dissolution surfaces from 52 samples from 16 localities. Statistical significance of trends or differences in the data are indicated by p, the probability that no correlation of observations or difference in means exists. Values of p were calculated from values of r (correlation coefficients) and n (numbers of observations) using methods in McClave and Dietrich (1985), and thus reveal more than r alone. Values of p less than 0.10 indicate statistical significance.

All observations reported here pertain to carbonate rocks, and so references to “lithology” refer to different carbonate lithologies. To characterize lithology, all limestones examined were categorized using Dunham’s (1962) classification, and use of the terms mudstone, wackestone, packstone, and grainstone below follows Dunham’s usage. “Carbonate mud” refers to fine-grained carbonate material. “Mudstone” does not imply that

![Fig. 3. — Sketch of a stylolite showing parameters used to quantify morphology of pressure-dissolution surfaces. T is thickness of stylolite itself, I is amplitude of irregularity of stylolite surface, O is amplitude of (usually vertical) offsets of stylolite, and F is frequency of such offsets (in direction parallel to stylolite surface). Scheme assumes that offsets are perpendicular to the trend of the stylolite surface; additional parameters could be added for more complex stylolites.](image-url)
any fine-grained siliciclastic material is present, and the results below show that the rocks examined in this study contain little such material.

RESULTS

Parametric Relationships

Among the samples described above, thickness $T$ of pressure-dissolution surfaces ranges from 0.01 mm, the minimum resolution with the binocular microscope, to 1.70 mm. The three parameters $O$, $I$, and $F$ all show statistically significant relationships with $T$. For example, $O$ (vertical offset) increases with $T$ in styloites, with $p < 0.01$ (Fig. 4). $I$ (amplitude of irregularity) likewise increases with $T$ ($p < 0.01$); this correlation is stronger for styloites than for all pressure-dissolution surfaces (Fig. 5). In contrast, $F$ (frequency of offset) is correlated with $1/T$ within data for styloites, with $p = 0.07$. (Fig. 6). Frequencies of more than 15.0 cm$^{-1}$ are found only in styloites with thicknesses less than 0.25 mm, whereas all styloites with thicknesses greater than 0.4 mm have frequencies less than 7.0 cm$^{-1}$.

Pressure-dissolution surfaces were categorized as undulatory or serrate, and lithologies or fabrics above and below each pressure-dissolution surface were determined according to Dunham’s(1962) classification of limestones. The results show that undulatory surfaces (dissolution seams) are most abundant in mudstones and progressively less so in wackestones and packstones, and there are none in the grainstones studied (Fig. 7). Thus, rocks richer in carbonate mud have greater relative abundance of dissolution seams. Note that not all undulatory surfaces in mudstones were measured, because their commonly anastomosing pattern provides a large number of surfaces to measure; only a representative sample of each “swarm” was measured. The proportion of undulatory pressure-dissolution surfaces in mudstones could thus be greater than shown in Figure 7, but could not be greater for wackestones, packstones, and grainstones.

$F$, $O$, and $T$ of styloites also vary with fabric and lithology. The mean value of $F$ for styloites in mudstones and wackestones is 2.15 cm$^{-1}$, whereas the mean value in packstones and grainstones is 6.26 cm$^{-1}$; the means differ with $p < 0.001$ (Fig. 8). Within these groupings, neither the
Fig. 8.—Histograms showing distribution of values of $F$ (frequency of vertical offset) for stylolites in mudstones and wackestones (top) and packstones and grainstones (bottom) in samples from midwestern U.S. Arrows indicate means for each group. Means for two groups differ with $p < 0.001$ (i.e., chance of less than 0.1% that means of two populations do not differ). See Figure 3 for definition of $F$.

Fig. 9.—Histograms showing distribution of values of $O$ (amplitude of vertical offset) for stylolites in mudstones and wackestones (top) and packstones and grainstones (bottom) in samples from midwestern U.S. Arrows indicate means for each group. Means for two groups differ with $p < 0.001$. See Figure 3 for definition of $O$.

Fig. 10.—Plot of values of $T$ (thickness) for different Danum lithologies in samples from midwestern U.S. Lines connect means for each group. See Figure 3 for definition of $T$.

The difference in mean values of $F$ between mudstones and wackestones is statistically significant. Similarly, the mean value of $O$ for mudstones and wackestones is 1.61 mm, whereas the mean value in packstones and grainstones is 6.73 mm, and the means again differ with $p < 0.001$ (Fig. 9). Within these groupings, the difference in mean values of $O$ for mudstones and wackestones is significant ($p < 0.05$), but not between packstones and grainstones. Statistical significance ($p = 0.001$) also exists in the difference between the mean value of $T$ in mudstones and wackestones (0.13 mm) and that in packstones and grainstones (0.24 mm). However, the mean value of $T$ is significantly larger in mudstones than in wackestones, and significantly larger in packstones than in grainstones (Fig. 10). The mean value of $T$ in all limestones examined is 0.26 mm, but the mean value of $T$ in dolostones is 0.13 mm, and the means differ significantly with $p = 0.01$ (Fig. 11—Histograms showing distribution of values of $T$ (thickness) in dolostones and limestones in samples from midwestern U.S. Dolostones include all rocks with $> 95\%$ dolomite, and limestones are all rocks with $\leq 5\%$ dolomite. Arrows indicate means for each group. Means for two groups differ with $p = 0.01$. See Figure 3 for definition of $T$. 

11). The mean value of $O$ in limestones (3.86 mm) is likewise greater than in dolostones (mean = 1.03 mm), but with a lesser statistical significance ($p = 0.04$).

Pressure-dissolution surfaces show little relationship to lithologic contrasts. Surfaces were categorized as occurring between rock units with no lithologic or fabric change, or between rocks in which Dunham lithology, grain size, grain packing, or some other petrographic characteristic differed. Only 26% of all pressure-dissolution surfaces coincide with lithologic changes. Mean values of $T$ do not differ significantly between the two groups, either for all surfaces or for stylolites ($p = 0.23$ and $p = 0.15$, respectively). Similarly, mean values of $F$ do not differ significantly for the two groups (for stylolites, $p = 0.36$). Mean values of $O$, however, differ significantly: the mean value of $O$ for stylolites at lithologic contrasts is 4.88 mm, whereas the mean value is 2.02 mm for stylolites between identical lithologies ($p = 0.02$) (Fig. 12).

The limestones examined in this study contain relatively little insoluble material outside their pressure-dissolution surfaces. Dissolution of some of the most visibly argillaceous or impure limestones yielded insoluble residues that constituted 0.5 to 1.0% of the bulk sample by weight (Table 2).

**Stylolites and Cement in Grainstones**

$T$ and $O$ do not correlate significantly with abundance of intergranular cement in grainstones ($p \approx 0.17$ and $p = 0.5$, respectively). Figure 13 shows, however, that $O$ does not exceed a line defined by $T = 0.05C - 0.35$, and $O$ does not exceed $O = 2.3C - 25.0$, where $C$ represents abundance of cement. Samples that approach these limits are not easily distinguished from samples with less well developed stylolites; for example, there is no systematic difference in types of grains or cements, or in abundance of intergranular porosity. In fact, some stylolites approaching these limits are less than 10 cm from stylolites with much lower values of $O$ and $T$ (Fig. 13).

Abundances of petrographic components at ends of columns or “teeth” of stylolites are not random. In all grainstones examined, grains are more abundant than penetrative components of stylolite columns than through the entire rock (Fig. 14), and cements are less abundant than penetrative components than in the rest of the rock. Among the penetrating components, echinoderm grains are most abundant (Table 3). Ratios of occurrences as penetrating elements to occurrences as elements penetrated by stylolite projections show that noncarbonate grains are consistently the favored component, that carbonate grains other than bryoaceous are generally favored, and that calcite cements and bryoaceous are disfavored to survive pressure dissolution (Table 3).

**Dissolution Seams**

Among the samples examined, dissolution seams are present only in rocks with more than 20% dolomite, and only in rocks with dolomite crystal sizes less than 0.20 mm (Fig. 15). This contrasts with the distribution of stylolites, which are present in rocks containing 0 to 100% dolomite, and in rocks with a broad range of dolomite sizes. Dolomite

![Graph showing distribution of values of $O$ (amplitude of vertical offset) for stylolites between identical lithologies (top) and between rocks differing in lithology, grain size, or packing (bottom) in samples from southeastern U.S. Arrows indicate means for each group. Means for two groups differ with $p = 0.02$. See Figure 3 for definition of $O$.](image)

![Graph showing plots of $T$ (thickness of stylolites) vs. abundance of intergranular cement in grainstones. Plots of $O$ (vertical offset of stylolites) vs. abundance of intergranular cement in grainstones. Dashed lines connect stylolites less than 10 cm apart.](image)

**Table 2. Results of analyses of insoluble residues in host rocks**

<table>
<thead>
<tr>
<th>Sample Locality</th>
<th>Sample Lithology</th>
<th>Parameters of Nearest Stylolite</th>
<th>Weight % Insoluble Residue</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>$T$ (mm)</td>
<td>$I$ (mm)</td>
</tr>
<tr>
<td>6</td>
<td>Mudstone</td>
<td>0.06</td>
<td>0.7</td>
</tr>
<tr>
<td>16</td>
<td>Grainstone</td>
<td>0.23</td>
<td>1.2</td>
</tr>
<tr>
<td>34</td>
<td>Packstone</td>
<td>0.01</td>
<td>0.2</td>
</tr>
<tr>
<td>70</td>
<td>Grainstone</td>
<td>0.28</td>
<td>1.0</td>
</tr>
</tbody>
</table>
rhombs penetrate other rhombs and grains, indicating that dolomite preceeds development of dissolution seams in these rocks. Both $T$ and $F$ show weak negative correlations with dolomite abundance (Fig. 16), but no significant relationship exists between $T$ or $F$ and dolomite crystal size.

**DISCUSSION**

**Thickness. Offset, and Frequency, and their Relation to Carbonate Lithology**

Covariance of $T$ and $O$ presumably reflects accumulation of insoluble materials at a stylolite as it develops and removes more carbonate material to generate greater offset. Covariance of $T$ and $I$ similarly supports the notion that pressure-dissolution surfaces typically develop from thin, relatively flat surfaces, and thus larger stylolites are just older or more extensively developed stylolites. The inverse relationship between $T$ and $F$ may reflect the increased strength of a thicker mass of insoluble material and thus the lesser propensity of the surface to break into segments capping separate stylolite columns. The inverse relationship between $T$ and $F$ suggests that offsets or columns must begin to form early in stylolite development, before the surface becomes thick enough to preclude formation of offsets. Once offsets have begun to form, however, $O$ and $T$ increase together.

The differences found between grain-supported carbonate rocks (grainstones and packstones) and mud-supported carbonate rocks (wackestones and mudstones) suggest that variability of carbonate fabric controls stylolite morphology (Figs. 7–10). Particles of carbonate mud, which dominate in mudstones and wackestones, may be so small (relative to insoluble particles) and uniform in susceptibility to pressure dissolution that these rocks lack contrasts along pressure-dissolution surfaces sufficient to generate offsets. This would account for both lower values of $F$ and $O$ in mudstones and wackestones and greater abundance in these rocks of dissolution seams, in which $F$ and $O$ approach zero. On the other hand, grain-supported rocks show greater variability in susceptibility to pressure dissolution (Table 3), and their particles are much larger than the insoluble materials making up the surface that must be broken as offsets form. Greater values of $F$ in grain-supported rocks may also result from lower concentrations of insoluble minerals, especially clays, in winnowed rocks so that $T$ does not increase rapidly during stylolite development to preclude development of offsets.

Buxton and Sibley (1981) concluded that stylolites form preferentially at changes in carbonate (Dunham) lithology. The data presented here show a lower percentage of pressure-dissolution features at lithologic contacts than found by Buxton and Sibley and so do not support their claim. The greater mean value of $O$ for stylolites at changes in carbonate lithology (Fig. 12) suggests instead that such stylolites are more easily recognized. However, this study was not designed to test Buxton and Sibley’s hypothesis; that would have required tabulation of both lithologic changes and pressure-dissolution surfaces in continuous vertical sections.

**Stylolites and Cementation**

Covariance of maximum values of $O$ and $T$ with abundance of cement in grainstones (Fig. 13) agrees with the generalization that stylolites are more likely to develop in more extensively cemented limestones (Garrison and Kennedy 1977; Buxton and Sibley 1981; Railback 1993) or in “limestone units having structural resistance to stress” (Wanless 1979). Two interpretations are commonly made: (1) that stylolites generate dissolved

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**Table 3.**—Occurrence of rock components as penetrating elements leading columns or teeth of stylolites in grainstones

<table>
<thead>
<tr>
<th>Components</th>
<th>Occurrences as Penetrating Components</th>
<th>Penetration-to-generated Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dolomite Rhombs</td>
<td>7</td>
<td>∞</td>
</tr>
<tr>
<td>Quartz Grains</td>
<td>4</td>
<td>∞</td>
</tr>
<tr>
<td>Phosphate Grains</td>
<td>2</td>
<td>∞</td>
</tr>
<tr>
<td>Belemnoids</td>
<td>22</td>
<td>11.0</td>
</tr>
<tr>
<td>Echinoderms</td>
<td>109</td>
<td>3.5</td>
</tr>
<tr>
<td>Muc. Calcite Grains</td>
<td>80</td>
<td>3.3</td>
</tr>
<tr>
<td>Peloids</td>
<td>14</td>
<td>2.3</td>
</tr>
<tr>
<td>Ooids</td>
<td>33</td>
<td>1.1</td>
</tr>
<tr>
<td>Calcite Cements</td>
<td>18</td>
<td>0.9</td>
</tr>
<tr>
<td>Bryozoans</td>
<td>3</td>
<td>0.3</td>
</tr>
<tr>
<td>Multiple Calcite Components</td>
<td>35</td>
<td>0.2</td>
</tr>
</tbody>
</table>

1 Ratio of occurrences as penetrating element to occurrences where penetrates by a penetrating element.

2 Grains plus cement.
solids that form cements nearby, or (2) that well cemented rocks form rigid frameworks that allow development of macroscopic stylolites, whereas less cemented rocks undergo intergranular pressure dissolution. Although pressure dissolution clearly generates material for cement, Buxton and Sibley (1981, p. 24) questioned the first explanation by pointing out that both porosity near stylolites and lack of cement in grainstones that have undergone intergranular pressure dissolution demonstrate that cement does not always precipitate nearby. The second explanation accounts for the lack of intergranular compaction in cement-rich rocks, but it fails to explain why those rocks, having achieved structural uniformity, do not form planar pressure-dissolution surfaces rather than the stylolites with high values of $T$, $O$, and $F$ commonly observed.

A third interpretation of the relationship between cementation and stylolitization arises from Figure 14 and Table 3, which show that formation of stylolite columns involves preferential penetration of grains at the expense of cements (and bryozoan fragments). Interaction of less pressure-soluble components (grains and non-CaCO$_3$ components) with more pressure-soluble components apparently favors stylolite development. This perspective suggests that stylolite development in cement-rich grainstones does not necessarily result from a stable lithologic framework but may also be caused by abundance of pressure-soluble cement. This interpretation not only accounts for the data in Figures 13 and 14 but also explains Buxton and Sibley’s (1981) observation that poorly cemented grainstones do not develop stylolites, even after intergranular pressure dissolution has generated a stable lithologic framework of pressure-welded grains. Reasons for the seemingly greater pressure solubility of cements relative to many grains are not readily apparent but may lie in Meike’s (1983) TEM observations that pressure dissolution acts at microfractures and dislocations, which may be unequally distributed among limestone components. The contention that rocks rich in pressure-resistant components are less likely to contain well-developed pressure-dissolution surfaces is also supported by data on dissolution seams and dolomite abundance, which are shown in Figure 16 and discussed below.

**Dissolution Seams and Dolomite**

Association of dissolution seams with dolomite (Fig. 15) has been noted by Wannless (1979), but among the carbonates examined in this study, dissolution seams are present only in dolomitic limestones and dolostones of fine crystal size. Wannless (1979) argued that dolomitization is a response to the same stresses that cause pressure dissolution, but petrographic evidence shows that dolomite predates pressure dissolution in these dissolution seams. Furthermore, dissolution seams in carbonates with dolomite abundances approaching 100% present the problem of how clays in dissolution seams, or any other internal source, could expel sufficient Mg to cause volumetrically significant dolomitization (Morrow 1982, Land 1982).

An alternative explanation of association of dissolution seams with fine-grained dolomite is that dolomite generates a rock fabric favoring development of dissolution seams. Table 3 suggests that dolomite has a lower susceptibility to pressure dissolution than calcite, and Trumit’s (1968) ranking of minerals by pressure solubility placed calcite before dolomite. This relationship suggests that dolomitic rocks contain enough pressure-resistant elements to inhibit the interpenetration characteristic of stylolites and may instead allow only development of dissolution seams. Trends within dissolution seams support this hypothesis: Figure 16 shows that development of dissolution seams decreases with increasing dolomite abundance. The presence of stylolites in dolomite-rich rocks is a potential rebuff of this hypothesis, but lower mean values of $O$ and $T$ in dolostones (Fig. 11) support the contention that abundance of pressure-resistant dolomite inhibits stylolite development. Glover (1968) also noted this inhibition in Ordovician carbonates from Western Australia, where stylolites were found in limestones and dolomitic limestones but not in dolostones.

**Development of Pressure-Dissolution Surfaces**

Combination of data regarding different Dunham lithologies (Figs. 7-10) and abundance of specific components (Figs. 13-16) suggests that development and elaboration of pressure-dissolution surfaces depend on superposition or interaction of materials of contrasting pressure solubility. Rocks rich in just one component, whether it is carbonate mud (as in mudstones and wackestones), carbonate grains (as in cement-poor grainstones), or dolomite, all have dissolution surfaces lower in $T$, $O$, and $F$ than rocks in which grains and mud (packstones) or cement and grains (cement-rich grainstones) or calcite and dolomite are abundantly juxtaposed. Stylolite growth may require a stable rock framework, but the results of this study suggest that contrasting pressure solubility at the component-to-component level is necessary as well.

The results of this study should allow better prediction of (1) which strata undergo pressure dissolution and (2) the extent to which pressure-dissolution surfaces develop. Such insight should in turn allow better prediction of destruction of porosity by pressure dissolution, and better prediction of surfaces sufficiently extensive to control migration and location of hydrocarbons. It should also allow better estimation of lost section, a matter of great importance to modelers of sedimentary sequences. Furthermore, carbonate lithologies (trend Dunham 1962) have differing susceptibilities to pressure dissolution, so that some lithologies (packstones and well cemented grainstones) may be underrepresented in the strati-
graphic record. The mean of value of $T$ for packstones and grainstones is twice that for mudstones and wackestones, suggesting a twofold loss of section among the former, but lesser concentration of clays and other noncarbonate phases in grain-supported rocks may make this an underestimate. Other workers have used maximum amplitude of stylolites as a measure of lost section, and the mean value of $O$ for grain-supported limestones is four times that in mud-supported rocks, suggesting a fourfold preferential loss of section in packstones and grainstones.

Controls on Pressure Dissolution Not Related to Carbonate Fabrics

Insoluble-residue contents of $\pm 1\%$ in the carbonates examined illustrate that development of pressure-dissolution surfaces does not require large concentrations of noncarbonate material. Marshak and Engelder (1985) reported that, in the Devonian of eastern New York, only limestones with clay-quartz contents of more than $10\%$ developed widespread stylolitic cleavage, whereas limestones with lesser insoluble contents contained isolated stylolites. The work of Marshak and Engelder work focused on deformed limestones with cleavage in a very different tectonic setting, so their study is not directly comparable with this one, but the results here do not suggest that pressure dissolution requires high insoluble-residue contents.

One possible explanation for the development of pressure-dissolution surfaces is the relatively clay-poor undeformed rocks studied: these lie in their shallow burial. Extrapolation from Engelder's (1984) work in other settings suggests that shallow burial may have allowed extensive circulation of dilute groundwaters, thus promoting pressure dissolution. Engelder (1984) pointed out that pore waters above 1 to 2 km depth are typically relatively dilute, whereas deeper pore fluids tend to have higher concentrations of dissolved solids. Comparison of deformation mechanism maps (Rutter 1976; Meikle 1990) for wet and dry conditions amplitudes this point, because the "wet" deformation mechanism maps contain a larger field for pressure dissolution, whereas the "dry" maps have no field for pressure dissolution. Because the rocks examined here were never buried more than about 1.5 km, and may have often been much shallower, they have probably always contained dilute pore fluids that promoted pressure dissolution, both with their low concentrations to facilitate dissolution and with their active circulation to remove the dissolved products.

If shallow hydrologic activity is one key to enhanced pressure dissolution, it would explain the many observations of extensive pressure dissolution and stylolitization during shallow burial (e.g., Buxton and Sibley 1981; Finkel and Wilkinson 1990; Raisback 1993). Dependence of extensive pressure dissolution on shallow dilute pore waters would also undermine the usual characterization of pressure dissolution as a deep-burial diagenetic process (e.g., Bathurst 1980; Scholle and Halley 1985; Soffin 1987, p. 114), which may have arisen from many workers' focus on pressure. The perspectives of Engelder (1984) and Meikle (1990) suggest that "dissolution" may be as important as "pressure" in pressure dissolution.

Finally, one should note the limitations of this study. Only carbonate rocks from stable cratonic regions were sampled and studied, so the results do not pertain to structurally deformed rocks in which stylolites form cleavage surfaces (as studied by Alvarez et al. 1976, 1978, Marshak and Engelder 1985, Engelder and Marshak 1985, and de Roo et al. 1992). Also, although the results reveal controls on morphology of pressure-dissolution surfaces, they account only partly for why pressure dissolution occurs in some settings and not others. Catalysis of pressure dissolution at any depth by phases such as clays (Weyl 1959; Schwander et al. 1981; Marshak and Engelder 1985) and small-scale mechanisms of pressure dissolution (e.g., Meikle and Wenk 1988; Meikle 1990) persist as problems independent of controls on morphology and growth of pressure-dissolution surfaces, and answers may rest in both the noncarbonate phases present during compaction as well as the fabric or lithology of the carbonate itself.

CONCLUSIONS

Examination of pressure-dissolution surfaces in relatively undeformed carbonate rocks from the midwestern U.S. supports the following conclusions:

1) Pressure-dissolution surfaces can be usefully described using quantitative measurements ($T$, $I$, $O$, and $F$ in Figure 3), as opposed to previous qualitative categorization.

2) $O$ and $I$ increase with $T$ in stylolites, reflecting elaboration of stylolites as they grow; $F$, in contrast, is inversely proportional to $T$, suggesting that the physical strength of the stylolite may inhibit development of offsets.

3) Dissolution seams are more common in rocks richer in carbonate mud, whereas stylolites increase in abundance in the transition from mudstones to grainstones, and $T$, $F$, and $O$ of stylolites are greater in grain-supported rocks than in mud-supported rocks. These results suggest that elaboration of pressure-dissolution surfaces from seams to stylolites high in offset and frequency results from increased heterogeneity of rock fabric.

4) Maximum values of $T$ and $O$ increase with abundance of cement in grainstones. These data, combined with data showing that many grains are less susceptible to pressure dissolution than cements, suggest that stylolite development in well cemented rocks results from juxtaposition of grains and cement, whereas abundance of pressure-resistant grains in cement-poor grainstones inhibits stylolite development.

5) Within the samples studied, dissolution seams are found only in rocks containing fine-grained dolomite. This co-occurrence, combined with an inverse relationship between dolomite abundance and seam development, suggests that presence of fine-grained, pressure-resistant dolomite inhibits elaboration of pressure-dissolution surfaces.

6) Points 3, 4, and 5 suggest that development of pressure-dissolution surfaces into stylolites high in $T$, $O$, and $F$ generally requires a rock with an abundance of both pressure-soluble and pressure-resistant components. Rocks that contain only one component, such as carbonate mud (in mudstones), carbonate grains (in cement-poor grainstones), or dolomite, all generally have dissolution surfaces with lower values of $T$, $O$, and $F$ than rocks containing mixtures of grains and mud (packstones) or cement and grains (cement-rich grainstones) or calcite and dolomite. Rocks of the latter kind may thus be underrepresented in the stratigraphic record because of their loss to pressure dissolution.

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