



Stress Sensitivity of Carbonate Rocks

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Abstract

Carbonate rock characterisation has been a challenging task in reservoir studies due to their complex pore structure and various diagenetic evolutions. Proper understanding of physical and chemical characteristics of rocks facilitate more accurate analysis. As a matter of fact, understanding the evolutions of sediments is a key to characterize the current state of the rock. One of the essential rock properties is porosity, which is typically a naive function of depth (overburden) in most sedimentary rocks, known as compaction trend. However, porosity in carbonates doesn't necessarily follow a uniform compaction trend, because of various physical and chemical processes that take place hand in hand with the physical rock deformation due to stress. It is imperative to assess the stress-sensitivity of carbonate rocks to be able to validate any analysis that relies on compaction. The conclusion is that density and porosity doesn't necessarily correlate with depth or overburden stress, making carbonate rocks much less stress-sensitive than other sedimentary rocks. In this paper, a holistic review of literature is conducted to document the deposition and sedimentation of carbonate rocks, then explain the diagenesis and various forms it takes. Then, stress-sensitivity of carbonate rocks is examined, and implications are discussed.

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The sedimentation regime of carbonate rocks is a combination of mechanical and chemical compaction that varies with burial depth, temperature, mineral composition, and type of pore fluid. Porosity, and some other rock properties, is modified in three environments known as diagenesis regimes [1,2].

1. Marine environment: characterised by supersaturated sediments with normal or modified pore fluid [3].
2. Meteoric environment: occurs in the vadose zone associated with precipitation of calcite (needle fibres, meniscus cements, and micritic networks), neomorphism and dissolution [4]. This zone is characterised by fresh water or very low salt saturation with respect to most carbonate mineral species [2].
3. Subsurface environment: where the pore fluid is either marine or meteoric water [1,5] or there is a mixture of water-rock interactions [6,7].

In each of the above mentioned environments, different types of burial process takes place. The first stage of burial in shallower depths is mainly dominated by a mechanical compaction and results in de-watering. First, a set of grain rearrangements takes place to reach the most stable packing geometry [2]. This phase is generally associated with rock fracturing or plastic deformation [8,9] and since allochems in unlithified sediments are in a mud context, the deformation is less homogenous because the load is applied on both the mud and grain. The force on the grain surface would vary in different parts of the deposit depending on the grain/mud proportion, mud strength and magnitude of the load. To put it simply, compaction in post-lithified sediments exhibits more homogeneity as the load is uniformly applied on the rigid substance. Therefore, co-existence of deformed and undeformed allochems is essential for distinguishing the relative time of compaction with respect to lithification[9].

It is important to notice that mechanical compaction would be different in grain-dominated deposits (such as carbonate sands) and

mud-dominated sediments [2] and therefore, clay plays a critical role in controlling a variety of mechanical and chemical responses to the applied load. Clay in the rock fabric could enhance force diffusion or, at the same time, impede cementation [10].

Mechanical compaction influenced by the mineralogy of the sediments (controlled by depositional environment) creates the rock frame with the primary porosity. After that, several processes are applied to the carbonate rocks which modify the existing porosity, either by creating more pore space or reducing pore space by filling them. Additional porosity created by post-depositional events is known as secondary porosity and could form from either of the following processes:

1. Dissolution
2. Dolomitization
3. Brecciation
4. Fracturing

Similar to this classification, Lucia [11] grouped diagenetic processes according to their conformance to depositional patterns in order to study the petrophysical rock fabric:

1. Cementation, compaction, and selective dissolution
2. Reflux dolomitization and evaporite mineralization
3. Massive dissolution, collapse brecciation, and fracturing

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The dissolution process is occasionally followed by mineral precipitation and the resultant porosity is discussed herein under the title of chemical compaction.

Chemical Compaction

Once carbonate sediments become stiffer and lithified, increasing load results in greater elastic strain at the individual grain contact points. The increased strain leads to a high potential for chemical reactions, such as solubility at grain contact points [2]. Under such conditions, as a crystal is immersed in its saturated solution, the contact points of the grains under stress may be dissolved [12]. This phenomenon is known as the ‘pressure solution’ which is attributed to the dissolution of minerals under non-hydrostatic stress [13]. Indeed, the original mineralogy of the sediment and the pore-filling fluid properties are key controlling factors over the whole process of ‘chemical compaction’, which involves one or more cycles of pressure solution and cementation [2]. As a result, the dissolution of rock mass into the pore fluid creates several textural features including: microstylolite, stylolite, wispy seam, and solution seam [14]. A numerical model for rock deformation and mass transfer by pressure solution is provided by Rutter and Elliott [15] as a function of pressure and temperature. Meanwhile, Sibley and Blatt [10] observed less pressure solution in the well-cemented samples of the Tuscarora Orthoquartzite samples than was seen in more friable samples, but with more cemented samples also displaying greater amounts of pore fill mineralisation. They pointed to four factors that control the amount of pressure solution: proximity to folds; grain size; sorting; and clay content.

In carbonate sequences, massive dissolution can be either carried out by near surface water [11] or deep burial fluid [16], and leads to substantial secondary porosity. Several major oil and gas reservoirs in the Middle East, such as the Ghawar supergiant field in Saudi Arabia, are enhanced by this type of diagenetic process [17,18]. Pressure solution also creates stylolites, which can be filled by several chemicals and minerals such as dead oil, sparry calcite, pyrite, black organic matter and rare dark grey argillaceous matter. Aharonov and Katsman [19] suggested that stylolites propagate only when both the pressure solution and clay-enhanced dissolution take place in unison. A list of key parameters influencing pressure-solution/recrystallization reactions in carbonate rocks is provided in Table 1.

Dissolution can be accelerated by high stress at grain-to-grain contact points, and the dissolved particles may precipitate in the adjacent pore space where the regional stress is lower. Precipitation is found in the form of ‘calcite cement’, which is composed of calcite,

high-magnesium calcite, or aragonite [11]. The porosity loss by cementation is equal to the proportion of total inter-granular cement to total pore space [21]. After, or during, matrix texture precipitation, grains could undergo partial leaching which creates porosity. On some occasions during meteoric diagenesis, grains are leached out completely by selective dissolution while the matrix remains unaltered. In such cases, the main porosity is ex-grain, and the rock framework consists of phreatic cement.

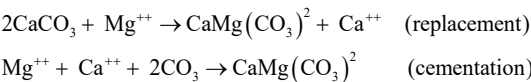
In this paper, the definition of cement by Bathurst [12] is followed to include all passively precipitated and space-filling carbonate crystals that grow attached to a free surface. Nevertheless, it is worth noting that there are three main requirements for calcium and carbonate ions to undergo continuous cementation [11]:

- 1. Grain dissolution associated with chemical compaction.
- 2. Dissolution of unstable minerals, such as aragonite.
- 3. Long distance transport of ions by ground water flow.

As another control on chemical compaction, thermal degradation of hydrocarbons in carbonate environments at temperatures higher than 150°C generate CO₂, H₂S, methane and solid pyrobitumen [22]. Although the combination of CO₂ and H₂S in an aqueous environment could accelerate dissolution and enhance porosity, solid pyrobitumen precipitation can significantly degrade formation porosity [2]. Therefore, cementation will not proceed in pores filled by hydrocarbon, and any further dissolution (without cementation and in the absence of compaction) may create additional secondary porosity.

Dolomitization

Hypersaline evaporated sea water accumulated in hypersaline tidal flat environments and associated hypersaline ponds, lakes, and lagoons could reflux from the surface down into underlying strata, replacing seawater and interacting with ground-water [11]. Sea water flows down towards the sea by a hydrodynamic force due to the higher density of sea water and the higher elevation of tidal flats. As a result of hypersaline sea water and groundwater interaction, dolomites are produced, and gypsum and anhydrites are precipitated. Chemical reactions forming dolomite are in two forms of replacement and cementation [23].

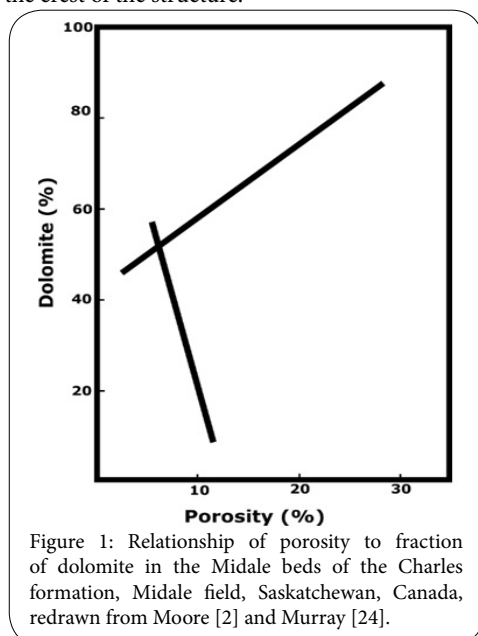


The role of magnesium, and the need for fluid flow to introduce the magnesium to the system, is obvious from the above chemical reactions. Murray [24] provided an interesting relationship between dolomite content and porosity in the Midale beds (lime muds) of the Charles Formation, Midale field, in Saskatchewan, Canada. Floating dolomite rhombs occupy available pore spaces as long as the mud consists of up to 50% dolomite. From this point, dolomite rhombs provide a rigid framework for the sediment and prevent further compaction. Therefore, porosity increases with increasing dolomite percentage (Figure 1). Lucia [23] reported a porosity increase by dolomitization in mud-dominated carbonates, but with no significant effect on pore size distribution in grainstones. Saller and Henderson [25] observed a varying effect of dolomitization on the porosity in the Permian platform dolomites of west Texas. Saller and Henderson [25] also investigated the role of differential compaction in the

| Porosity (permeability) | | |
|-------------------------|-----------------------------------------------------------------------------|--------------------------------------------------------------------------------------------|
| low porosity | high porosity | |
| transport control | surface area | |
| | low | high |
| | rate of dissolution control; stress concentration is an important predictor | rate of dissolution and crystal growth, about equal magnitude |
| | | inhibition is important; low super-saturation (high silicates content cause inhibition) |

Table 1- Mechanisms and controls of pressure-solution/recrystallization reactions in carbonate rocks [20].

formation of platform-margin dolomites over structural highs, and platform-interior dolomites in down-dip and basin-ward structures. Differential compaction provided the best reservoir quality on the basin-ward flanks of the structure, while the reservoir quality was the worst on the crest of the structure.



There is another process that occurs in dolomites, similar to differential compaction, that is known as 'evaporite mineralization' [11]. The controlling factors in this reaction are temperature and water activity. Hardie [26] provided a relationship between these parameters, which showed a reasonable compatibility to the recent supertidal flat sediments of the Trucial Coast in the Persian Gulf.

Lucia [11] classified anhydrite and gypsum textures in carbonate rocks, and their effect on the formation porosity and permeability, as follows.

1. Poikilotopic anhydrite crystals are formed by a combination of replacement and pore-filling mechanisms that reduce porosity. Since poikilotopic anhydrite is typically scattered across the rock, the matrix between crystals remains constant and permeability is not affected while pore size remains constant.
2. Nodular anhydrite is a form of diagenetic texture that does not reflect the depositional environment conditions. It is formed within the sediments by displacement as either anhydrite or gypsum in the shape of microcrystalline masses. While it locally reduces porosity, total rock porosity and permeability does not change as these nodules only make up a small percentage of the bulk volume.
3. Pore-filling anhydrites are pervasive, evenly distributed in the rock, and occlude the inter-grain, inter-crystal, and vuggy pore space. They have a destructive impact on porosity and permeability.
4. Bedded anhydrite is deposited out of hypersaline water as gypsum, which converts to anhydrite later in the form of laminated or coalesced nodules. It is formed either by precipitation out of water or by displacement and replacement of near surface sediments as gypsum or anhydrite. This type of anhydrite is laterally continuous and provides an impermeable barrier and seal for reservoirs.

Brecciation

Several conditions in carbonate rocks lead to massive dissolution that can cause large scale collapse and brecciation [11]. This phenomenon happens in situations such as limestone solution collapse, evaporite solution collapse, faulting and soil formation [27]. Four types of brecciation were classified by Kerans [28] as: fracture breccia, mosaic breccia, chaotic breccia (siliclastic-matrix supported), and chaotic breccia (carbonate-clast supported). In a typical burial evolution of cave-related breccia, several facies may be formed during massive dissolution along with several mineralisation processes. The additional pore space created in such conditions provides improved reservoir quality, and can host hydrocarbons under some circumstances, such as those of the Mississippian Northwest Lisbon Field, Utah [29] and the Permian Yates Field [30]. Additional evidence of karstic limestone collapse associated with unconformities was provided by Kerans [31] in west Texas, which exhibits improved reservoir quality.

Fracturing

A fracture is a discontinuity or parting in material caused by brittle failure [32]. According to this definition, joints and faults are types of naturally occurring fractures. Fractures are the deformation strain of a rock body against stress, and can form from many processes. A natural fracture system can play a significant role in reservoir performance during primary, secondary, and tertiary recovery. Despite the low intrinsic porosity and permeability of carbonate grain and matrix, a remarkable amount of hydrocarbon is produced from fractured carbonate formations. As an example, the Asmari limestone (Oligocene-Miocene) of the Gachsaran Field in Iran has 25% porosity and greater than 100 md permeability, even though the matrix porosity is only 5-9% [33,34]. Fractures are particularly common in carbonate rocks due to their brittle nature when compared with fine-grained siliclastics. Since naturally occurring fractures originate as a result of various mechanisms, fracture patterns vary in their attributes and are primarily related to stress conditions at the time of fracturing. Though, some fracture systems tend to be simplified when the paleo-stratigraphical model is constructed [35].

Three common driving forces that create fractures are tectonic stresses, pore fluid overpressure, and the formation of large voids [11]. Tectonic stresses are explained in terms of stress analysis [36], and are typically considered to result in three different types of fractures: shear fractures, extensional fractures, and tensile fractures. On the other hand, Nelson [37], classified naturally occurring fractures in four types, based on the nature of causing effects.

1. Tectonic fractures (due to surface forces)
 - Fault-related fractures
 - Fold-related fractures
 - Other tectonic fractures (such as piercement and impact structures).
2. Regional fractures (due to surface forces or body forces), a result of:
 - desiccation
 - syneresis
 - thermal gradients
 - mineral phase changes.

3. Contractional fractures (due to body forces).
4. Surface-related fractures (due to body forces).

According to this classification, tectonic fractures are those created, or attributed to, tectonic events and are associated with the folds and faults. Nelson [37] stated that the majority of tectonic fractures in outcrops tend to be shear fractures. Regional fractures are those that are developed over large areas of the earth's crust with relatively little change in orientation. Desiccation fractures are like mud cracks, which result due to shrinkage upon loss of water in subaerial drying. Syneresis fractures are the result of chemical processes during subaqueous or subsurface dewatering which cause volume reduction. Thermal contractional fractures are created when a hot rock cools due to different thermal gradients [38]. Mineral phase change fractures are those created due to the mineral phase change in carbonate or clay constituents where the substances have different molar volumes before and after a mineral transformation [39]. Under proper conditions, mineral phase changes can create a 'chicken wire pattern' of fractures. Contractional fractures can be formed due to tension or extension, and result in a general bulk volume reduction in the rock. Finally, surface-related fractures are the result of body forces during unloading [40]. The most common feature of this type is observed in quarrying operations when a piece of rock body is removed from the quarry. The rock relaxes due to reduced surface load, and spalls or fractures on a plane parallel to the newly developed free space. Extensive fracturing creates a brecciation pattern in rocks. Indeed, the characteristics of fold-related fractures, such as opening, propagation and distribution, are generally predictable from numerical modelling of the bedding curvature [41-44].

Stress Sensitivity of Carbonates

The compaction of carbonate rocks is a more complex process than in siliciclastic deposits, because post-deposition petrophysical attributes of carbonate rocks are not solely controlled by mechanical factors. Compaction in carbonate rocks is often not clearly related to effective vertical stress, as several diagenetic processes create or destroy porosity after burial and, therefore, porosity-depth relationships cannot be defined as a function of overburden stress. The prior review of carbonates and their classification schemes highlights that it is critical to know the conditions under which a rock was deposited in order to develop any understanding of a carbonate's mechanical attributes. The composition of carbonate rocks can reflect depositional environment because of two reasons: lack of transport in carbonate regimes and the direct tie to the biological components of the environment [2]. Due to the vast variety of organic and chemical parameters in the marine environment, where the majority of carbonate rocks are deposited [45-47], a wide variety of petrophysical properties are observed. Although there is strong evidence that carbonate porosity decreases exponentially with depth [21, 48], predicting compaction trend as a function of depth or overburden stress is still generally an invalid approach. Furthermore, it is also noted that carbonates are not the only rocks to often exhibit a lack of any robust porosity-depth relationship, as this is also observed in shales of the North Sea due to diagenetic phenomena occurring along with mechanical compaction [49,50].

As noted earlier, carbonate rocks mainly consist of the grain and inter-particle cement, which binds the grains together. Any stress on the rock is carried by both the grain and cement, and the rock can deform in different ways depending on the strength of either part. On the other hand, carbonate rock burial history consists of two mechanical and

chemical compactions. At shallow depth, non-cemented rocks undergo a mainly mechanical compaction, which rearranges the position of grains with respect to each other as well as cement. This configuration and the grain-to-grain contact is, in turn, linked to chemical compaction in deeper sediments. Chemical compaction is dependent on the rock's petrophysical properties as well as pore fluid. Another controlling parameter in the extension of the chemical compaction processes is the diffusion of solids into the pore fluid, which is also a time dependent variable of mineral and fluid interactions.

The complexity of porosity in the carbonate pore system is due to a variety of reasons related to diagenesis, which converts a carbonate rock from a shallow porous sediment to a low porosity rock [21,51] and from an unstable mineral combination to a stable assemblage of low-magnesium calcite, etc. [51-53]. Geological structures also control the occurrence and frequency of the petrophysical features, such as dissolution and karstification, associated with subaerial exposures that occur over the depositional high and crest of structures. A brief explanation of the diagenetic processes in carbonate rocks was provided earlier whilst remarkable work has been done on this subject to understand the whole phenomenon, which is available in the literature [3,4,12,48,52,54-58]. In the context of diagenetic processes, the complexity of pore shape and size in carbonates is attributed to the wide range of particle sizes and shapes of the rock. The skeletal organs that form the grain, or primary hard framework, also exist in a wide range of textures, shapes and sizes. Initial pores are, therefore, expected to be highly varied as a result of particle positioning. Additionally, pore-filling cement also varies extensively in nature and size. In Choquette and Pray's [51] classification, the grain and pore-filling fluid interactions have been considered as a fabric-selective attribute of the rock, which takes into account the porosity enhancement potential of the mineral dissolution into the fluid phase. In many limestones, the volume of pore-filling cement is almost the same as, or even exceeds, that of the initial sediment, and additional porosity as a result of dolomitization is quantitatively minor compared to the porosity loss that happened in the subsurface [59].

In terms of the chemical reactions taking place in carbonate depositional environments, it is noted that pressure solution and calcium carbonate precipitation are two concurrent processes that have different impacts on rock porosity, and make porosity-depth predictions problematic [51,60]. In carbonates, chemical compaction is also strongly affected by pre-existing fracture patterns and vadose/phreatic contact [11]. Dissolution is partially affected by the depositional texture, where the soluble minerals, such as evaporites, are selectively dissolved in groundwater and create karstic porosity. This means that the existence of fractures filled with various minerals, including calcite, dolomite, anhydrite, galena, sphalerite, celestite, strontianite, and fluorite, is another source of carbonate rock complexity [2]. This causes increased heterogeneities in the petrophysical properties of these sediments and makes them less predictable.

Fractures are created under different conditions and they add huge uncertainty to geological models by their complex distribution in non-homogenous rock types. Fractures are dependent not only on stresses, but also on rock fabric and mechanical strength, which varies spatially over depth and width within a specific formation. Different fracture attributes, such as opening and length, make the rock properties highly variable. By increasing the load on carbonate rocks, cracks

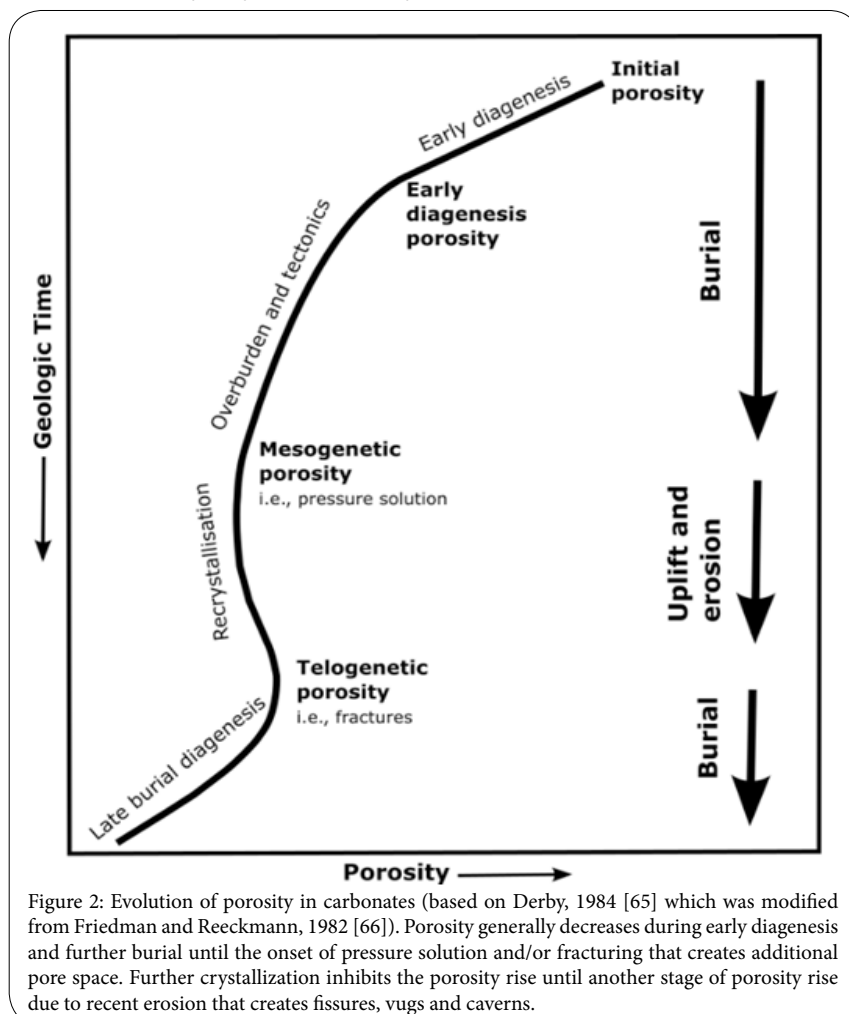
may originate and develop along the matrix (by-passing the harder grains and developing fissures in the fine-grained mud) or across the grain and matrix together. Fractures can develop perpendicular to the direction of least stress in the environment [36,61-63]. But the matrix and particles would have different response to the stresses. A test done by Shinn, Halley [64] on carbonate rocks revealed a surprising result of compressive behaviour of the rock particles. While there is a measurable bulk compressibility, the shells, foraminifera and other fossils remained uncrushed under compressive stress up to 7900 psi, confirming the selective creation of primary cracks in the matrix. In the folded structures, the highest intensity of fractures exists at the crest while there is a preferred orientation in the lowermost section of the layer that improves the vertical permeability parallel to the formation's structural axis.

It must be noted that diagenesis is not limited to compaction and, in many carbonate sediments, it involves tectonism and post-tectonism events. Therefore, recent diagenetic processes could also alter pre-existing porosities by a large amount, as pore fluid may flow through high permeability conduits and cause excess pore space. This may place formation porosity and permeability in complete opposition to Saller and Henderson's [25] findings, where they showed a destructive effect of dolomitization on porosity. However, the mineralogical modifications occur when the metastable minerals such as magnesian calcite and aragonite, as the main parts of the carbonate sediment, are exposed to the meteoric waters over the geological time leading

to dissolution and then stable mineral precipitation, in the form of calcite and dolomite [2,3]. A schematic of the varying relationship between porosity and depth in a carbonate rock (Figure 2) shows several stages of porosity evolution over time.

Although porosity normally decreases with depth (overburden), obtaining a function that relates compaction to just a single parameter, and specifically a function of effective stress, simply is not plausible for most carbonate rocks. As such, all existing pore pressure prediction methods struggle to provide reliable and accurate estimates in many carbonates. Several other important factors directly or indirectly control the porosity in the carbonate rocks. For example, Swarbrick [67] outlined the effect of temperature and its impact on cementation and seal development (and, consequently, overpressure generation). From that point of view, there are four aspects of traditional pore pressure prediction that are affected by temperature and include:

1. rocks become stiffer in higher temperature, which accelerates cementation;
2. higher temperature augments dissolution, and further rock body collapse reduces pore space;
3. higher temperature makes grains more ductile and weakens the rock framework, and;
4. mineral transformations (e.g., clay diagenesis) will be accelerated in higher temperatures.



Due to the large diversity in trends, there is no unique porosity-depth relationship that can be used to relate porosity to depth of burial and estimate porosity in present day. A list of rock body compaction models on chronological order has been presented by Chilingar, Robertson Jr [68], though all are based on porosity variation of clay versus depth of burial. Figure 3 shows different depth-porosity relationships for rocks in several geographical locations.

Since the nature and fabric of carbonates vary, petrophysical characterisation also requires a specific set of empirical relationships. Permeability-porosity relationships in carbonates, as a characteristic of the formation, also exhibit a wide range. Wackestones and mudstones exhibit the lowest porosity and permeability while reefal rocks provide the highest value of permeability, and grainstone as well as coccolith chalk provide the highest porosity. According to Lun et al. [74], stress sensitivity of porosity is lower than that of permeability in a carbonate environment. They analysed porosity and permeability in four types of core samples with different fracture attributes, including matrix cores and cores with non-packed, semi-packed and fully-packed fractures in the littoral Caspian Basin. As a result, the packing degree of fractures had a great impact on the stress sensitivity due to the closure of fractures under the overburden load.

In the tabulation of data gathered from various types of rock and reservoirs in several locations within the United States of America [75], carbonate rocks showed a narrower range of variation in porosity and bulk density in relation to depth of burial, age, degree of tectonic disturbance, and departure from homogeneous texture. Pure shale formations had the highest degree of relationship between porosity and increase in bulk density (stress sensitivity) with depth of burial and degree of tectonic disturbance, while sandstone generally, but not invariably, had good dependencies. However, such relationship is not a single-parameter function in carbonates. Changes in carbonate rock properties, including those due to burial are caused by different physicochemical processes occurring in the pore spaces of sediments

and rocks [76-79]. Putting all of this information together, the influence of gravitational compaction plays a secondary role in a carbonate rock evolution [80] and most carbonates are classified as non-stress-sensitive deposits. As a result, for example, the application of conventional pore pressure prediction methods on this type of rock becomes invalid.

Conclusions

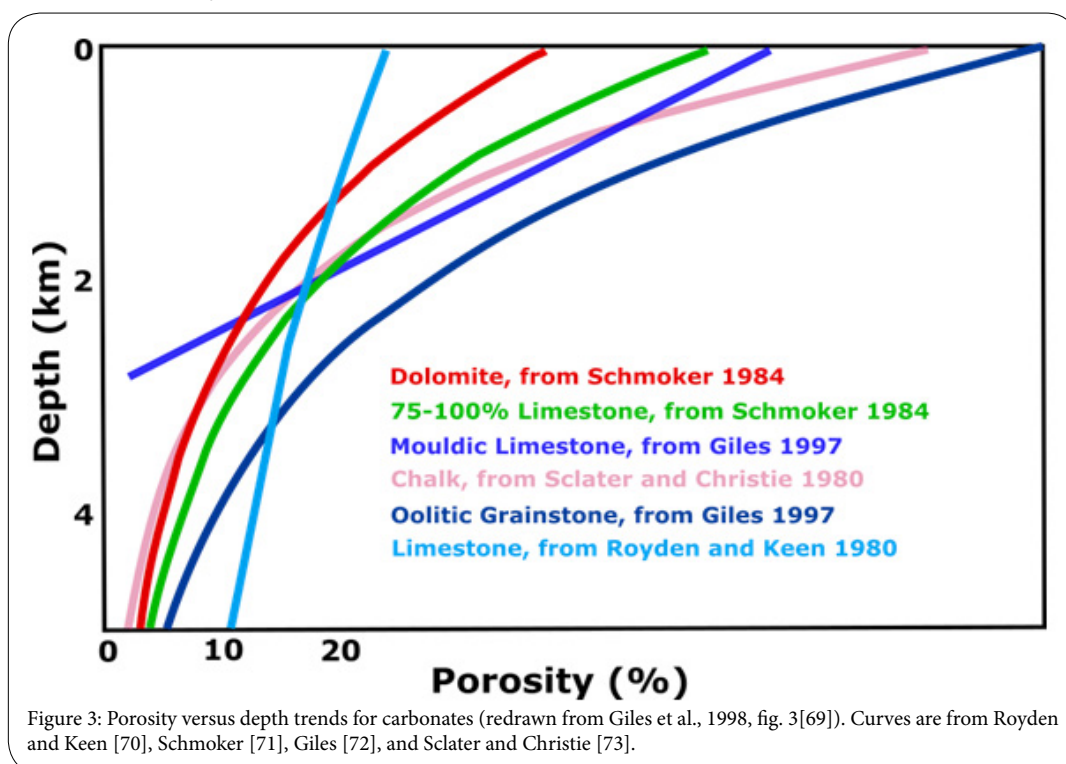
Porosity is controlled by several depositional and diagenetic processes in carbonate rocks, such as pressure solution, cementation, dolomitization, brecciation and fracturing. Some of these processes create additional pore space while some diminish the existing porosity. Any of them can occur in specific conditions regardless of burial depth. For example, carbonate deposits may be associated with extensive fracturing, which further creates brecciation patterns in the rocks. Pressure solution and stylolites filled with calcite and dolomite cements are also frequently evident in these formations. Although in-situ stresses significantly affect some of those processes, the overall porosity evolution isn't entirely controlled by stress. Therefore, carbonate rocks are classified as non-stress-sensitive and hence, hard to attribute a normal compaction trend for the purpose of predicting pore pressure.

Competing Interests

The authors declare that they have no competing interests.

References

1. Cui H (2012) Carbonate Diagenesis (stories behind the microscopic images).
2. Moore CH (2001) Carbonate Reservoirs: Porosity, Evolution & Diagenesis in a Sequence Stratigraphic Framework. Elsevier.
3. Moore CH (1989) Carbonate diagenesis and porosity. Elsevier.



4. Sherman C, Fletcher C, Rubin K (1999) Marine and meteoric diagenesis of Pleistocene carbonates from a nearshore submarine terrace, Oahu, Hawaii. *Journal of Sedimentary Research* 69: 1083-1097.
5. Connolly CA, Walter LM, Baadsgaard H, Longstaffe FJ (1990) Origin and evolution of formation waters, Alberta Basin, Western Canada sedimentary Basin. I. Chemistry. *Applied Geochemistry* 5: 375-395.
6. Stoessell RK, Moore CH (1983) Water-rock interactions. *The American Association of Petroleum Geologists Bulletin* 67: 896-906.
7. Stueber AM, Walter LM (1991) Origin and chemical evolution of formation waters from Silurian-Devonian strata in the Illinois basin, USA. *Geochimica et Cosmochimica Acta* 55: 309-325.
8. Coogan AH (1970) Measurements of compaction in oolitic grainstone. *Journal of Sedimentary Research*.
9. Bhattacharyya A, Friedman GM (1979) Experimental compaction of ooids and lime mud and its implication for lithification during burial. *Journal of Sedimentary Research* 49: 1279-1286.
10. Sibley DF, Blatt H (1976) Intergranular pressure solution and cementation of the Tuscarora orthoquartzite. *Journal of Sedimentary Research* 46: 881-896.
11. Lucia FJ (1999) *Carbonate Reservoir Characterization*. Springer.
12. Bathurst RG (1972) *Carbonate sediments and their diagenesis*. Elsevier.
13. Durney DW (1972) Solution-transfer, an important geological deformation mechanism. *Nature* 235: 315-317.
14. Koepnick RB (1984) Distribution and vertical permeability of stylolites within a lower Cretaceous carbonate reservoir, Abu Dhabi, United Arab Emirates. *Stylolites and Associated Phenomena: Relevance to Hydrocarbon Reservoirs*, Abu Dhabi National Reservoir Foundation Special Publication.
15. Rutter EH, Elliott D (1976) The Kinetics of Rock Deformation by Pressure Solution. 283: 203-219.
16. Dravis JJ, Muir ID (1993) Deep-Burial Brecciation in the Devonian Upper Elk Point Group, Rainbow Basin, Alberta, Western Canada. *Paleokarst Related Hydrocarbon Reservoirs*, SEPM Society for Sedimentary Geology.
17. Cantrell DL, Hagerty RM (1999) Microporosity in Arab Formation Carbonates, Saudi Arabia. *Geo Arabia* 4: 129-154.
18. Al-Ramadan K, Hussain M, Imam B, Saner S (2004) Lithologic characteristics and diagenesis of the Devonian Jauf Sandstone at Ghawar Field, Eastern Saudi Arabia. *Marine and Petroleum Geology* 21: 1221-1234.
19. Aharonov E, Katsman R (2009) Interaction between pressure solution and clays in stylolite development: Insights from modeling. *American Journal of Science* 309: 607-632.
20. Baker PA, Kastner M, Byerlee JD, Lockner DA (1980) Pressure solution and hydrothermal recrystallization of carbonate sediments - An experimental study. *Marine Geology* 38: 185-203.
21. Heydari E (2000) Porosity Loss, Fluid Flow, and Mass Transfer in Limestone Reservoirs: Application to the Upper Jurassic Smackover Formation, Mississippi. *AAPG Bulletin* 84: 100-118.
22. Sassen R, Moore CH (1988) Framework of hydrocarbon generation and destruction in eastern Smackover trend. *AAPG Bulletin* 72: 649-663.
23. Lucia FJ (2007) *Carbonate reservoir characterization: an integrated approach*. Springer Science & Business Media.
24. Murray RC (1960) Origin of Porosity in Carbonate Rocks. *Journal of Sedimentary Research* 30: 59-84.
25. Saller AH, Henderson N (1998) Distribution of porosity and permeability in platform dolomites: Insight from the Permian of west Texas. *AAPG bulletin* 82: 1528-1550.
26. Hardie LA (1967) The Gypsum-Anhydrite Equilibrium at One Atmosphere Pressure. *The American Mineralogist* 52: 171-200.
27. Blount DN, Moore CH (1969) Depositional and Non-Depositional Carbonate Breccias, Chiantla Quadrangle, Guatemala. *Geological Society of America Bulletin* 80: 429-442.
28. Kerans C (1990) Depositional systems and karst geology of the Ellenburger Group (Lower Ordovician), subsurface West Texas.
29. Miller JA (1985) Depositional and reservoir facies of the Mississippian Leadville Formation, northwest Lisbon field, Utah, in *Carbonate Petroleum Reservoirs*. Springer.
30. Craig D (1988) *Caves and Other Features of Permian Karst in San Andres Dolomite, Yates Field Reservoir, West Texas*. Springer New York.
31. Kerans C (1993) Description and interpretation of karst-related breccia fabrics, Ellenburger Group, West Texas.
32. Narr W, Schechter DW, Thompson LB (2006) *Naturally fractured reservoir characterization*. Richardson, TX: Society of Petroleum Engineers.
33. Nairn A, Alsharhan AS (1997) *Sedimentary Basins and Petroleum Geology of the Middle East*. Burlington: Elsevier Science.
34. McQuillan H (1985) Fracture-controlled production from the Oligo-Miocene Asmari Formation in Gachsaran and Bibi Hakimeh fields, southwest Iran. *Carbonate Petroleum Reservoirs*. Springer.
35. Zahm CK, Zahm LC, Bellian JA (2010) Integrated fracture prediction using sequence stratigraphy within a carbonate fault damage zone, Texas, USA. *Journal of Structural Geology* 32: 1363-1374.
36. Zoback MD (2007) *Reservoir Geomechanics*. New York: cambridge university press.
37. Nelson R (2001) *Geologic analysis of naturally fractured reservoirs*. Gulf Professional Publishing.
38. Turcotte D (1974) Are transform faults thermal contraction cracks? *Journal of Geophysical Research* 79: 2573-2577.
39. Kulander B, Dean S, Ward Jr B (1990) *Natural Fractures in Core: Chapter 2*.
40. Fertl WH (1980) Evaluation of Fractured Reservoir Rocks Using Geophysical Well Logs. *Unconventional Gas Recovery Symposium*. Society of Petroleum Engineers: Pittsburgh, Pennsylvania.
41. Ramsay JG (1967) *Folding and fracturing of rocks*. New York: McGraw-Hill.
42. Lisle RJ (1995) Detection of zones of abnormal strains in structures using Gaussian curvature analysis. *American Association of Petroleum Geologists Bulletin* 78: 1811-1819.
43. Calladine CR (1986) *Gaussian curvature and shell structures. The mathematics of surfaces*. Clarendon Press: Oxford.
44. Fischer MP, Wilkerson MS (2000) Predicting the orientation of joints from fold shape: Results of pseudo-three-dimensional modeling and curvature analysis. *Geology* 28: 15-18.
45. Kendall CGSC, Schlager W (1981) Carbonates and relative changes in sea level. *Marine Geology* 44: 181-212.
46. Milliman JD (1974) *Marine carbonates*. Springer.
47. Wilson JL (1975) *Carbonate facies in geologic history*. Springer.
48. Schmoker JW, Halley RB (1982) Carbonate porosity versus depth: a predictable relation for South Florida. *AAPG Bulletin* 66: 2561-2570.
49. Teige GMG, Hermanrud C, Wensaas L, Bolas HMN (1999) The lack of relationship between overpressure and porosity in North Sea and Haltenbanken shales. *Marine and Petroleum Geology* 16: 321-335.
50. Carstens H, Dypvik H (1981) Abnormal formation pressure and shale porosity. *AAPG Bulletin* 65: 344-350.
51. Choquette PW, Pray LC (1970) Geologic nomenclature and classification of porosity in sedimentary carbonates. *AAPG bulletin* 54: 207-250.
52. Land LS (1967) Diagenesis of skeletal carbonates. *Journal of Sedimentary Research*.
53. Runnegar B (1985) Shell microstructures of Cambrian molluscs replicated by phosphate. *Alcheringa: An Australasian Journal of Palaeontology* 9: 245-257.
54. Neugebauer J (1973) The diagenetic problem of chalk - the role of pressure solution and pore fluid. *Neues Jahrbuch Geologie Palaeontologie Abhandlung* 143: 223-245.
55. Scholle PA (1977) Chalk diagenesis and its relation to petroleum exploration: oil from chalks, a modern miracle? *AAPG Bulletin* 61: 982-1009.
56. Garrison RE (1981) Diagenesis of oceanic carbonate sediments: a review of the DSDP perspective.
57. James NP, Choquette PW (1983) *Diagenesis 6. Limestones - The Sea Floor Diagenetic Environment*.
58. Koepnick RB (1987) Distribution and Permeability of Stylolite-Bearing Horizons within a Lower Cretaceous Carbonate Reservoir in the Middle East. *SPE Formation Evaluation* 2: 137-142.
59. Pray LC, Choquette PW (1966) Genesis of carbonate reservoir facies: ABSTRACT. *AAPG Bulletin* 50: 632-632.
60. Harris WH, Matthews R (1968) Subaerial diagenesis of carbonate sediments: efficiency of the solution-reprecipitation process. *Science* 160: 77-79.

61. Nordgren RP (1972) Propagation of a Vertical Hydraulic Fracture. *Society of Petroleum Engineers Journal* 12: 306-314.
62. Salz LB (1977) Relationship between fracture propagation pressure and pore pressure. Annual Fall Technical Conference and Exhibition, American Institute of Mining, Metallurgical, and Petroleum Engineers, Inc.: Denver, Colorado.
63. Ingraffea AR (1987) 3 - Theory of crack initiation and propagation in rock. *Fracture Mechanics of Rock*.
64. Shinn EA, Halley RB, Hudson JH, Lidz BH (1977) Limestone compaction: An enigma. *Geology*.
65. Derby JR (1984) Exploration for Carbonate Reservoirs or "Just Find a Fat Limestone with Holes in it".
66. Pau EAC, FriedmanGM, Reeckmann A (1982) Exploration for carbonate Petroleum Reservoirs. *J Wiley*.
67. Swarbrick R (2012) Review of pore-pressure prediction challenges in high-temperature areas. *The Leading Edge* 31: 1288-1294.
68. Chilingar GV, Robertson Jr JO, Rieke Iii HH (2002) Chapter 2 Origin of abnormal formation pressures. *Developments in Petroleum Scienc*.
69. Giles MR, Indrelid SL, s, Compaction - The Great Unknown in Basin Modelling. *Geological society, London, Special Publications*: 141.
70. Royden L, Keen CE (1980) Rifting process and thermal evolution of the continental margin of Eastern Canada determined from subsidence curves. *Earth and Planetary Science Letters* 51: 343-361.
71. Schmoker JW (1984) Empirical relation between carbonate porosity and thermal maturity: An approach to regional porosity prediction. *AAPG Bulletin* 68: 1697-1703.
72. Giles MR (1977) Diagenesis: a quantitative perspective: implications for basin modelling and rock property prediction. *Kluwer Academic Publishers: Dordrecht*.
73. Sclater JG, Christie PAF (1980) Continental Stretching - an Explanation of the Post-Mid-Cretaceous Subsidence of the Central North-Sea Basin. *Journal of Geophysical Research* 85: 3711-3739.
74. Lun Z, Yefei C, Zhengfu N, Xuelin W, Lifang L, et al. (2013) Stress sensitive experiments for abnormal overpressure carbonate reservoirs: A case from the Kenkiyak fractured-porous oil field in the littoral Caspian Basin. *Petrol Explor Develop* 40: 208-215.
75. Magner E (1963) Porosity and Bulk Density of Sedimentary Rocks. *US Atomic Energy Commission. Washington*.
76. Bagrintseva KI (1977) Carbonate Oil and Gas Reservoir rocks. *Moscow: Nedra Publishing House*.
77. Pavlova NN (1975) Deformation and Reservoir Properties of Rocks. *Moscow: Nedra Publishing House*.
78. Proshlyakov BK (1974) Secondary Changes of Oil and Gas Reservoir Rocks. *Moscow: Nedra Publishing House*.
79. Chilingar GV, Bissell HJ, Wolf KH (1979) Diagenesis of carbonate sediments and epigenesis. *Diagenesis in sediments and sedimentary rocks* 25: 247.
80. Buryakovskiy LA, Djevanshir RD, Chilingarian GV (1991) Mathematical simulation of sediment compaction. *Journal of Petroleum Science and Engineering* 5: 151-161.