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A model for porosity evolution during creep compaction of sandstones

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Abstract

A coupled creep-compaction and chemical-reaction model is developed to predict the porosity evolution for quartzose sandstones as a function of strain. The model also demonstrates the relative importance of grain-contact dissolution and cementation for both uniaxial and isotropic compaction. Theoretical analysis indicates that porosity reduction during compaction of sandstones is nonlinearly related to strain. In open systems, porosity loss is also related to grain packing, stress state, and pore-fluid saturation state. Grain-contact dissolution is the dominant mechanism for porosity loss in a closed system and, with increasing compaction, cementation becomes increasingly important. Compared to uniaxial compaction, isotropic compaction leads to more porosity loss due to grain-contact dissolution, but less porosity loss due to cementation. With compaction, pore-fluid saturation state has an increasing effect on porosity loss. Higher saturation state enhances porosity loss due to cementation. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: porosity; compaction; pressure solution; cementation; saturation

1. Introduction

During the past 50 years, porosity evolution during diagenesis of sandstones has been the subject of numerous investigations. Porosity information is important to constrain models of fluid flow and heat and mass transport in sedimentary basins and in groundwater and petroleum reservoirs. It is commonly accepted that porosity reduction in sandstones is caused by the combined effects of mechanical compaction, grain-contact dissolution,

and cementation [1,2]. Purely mechanical processes such as rotation, slippage, and brittle and plastic deformation of grains probably dominate during the early stages of burial [3,4]. Grain-contact dissolution is a grain-scale deformation mechanism that allows large-scale compaction of sediments during diagenesis. Material dissolves along grain–grain contacts under high stress, diffuses along the contacts, and precipitates in pore spaces [5–8]. The fundamental mechanisms and driving forces for grain-contact dissolution include intergranular pressure solution [2,5,8] and microcracking and microgranulation at grain contacts [8,9]. Both mechanical compaction and grain-contact dissolution cause the reduction of bulk volume. Cementation or precipitation of minerals in the

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pore space results in the reduction of intergranular porosity, but is not directly related to the reduction of bulk volume. In addition to the three important porosity-reducing mechanisms, a porosity-increasing mechanism is free-surface dissolution or dissolution of minerals at pore walls.

The studies of porosity evolution during compaction of sandstones are mainly focused on three aspects: (1) theoretical modeling [10–15], (2) experiments [8,9,16–18], and (3) the observations of naturally deformed samples [1,19–21]. It is clear from these studies that grain-contact dissolution reduces porosity of sandstones by compaction strain and by providing a source for cement. Despite these efforts, porosity evolution during compaction is still poorly understood, especially with regard to how porosity loss as a function of strain depends on loading conditions (e.g., uniaxial vs. isotropic) and solute concentration in the pore fluid. The objectives of this paper are to quantify porosity reduction during compaction of sandstones as a function of strain for both uniaxial and isotropic compaction, to evaluate the dependence of porosity loss on saturation state of solution, and to establish the relative importance of grain-contact dissolution and cementation to porosity reduction. For our analysis, we assume that mechanical compaction is finished in the early stages of compaction and focus on the porosity reduction after mechanical compaction.

2. Theoretical constraints for porosity evolution during creep compaction of sandstones

Porosity evolution during creep compaction of sandstones is the combined result of the coupled processes of compaction and chemical reaction (Fig. 1). Compaction is the bulk volume reduction caused by the grain-contact dissolution. Chemical reactions take place at both grain–grain contacts and free surfaces. Assuming that porosity (ϕ) is uniform in space ($\nabla\phi=0$), the porosity evolution during compaction of sandstones can be described by the following mass conservation equation [22]:

$$\frac{d}{dt}((1-\phi)V_m^{-1}) - \frac{1-\phi}{1-\varepsilon}V_m^{-1}\dot{\varepsilon} = -R \quad (1)$$

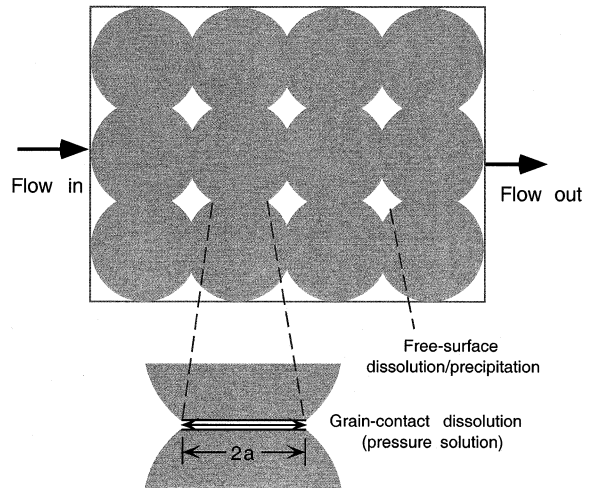


Fig. 1. Schematic diagram showing coupled compaction and chemical reactions during creep compaction. Material dissolves along grain–grain contact, which leads to grain convergence. The dissolved material transports along grain contacts and precipitates in pore spaces.

where ε and $\dot{\varepsilon}$ are the volumetric strain and strain rate, respectively, R is the total chemical reaction rate, V_m is the molar volume of quartz, and t is the time. The first term on the left represents the rate of change of solid materials ($\text{mol}/(\text{m}^3 \text{ s})$), the second term accounts for compaction, and the term on the right represents chemical reaction. Volumetric strain rate is defined as:

$$\dot{\varepsilon} = \frac{d\varepsilon}{dt} = -\frac{1}{V_0} \frac{dV}{dt} \quad (2)$$

where V is the system volume and V_0 is the initial system volume.

Consider a single intergranular contact between two identical spherical grains of quartz with grain size d (Fig. 1). The dissolution rate along a single intergranular contact (r_i) can be given by:

$$r_i = \frac{1}{2}\pi a^2 \dot{\varepsilon}_1 d V_m^{-1} \quad (3)$$

where $\dot{\varepsilon}_1$ is the linear convergence rate of grains ($1/\text{s}$), a is the grain contact radius. r_i has unit of mol/s .

Assuming that the same amount of dissolution occurs along each grain contact and neglecting

the volume of the segments dissolved for each sphere, the dissolution rate at the grain–grain contacts corresponding to unit bulk volume of sample (solid+pore) (R_1 , mol/(m³ s)) can be written as:

$$R_1 = \frac{3n(1-\phi)a^2\dot{\epsilon}_1}{V_m d^2} \quad (4)$$

where n is the number of contacts per grain along which dissolution occurs.

Consider an idealized aggregate consisting of a simple cubic pack of spherical grains of diameter d . The linear convergence of grains $\dot{\epsilon}_1$ is linearly or approximately linearly related to the volumetric strain $\dot{\epsilon}$. We assume:

$$\dot{\epsilon}_1 = \gamma\dot{\epsilon} \quad (5)$$

The value γ depends on the stress states: γ is 1 for uniaxial compaction, while for isotropic compaction, γ is generally between 1/3 and 1/2 (approximately 1/3 if volumetric strain is less than 10%).

The grain contact radius (a) changes with strain (ϵ) and grain size (d) as follows:

$$a^2 = \left(\frac{d}{2}\right)^2 - \left(\frac{d-\epsilon_1 d}{2}\right)^2 \approx \frac{1}{2}d^2\epsilon_1 = \frac{1}{2}\gamma d^2\epsilon \quad (6)$$

Combination of Eqs. 4–6 yields:

$$R_1 = \frac{3n\gamma^2(1-\phi)\epsilon\dot{\epsilon}}{2V_m} \quad (7)$$

The dissolution/precipitation rate at free surfaces can be represented by:

$$R_2 = k_m A_m \left(1 - \frac{c}{c_{eq}}\right) \quad (8)$$

where k_m is the surface reaction rate constant of quartz, A_m is the quartz-sand surface area per unit volume of porous medium, c is silica concentration, and c_{eq} is the equilibrium concentration of pore fluid [23,24]. Dissolution rate is taken to be positive and the precipitation rate is negative.

The total reaction rate (R) of the system is the sum of the dissolution at grain contacts and dissolution/precipitation rate at free surfaces. After

replacing reaction rate R with R_1+R_2 , Eq. 1 can be rewritten as:

$$\frac{d\phi}{dt} + \frac{1-\phi}{1-\epsilon}\dot{\epsilon} = RV_m = \frac{3}{2}n\gamma^2(1-\phi)\epsilon\dot{\epsilon} + V_m k_m A_m \left(1 - \frac{c}{c_{eq}}\right) \quad (9)$$

which gives the evolution of porosity as a function of strain and saturation state.

3. Porosity evolution

3.1. Porosity evolution in a closed system

In a closed system, all material dissolved at grain contacts is locally precipitated as cement in pore space. In the continuing processes of dissolution at grain contacts, diffusion along grain contacts, and precipitation in pore spaces, the rate is controlled by the slowest one among them (rate-controlling step). Some studies indicate that diffusion is the rate-controlling process [8,25,26], and other studies indicate that precipitation or dissolution is the rate-controlling step [6,27,28]. If the system is in steady state, the grain-contact dissolution rate will be equal the precipitation rate at pore spaces [29]. The total reaction rate (R) in the system is zero. Assuming the molar volume of minerals does not change significantly with pressure, the differential Eq. 9 can be rewritten as:

$$\frac{d\phi}{dt} + \frac{1-\phi}{1-\epsilon}\dot{\epsilon} = 0 \quad (10)$$

Eq. 10 has the following solution:

$$\phi = 1 - \frac{1-\phi_0}{1-\epsilon} \quad (11)$$

where ϕ_0 represents the initial porosity ($\epsilon=0$). Eq. 11 gives the relationship between porosity and strain in a closed system. It indicates that porosity is nonlinearly related to volumetric strain (Fig. 2). As ϵ increases from zero to ϕ_0 , porosity (ϕ) nonlinearly decreases from ϕ_0 to zero. Porosity loss as a function of strain in a closed system does not

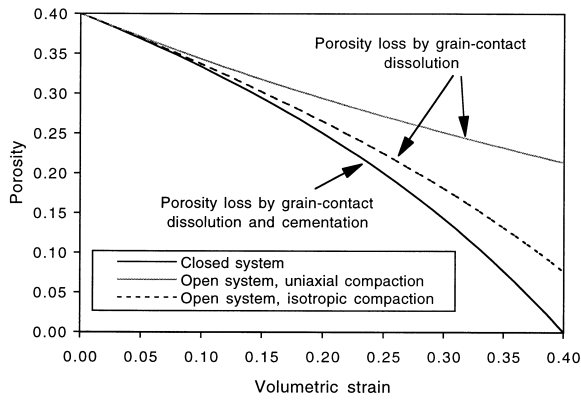


Fig. 2. Porosity evolution as a function of strain for a closed system and an open system with saturated pore fluid. Initial porosity is assumed to be 40%.

depend on the loading conditions, grain size, or grain shape. Mechanical compaction also obeys the same mathematical relationship as Eq. 11 because the total reaction rate in the system is zero.

Over the compaction path from $\varepsilon=0$ to $\varepsilon=\phi_0$ porosity loss is less than the achieved strain. The difference between achieved strain and porosity loss is expressed as (Δ):

$$\Delta = \varepsilon - (\phi_0 - \phi) = \frac{\varepsilon(\phi_0 - \varepsilon)}{1 - \varepsilon} \quad (12)$$

The maximum difference, $\Delta_{\max} = (1 - \sqrt{1 - \phi_0})^2$, occurs at $\varepsilon = 1 - \sqrt{1 - \phi_0}$. With increasing strain (compaction), the difference (Δ) gradually increases from zero to the maximum, and then gradually decreases to zero again. The difference also increases with initial porosity (ϕ_0).

3.2. Porosity evolution for an open system with saturated pore fluid

In an open system with pore fluid saturated with respect to quartz, no free surface dissolution or precipitation occurs ($R_2 = 0$). Assuming that all material dissolved at grain contacts is removed from the system, then the only chemical reaction is grain contact dissolution (R_1). For these con-

ditions, Eq. 9 can be rewritten as:

$$\frac{d\phi}{dt} + \frac{1-\phi}{1-\varepsilon} \dot{\varepsilon} = \frac{3}{2} n \gamma^2 (1-\phi) \varepsilon \dot{\varepsilon} \quad (13)$$

which has the following solution:

$$\phi = 1 - \frac{1-\phi_0}{1-\varepsilon} e^{-\frac{3}{4} n \gamma^2 \varepsilon^2} \quad (14)$$

Eq. 14 indicates that porosity loss is nonlinearly related to strain and also depends on loading conditions (γ), and grain packing which controls the number of contacts per grain (n).

Consider two cases of simple cubic-packed spherical grains of uniform size undergoing uniaxial and isotropic compaction creep. The values of n and γ are taken to be 2 and 1 for uniaxial compaction, and 6 and 1/3 for isotropic compaction, respectively. The porosity loss is shown in Fig. 2 with the closed-system results. Clearly, at any given volumetric strain, porosity loss in open systems is less than that in closed systems due to the removal of dissolved material. The difference in porosity loss between the open and closed systems increases with volumetric strain. Also, porosity loss due to grain-contact dissolution is greater for isotropic compaction than uniaxial compaction.

3.3. Porosity evolution as a function of saturation state of pore fluid

It is commonly accepted that porosity evolution depends on the saturation state of pore fluid. Supersaturation leads to solute precipitation (cementation), and undersaturation leads to free surface dissolution. According to Eq. 9, the dependence of porosity evolution on saturation state can be quantitatively evaluated. Assuming that initial pore fluid (at $\varepsilon=0$) is saturated, integration of Eq. 9 yields the following result:

$$\phi = 1 - \frac{1-\phi_0}{1-\varepsilon} \exp\left(\int_0^{\varepsilon} \frac{1-\phi_0 - R V_m}{1-\phi} dt'\right) =$$

$$1 - \frac{1 - \phi_0}{1 - \varepsilon} e^{-\frac{3}{4}n\gamma^2\varepsilon^2} \exp\left[\int_0^{\varepsilon} \frac{V_m k_m A_m}{1 - \phi} \left(\frac{c}{c_{eq}} - 1\right) dt'\right] \quad (15)$$

where ϕ_0 represents the initial porosity (at $\varepsilon=0$). Eq. 15 indicates that porosity (ϕ) is related to saturation state (c/c_{eq}).

Letting $S_m = c/c_{eq} - 1$, then the total reaction rate (R) is represented by:

$$R = R_1 + R_2 = (S_{m0} - S_m)k_m A_m \quad (16)$$

where $S_{m0} = [3n\gamma^2(1 - \phi)\varepsilon\dot{\varepsilon}]/(2V_m k_m A_m)$. When $S_m = S_{m0}$, $R=0$ and Eq. 15 represents the situation described by Eq. 11 (closed systems). If $S_m = 0$, pore fluid is saturated ($R_2 = 0$) and Eq. 15 represents the situation described by Eq. 14 (open system with saturated pore fluid).

In addition to the two cases, porosity evolution was modeled for three open systems with different pore-fluid chemistry and transport conditions: (A) pore fluid is supersaturated and more material is brought into the system than removed (net import, more supersaturated than a closed system), (B) pore fluid is supersaturated but more material is removed from the system than brought in (net

export, less supersaturated than a closed system), and (C) pore fluid is undersaturated. In the simulations, S_m is assumed to be $2S_{m0}$ for (A), $S_{m0}/2$ for (B), and $-S_{m0}$ for (C). The simulated results for the three different chemical conditions are shown in Fig. 3 along with the results of the closed and open systems shown in Fig. 2. Although the saturation state has little effect on porosity loss at low strains, it becomes increasingly important with increasing compaction. An increase in saturation state leads to accelerated porosity loss, primarily through its control on cementation.

4. Relative importance of grain-contact dissolution and cementation

Porosity loss due to grain-contact dissolution ($\Delta\phi_d$) is a function of strain and is independent of pore-fluid saturation state. It can be obtained from Eq. 14 and expressed as:

$$\Delta\phi_d = \phi_0 - \phi = (1 - \phi_0) \left[\left(\frac{1}{1 - \varepsilon} \right) e^{-\frac{3}{4}n\gamma^2\varepsilon^2} - 1 \right] \quad (17)$$

The porosity loss due to cementation ($\Delta\phi_c$) is the difference between the total porosity loss and the porosity loss due to grain-contact dissolution ($\Delta\phi_d$) (Eq. 17). In a closed system, the total porosity loss is $(\phi_0 - \phi)$ in which ϕ is represented by Eq. 11. The porosity loss due to cementation is expressed as:

$$\Delta\phi_c = \left(\frac{1 - \phi_0}{1 - \varepsilon} \right) \left(1 - e^{-\frac{3}{4}n\gamma^2\varepsilon^2} \right) \quad (18)$$

The porosity losses due to the two competing mechanisms as a function of strain in a closed system are shown in Fig. 4a,b. The relative importance of grain-contact dissolution, which is represented by $\Delta\phi_d/(\Delta\phi_d + \Delta\phi_c)$, is shown in Fig. 4c. While porosity losses due to both grain-contact dissolution and cementation increase with increasing volumetric strain, loading conditions have significant effects on the two competing

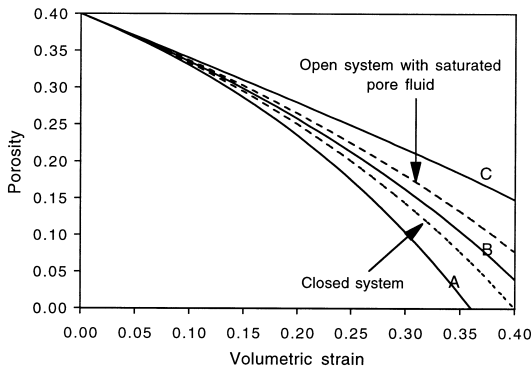


Fig. 3. Porosity evolution as a function of strain for different saturation states of solution. Isotropic compaction is assumed. Curve A, supersaturated ($S_m = 2S_{m0}$); curve B, supersaturated ($S_m = (1/2)S_{m0}$); curve C, undersaturated ($S_m = -S_{m0}$). The dashed lines represent the closed system ($S_m = S_{m0}$) and the open system with the saturated pore fluid ($S_m = 0$). Initial porosity is assumed to be 40%.

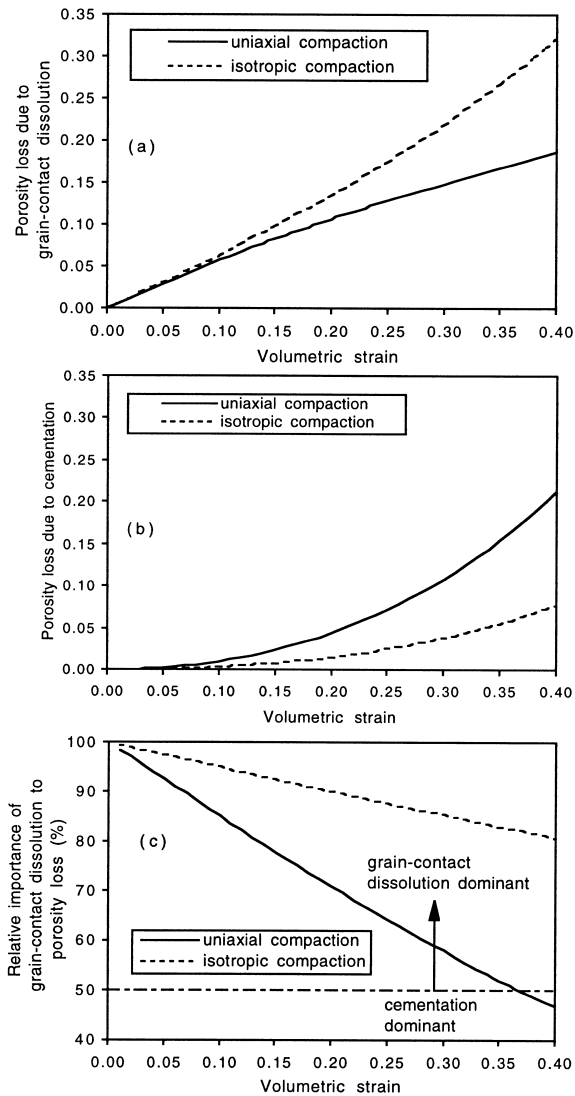


Fig. 4. Porosity loss as a function of strain for uniaxial and isotropic compaction due to (a) grain-contact dissolution and (b) cementation. The relative importance of porosity loss due to grain-contact dissolution (c). Initial porosity is assumed to be 40%.

mechanisms. Compared to uniaxial compaction, isotropic compaction leads to more porosity loss due to grain-contact dissolution (Fig. 4a), but less porosity loss due to cementation (Fig. 4b). For isotropic compaction, grain-contact dissolution is always the dominant mechanism for porosity loss. For uniaxial compaction, grain-contact dissolution is the dominant mechanism for strain

< 35%, which covers most geological conditions (Fig. 4c). Cementation becomes increasingly important with increasing strain. The relative importance of the two competing mechanisms given by the model is consistent with the theoretical results of Mitra and Beard [10]. It is also consistent with the observations of naturally deformed samples [1].

The change of the relative importance of grain-contact dissolution and cementation is related to the change of grain-contact area. With increasing compaction, grain-contact area increases, which leads to an increase in the amount of material that must dissolve from grain contacts per unit of grain convergence. This increase in dissolved material results in a larger contribution of cementation to porosity loss.

The difference in the porosity evolution for uniaxial and isotropic compaction is also related to the changes of grain-contact area. Assuming a simple cubic pack of spherical grains of the same size, the grain convergence for isotropic compaction is approximately a third of the grain convergence for uniaxial compaction at any given volumetric strain. Although grains undergoing isotropic compaction have three times more contacts than those undergoing uniaxial compaction, the average grain contact area is larger for uniaxial compaction. This leads to more porosity loss due to cementation for uniaxial compaction (Fig. 4b). Total porosity loss in a closed system is not dependent on the loading conditions (Eq. 11) because all dissolved material is precipitated in the system. Consequently, porosity loss by grain-contact dissolution is less important for uniaxial compaction than for isotropic compaction.

In an open system, the relative importance of grain-contact dissolution and cementation may be significantly influenced by saturation state. An increase in saturation state leads to more porosity loss (Eq. 15, Fig. 3) by increasing contribution due to cementation.

The effects of fluid flow and the transport of dissolved materials can also be observed in sedimentary basins [19,30,31]. Pore-fluid flow redistributes the cement and changes the relative importance of the different mechanisms for porosity loss by affecting the saturation state of pore

fluid. Some rocks experience strong grain-contact dissolution but weak cementation when pore fluid is undersaturated or slightly supersaturated [19,31] while others show the opposite effects when the pore fluid is highly supersaturated [19,30,31].

5. Conclusions

The theoretical model indicates that porosity reduction during creep compaction of sandstones is nonlinearly related to strain. In open systems, porosity is also related to grain packing, stress state, and pore-fluid saturation state. Grain-contact dissolution is the dominant mechanism for porosity loss in a closed system, but cementation becomes increasingly important with compaction. The relative importance of porosity-reducing mechanisms is related to the loading conditions. Isotropic compaction leads to more porosity loss due to grain-contact dissolution and less porosity loss due to cementation than does uniaxial compaction. With increasing compaction, pore-fluid saturation state has an increasing effect on porosity loss. Higher saturation state enhances porosity loss due to cementation.

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