The variation of porosity in quartzose sandstones is calculated as a function of depth, temperature gradient, burial rate, stylolite frequency, and hydrocarbon saturation. Calculations were performed by considering the effects of both mechanical compaction and chemical compaction/cementation. This latter process dominates at temperatures greater than approximately 90°C and is due to quartz redistribution within the sandstone. Quartz redistribution stems from clay-induced quartz dissolution at stylolite interfaces, coupled with diffusional transport of aqueous silica into the interstyrolite sandstone and precipitation on quartz surfaces as cement. Many model parameters are obtained from theoretical calculations or laboratory measurements, and few basin-dependent parameters are required to make porosity predictions. A set of porosity predictions is presented in porosity/depth figures. Close correspondence between computed results and measured porosities in cores from a variety of sedimentary basins demonstrates the accuracy of the predictions.

Porosity evolution in sandstones is the result of physical and chemical processes acting over geological time. Properties controlling sandstone porosity in sedimentary basins include mineralogy, texture, and pressure, temperature, and burial histories. Because of the need to assess the quality of sandstone petroleum reservoirs, numerous efforts have been made to correlate porosity to these fundamental properties. Many efforts are based on empirical correlations of porosity with one or more independent variables (Scherer, 1987; Schnooker and Gautier, 1988; Bloch, 1991). Because correlation does not imply causality, such correlations are limited in their application to specific basins, rock units, or depth regimes, and can lead to large uncertainties when extended to other settings.

To overcome these limitations, many workers have tried to understand the physical and chemical processes involved in porosity evolution. The majority of their studies assumed that quartz cementation stems from the influx of large volumes of SiO$_2$-rich aqueous fluids (Van Hise, 1904; Siever, 1959; Land and Dutton, 1978; Blatt, 1979; Wood and Hewett, 1984; Leder and Park, 1986; McBride, 1989; Gluyas and Coleman, 1992; Hogg et al., 1992; Robinson and Gluyas, 1992a; Canals and Meunier, 1995) or from the effects of pressure solution (Weyl, 1959; Durney, 1976; De Boer, 1977a, b; Elverhøi and Bjørlykke, 1978; Robin, 1978; Bjørlykke et al., 1979; Anegovine and Turcotte, 1983; Merino et al., 1983; Tada et al., 1987; Tada and Siever, 1989; Dewers and Ortoleva, 1990; Mullis, 1992, 1993; Ramm, 1992; Stephenson et al., 1992); however, mass balance calculations presented by Bjørlykke (1979, 1983, 1993) showed that insufficient H$_2$O is available to advectively transport the quantity of quartz cement found in sedimentary basins. Furthermore, petrographic evidence reported by Bjørkum (1996) demonstrated that pressure is not the major factor controlling quartz dissolution at stylolites. An additional complication is that predicting the extent and volume of quartz cementation due to silica influx or pressure solution requires information that cannot be
generally acquired without extensive basin-specific geological and petrographical data. Specifically for the case of the fluid influx model, one must know the temporal variation of fluid flow rates and direction; for pressure solution models, one must know the temporal variation of the pore pressure.

In an alternate model, quartz dissolves at clay or mica surfaces of stylolites, silica diffuses to interstylolite areas, and quartz cement precipitates on quartz grains (Oelkers et al., 1992, 1993, 1996; Bjørkum, 1994, 1996; Walderhaug, 1994a, 1996). This model, which has much in common with the mechanism of quartz cementation suggested 40 yr ago by Heald (1955), is supported by a variety of petrographic observations and modeling results (Bjørkum, 1994, 1996; Walderhaug, 1994a, 1996; Aase et al., 1996; Oelkers et al., 1996). The model permits cementation and porosity variation to be computed as a function of time, temperature, and depth using primarily laboratory measured and theoretically generated parameters. Consequently, the model is well suited for predicting porosity during the early stages of exploration, when little formation-specific data are available. The model has been successful in representing porosity and porosity heterogeneity as a function of depth and stylolite occurrences in North Sea sandstones (Oelkers et al., 1996). In our study, we use the clay-induced dissolution and local silica redistribution model with provisions for mechanical compaction during early stages of burial to provide a set of porosity predictions for quartzose sandstones.

**REVIEW OF POROSITY REDUCTION MECHANISMS**

Efforts in our study are limited to quartzose sandstones, where the effects of fluid flow and dissolution and precipitation reactions of aluminosilicate, carbonate, or highly soluble SiO$_2$ phases (e.g., amorphous silica or opal CT) on porosity are negligible. Three diagenetic processes are important in modifying porosity in quartzose sandstones (Houseknecht, 1987): (1) mechanical compaction, characterized by the reorientation and repacking of rigid grains and the deformation of ductile grains, (2) chemical compaction, specifically bulk volume reduction due to the dissolution of framework grains at points of contact, and (3) precipitation of quartz cement. The first of these porosity modifying processes is computed using the exponential mechanical compaction equation of Robinson and Gluyas (1992b) given by

$$
\phi_M = A \exp\left[-\frac{z}{B+Cz}\right]
$$

where $\phi_M$ refers to porosity of the sandstone experiencing only mechanical compaction, $z$ represents depth, and $A$, $B$, and $C$ designate empirical temperature-independent constants. Taking account of the porosities of a wide variety of sandstones, Robinson and Gluyas (1992b) proposed values of 0.5, 2.4 km, and 0.5 for $A$, $B$, and $C$, respectively, for quartzarkoses. The driving force for mechanical compaction, described in our study using this empirical expression, is overburden pressure (effective stress). The effects of temperature and time are considered negligible.

The effects of chemical compaction and cementation are assumed to be due to quartz redistribution in the sandstone. Following the clay induced dissolution model of Oelkers et al. (1992, 1993, 1996) and Bjørkum (1994, 1996), quartz is assumed to dissolve only at interfaces with mica or clay grains at stylolites within the sandstone. Aquous silica is transported by diffusion into the interstylolite region, where it is precipitated as quartz cement by kinetically controlled surface reactions. Fluid advection is assumed to be negligible in the silica transport process. This quartz redistribution process is outlined in Figure 1. The decrease in porosity due to chemical quartz redistribution stems from dissolution of quartz at stylolites leading to precipitation of cement in pore voids. Although small amounts of effective stress are required to ensure quartz grains remain in contact with mica or illite grains at stylolite interfaces, the overall rate of this chemical compaction and cementation process is controlled by the strong dependence of quartz dissolution or precipitation rate constants and aqueous diffusion coefficients on temperature.

The rate of silica redistribution in this system is computed by first taking account of the conservation of aqueous silica (SiO$_2$$_2$), which can be expressed as

$$
\frac{dm_{\text{tot, SiO}_2}}{dt} = \frac{dm_{\text{diff, SiO}_2}}{dt} + \frac{dm_{\text{rxn, SiO}_2}}{dt}
$$

where $\frac{dm_{\text{tot, SiO}_2}}{dt}$ refers to the change of the total aqueous silica concentration with time, and $\frac{dm_{\text{diff, SiO}_2}}{dt}$ and $\frac{dm_{\text{rxn, SiO}_2}}{dt}$ denote the change in aqueous silica concentration with time due to diffusional transport and chemical reactions, respectively. The change of aqueous silica concentration with time due to diffusional transport is computed using Fick’s law given by

$$
\frac{dm_{\text{diff, SiO}_2}}{dt} = \frac{D_{\text{SiO}_2}}{\tau} \frac{d^2 m_{\text{SiO}_2}}{dz^2}
$$
where $D_{SiO_2}$ denotes the diffusion coefficient of aqueous silica in pure solution, $\tau$ represents a tortuosity factor, and $z$ refers to distance in the $z$ direction. Quartz dissolution or precipitation is the only reaction considered. Quartz dissolution or precipitation rates are computed with an equation based on transition state theory of the form (Rimstidt and Barnes, 1980; Aagaard and Helgeson, 1982; Murphy and Helgeson, 1989; Berger et al., 1994)

$$\frac{dm_{qtz}}{dt} = k_{qtz} s_{qtz} \left(1 - \frac{c_{SiO_2}}{K_{qtz}}\right)$$

(4)

where $m_{qtz}$ designates the moles of quartz dissolved (which may be positive or negative), $k_{qtz}$, $s_{qtz}$, and $K_{qtz}$ are the rate constant, mineral surface area, and equilibrium constant for quartz dissolution, respectively, and $c_{SiO_2}$ refers to the concentration of aqueous silica.

Given the temperature and burial histories of a sandstone of known stylolite density, calculating chemical compaction and cementation within this model requires quartz dissolution and precipitation rate parameters, aqueous silica diffusion coefficients, and the chemical driving force of the process, each of which is independently characterized. Consequently, the model can be used to compute the extent and consequences of the quartz redistribution process over a wide variety of conditions. Calculations were performed using the temperature-dependent equilibrium constants (at hydrostatic pressure) obtained using the SUPCRT92 computer code (Johnson et al., 1992) and the temperature-dependent quartz dissolution rate parameters reported by Gislason et al. (1997). In accord with the distribution of quartz cement as a function of interstylolite distance and temperature in North Sea sandstones, the precipitation rate constant is set equal to 0.008 times the corresponding dissolution rate constant. Temperature-dependent diffusion coefficients for aqueous silica are assumed to be equal to those of NaCl electrolyte computed from equations reported by Oelkers and Helgeson (1988); tortuosities are generated from Archie’s law (Archie, 1942; Berner, 1980). The surface area for quartz dissolution at stylolites is set equal to 1 cm$^2$/cm$^2$ interface, and the surface area available for precipitation is taken to decrease linearly with porosity ($\phi$) (Oelkers et al., 1997), according to

$$s_{qtz} = A(\phi - 0.035)$$

(5)

where $A$ refers to a geometrically obtained constant equal to 200 cm$^2$. According to equation 5,
quartz surface area tends to zero as porosity tends to 3.5%. At such low porosities, pore throats may become obstructed and a few isolated pores may remain after most of the sandstone is totally quartz cemented.

The rock is assumed to consist of a quartzose sandstone containing parallel equally spaced stylolites and, consistent with equation 1, to have a porosity of 0.5 at \( z = 0 \). Calculation of porosity as a function of time along a specified time, temperature, and burial rate trajectory is performed by simultaneously considering the effects of mechanical compaction and quartz redistribution on porosity. The effects of mechanical compaction are computed using equation 1. Based on petrographic observations and computational results, quartz redistribution stemming from clay-induced dissolution is not significant at temperatures below 80°C (Konnerup-Madsen and Dypvik, 1988; Glasmann et al., 1989; Ehrenberg, 1990; Walderhaug, 1990, 1994b; Robinson and Glayas, 1992a; Walderhaug and Fjeldskaar, 1993). Mechanical compaction is the dominant process at low temperatures. Due to the increased strength of the sandstone following precipitation of quartz cement, mechanical compaction is assumed to be arrested by the quartz redistribution process. Consequently, the mechanical compaction was assumed to stop once 0.5 vol. % of the sandstone was filled with precipitated quartz cement. Porosity loss due to quartz redistribution is computed using equations 2–4 by a finite-difference algorithm. Aase et al. (1996) suggested that the inhibiting effect of microquartz coatings on the formation of normal syntaxial quartz overgrowths is due to the microquartz crystals being approximately 1% more soluble than larger quartz crystals; that is, the resulting increased silica saturation in the pore water is large enough to inhibit quartz dissolution at stylolites. To be consistent with these observations, the equilibrium constant for the quartz dissolution reaction at stylolite interfaces was set to be 1–1.2% greater than that of macrocrystalline quartz.

The illustrative calculations in the following paragraphs were generated assuming a constant temperature gradient and burial rate, although any temperature gradient and burial history can be accounted for in the model. The burial rate of a sandstone having an age designated by \( a \), which is currently at a depth of \( z_d \), is equal to \( z/a \). As a consequence, the depth of the sandstone as a function of time is given by

\[
z_t = \frac{z}{a} t
\]

where \( z_t \) refers to the depth of the sandstone at any time \( t \), which is assumed to be zero when the sandstone was deposited. The temperature of the rock as a function of time is therefore

\[
T_t = \frac{\partial T}{\partial z} z_t = \frac{\partial T}{\partial z} \left( \frac{z_d}{a} t \right)
\]

where \( T_t \) designates the temperature of the rock at time \( t \), and \( \partial T/\partial z \) is the temperature gradient. Because the porosity can be a function of distance from stylolites (Oelkers et al., 1996), the porosities reported below are spatially averaged porosities (\( \phi_{sa} \)) generated from

\[
\phi_{sa} = \frac{1}{x_s} \int_0^{x_s} \phi dz
\]

where 0 and \( x_s \) are the positions of stylolites.

Figure 2—Computed variation of porosity with depth for Jurassic (170 Ma) sandstones for basins having temperature gradients of 80, 50, 35, and 20°C/km.
Results presented in the following paragraphs are consistent with those of Walderhaug (1996), who used the same conceptual model we use to compute the effects of quartz distribution on porosity in sandstones. In his study, Walderhaug (1996) assumed that the rate of quartz cementation is limited by the rate of its precipitation on quartz grains, which was calibrated empirically using typical sandstone with a representative interstylolite spacing. A mathematically simple set of equations that could accurately account for the effects of grain size, temperature history, and the presence of grain coatings that inhibit cement nucleation was generated. In contrast, in our mechanistic model, rates are computed through the simultaneous consideration of the dissolution, diffusion, and precipitation subprocesses. The resulting model is more complex mathematically but, as a consequence, can accurately account for the heterogeneous porosity distribution between stylolites and can incorporate phenomena that affect dissolution and transport, such as stylolite density, silica activity, and the presence of hydrocarbons in pore voids.

**COMPUTATIONAL RESULTS**

**Variation of Porosity With Thermal Gradient**

Computed porosities as a function of depth are illustrated for Jurassic (170 Ma) sandstones in Figure 2. These results were generated assuming a 20/m stylolite density, a relatively common stylolite density on the Norwegian shelf. The curves in this figure correspond to constant basin thermal gradients ranging from 20 to 80°C/km. All the sandstones exhibit identical porosity/depth trends until the onset of the silica redistribution

![Figure 3](image-url)
process, which is indicated in Figure 2 by the departure of the porosity/depth curve from the mechanical compaction curve. Following the onset of chemical compaction due to quartz cementation, the computed porosities strongly depend on the temperature gradient as a consequence of the silica redistribution process, which is controlled by the relative variation of kinetic dissolution and precipitation rate constants and the diffusion coefficient. The rate constants increase more strongly with temperature than the diffusion coefficients, but all constants are far stronger functions of temperature than of pressure. As a result, the amount of porosity persisting at any given depth decreases dramatically with increasing temperature gradient. For example, at a depth of 3.5 km, porosities of 22, 18, and 3.5% were computed for sandstones in basins having temperature gradients of 20, 35, and 50°C/km, respectively.

Variation of Porosities With Rock Age

The computed variation of porosity with rock age can be assessed by comparing Figure 2 with Figure 3, where the results of corresponding calculations for Miocene (25 Ma) (Figure 3A) and Cretaceous (100 Ma) (Figure 3B) rocks are depicted as a function of depth. This comparison reveals that for a given depth and temperature gradient, the porosity decreases significantly with age. For example, for a sandstone at a depth of 4 km in a sedimentary basin having a temperature gradient of 35°C/km, the porosity is 20% for Miocene sandstones and 12% for Jurassic sandstones. This difference stems from the low quantity of precipitated quartz cement in the younger rocks, which have a slightly lower porosity prior to the arresting of mechanical compaction, but a higher overall porosity. This observation is consistent with the conclusions of Scherer (1987) who found a strong
Figure 6—Comparison between measured and independently calculated quartzose sandstone porosities as a function of depth. The outlined area shows measured porosities, and the solid line corresponds to predicted porosities (see text) for various locations.
inverse correlation between rock age and sandstone porosity.

A direct comparison of the variation of porosity with depth for sandstones of different ages but all from basins with a temperature gradient of 35°C/km is illustrated in Figure 4. This figure illustrates the predicted effect of age in controlling porosity. Economically significant levels of porosity apparently can persist in rocks to great depths and temperatures if the rocks are sufficiently young. Both quartz dissolution at stylolite interfaces and the precipitation of cement are controlled by kinetically controlled processes, and cementation is thus time dependent.

**Variation of Porosity as a Function of Stylolite Frequency**

The effect of stylolite density on computed porosities as a function of depth has been computed for various sandstone ages and temperature gradients. Representative results for Jurassic sandstones in a basin with a 35°C/km temperature gradient having stylolite densities of 2, 20, and 200/m are shown in Figure 5. Computed porosities increase with decreasing stylolite density because the stylolite-quartz interfaces are the source of the precipitated cement. For example, the computed porosities at a depth of 4 km are 12, 18, and 10% for stylolite densities of 2, 20, and 200/m, respectively. From the results depicted in Figure 5, one can infer that the effect of stylolite density is greatest at low densities. In these extreme cases, the quartz redistribution process is source limited. In contrast, at high stylolite densities the effect of increasing stylolite density becomes small because the quartz redistribution process tends to be precipitation controlled, and is effectively independent of stylolite density at stylolite densities greater than approximately 40/m.

**Comparison With Field-Measured Porosities**

Several comparisons of computed and measured porosities are presented in Figure 6. All computed results, represented by solid curves in this figure, were generated assuming a constant typical initial stylolite density of 20/m. The outlined area in Figure 6A represents the porosity trend reported by Taylor (1990) for Miocene sandstones from the Corsair trend, Brazos area, offshore Texas. The present temperature gradient of this region is approximately 32°C/km (Bodner and Sharp, 1988). The solid line in Figure 6A represents the porosity calculated for a 25-Ma rock in a basin with a 35°C/km temperature gradient. The outlined area in Figure 6B shows the porosity vs. depth trend for Jurassic Haltenbanken sandstones reported by Bjorlykke et al. (1989). Geothermal gradients in this area are approximately 38°C/km (Ehrenberg, 1990). The solid curve in this figure represents results of calculations generated for a 170-Ma sandstone in a basin with a 35°C/km temperature gradient. The outlined area in Figure 6C represents mean porosity values as a function of depth for Oligocene (30 Ma) southern Louisiana sandstones reported by Selley (1978). Geothermal gradients in this area are in the range 22 to 31°C/km (Maxwell, 1964; Kumar, 1977). The solid line in this figure was computed for a 30-Ma sandstone in a basin with a 25°C/km temperature gradient. The outlined area in Figure 6D depicts the porosity of the Jurassic Ness formation as a function of depth reported by Giles et al. (1992). Geothermal gradients in the North Sea mostly range from 30 to 40°C/km (Cornford, 1984). The solid curve in Figure 6D illustrates porosity calculations for a 170-Ma sandstone in a basin with a 35°C/km temperature gradient. Correspondence between the mean measured porosities at any depth and the independently calculated porosities in each of the diagrams presented in Figure 6 is generally good, within 5% (5 porosity units) or less. Although empirical relationships in any given basin and reservoir commonly can predict porosity to about 2% (2 porosity units) (Bloch, 1991), the fact that our model calculations can reproduce the porosities in a wide variety of sandstones from sedimentary basins of different ages and temperature gradients supports the use of our computational model as a general predictive tool. Furthermore, variations in thermal histories, sandstone composition, grain size, grain coatings, and the possible effects of early overpressure on mechanical compaction have not been considered here. Such information is required to achieve a better degree of fit for any particular reservoir and basin situation.

**Variation of Porosity as a Function of Hydrocarbon Saturation**

Calculations performed to assess the effects of the presence of hydrocarbons on the porosity evolution of sandstones were based on the following assumptions: (1) the surfaces of the minerals are water wet, and (2) the role of the hydrocarbons on the quartz redistribution process is to take up space in the pore voids, thus lowering the percent of aqueous fluid available for diffusional transport of aqueous silica and increasing the tortuosity in accord with Archie’s law. In addition, the percent of hydrocarbon fill in the pore voids is assumed to
be constant with time. This latter assumption is not essential, but is taken to provide illustrative results. Computed results generated for Jurassic sandstones in a basin with a 35°C/km temperature gradient and a stylolite density of 2/m and an 85% hydrocarbon fill are compared to corresponding results obtained in hydrocarbon-free sandstones in Figure 7. The presence of hydrocarbons preserves a significant amount of porosity in this sandstone, which has a large interstylolite spacing. The result is a consequence of the decreased diffusive flux of aqueous silica to the interstylolite region. Similar calculations, however, show a decreasing effect of the presence of hydrocarbons with increasing stylolite density, where effects of diffusion become minimal as a rate limiting process. For example, less than 1% porosity difference is found in the sandstones whose pores are hydrocarbon free and sandstones whose pores are 85% hydrocarbon filled for all depths for a Jurassic sandstone in a basin with a 35°C/km temperature gradient and a stylolite density of 20/m. This observation is consistent with the results of Murphy et al. (1989), who noted that diffusional control of geological process is greatest for systems with long system lengths requiring the furthest transport distances. Similar computed results were obtained for sandstones of different ages and in basins having different temperature gradients. Modeling results show that sandstones with a stylolite density of 20/m or more exhibit little porosity difference in response to the presence of hydrocarbons. These results are consistent with the observations of Walderhaug (1990) of quartz cementation during hydrocarbon emplacement, and of Giles et al. (1992), who observed no difference in the porosities of hydrocarbon-filled vs. hydrocarbon-free North Sea sandstones. The results presented here also suggest that hydrocarbon reservoir sandstones are predominantly water wet.

CONCLUSIONS

Our calculations represent a general set of computational results on the variation of sandstone porosity as a function of depth, age, basin temperature gradient, stylolite density, and hydrocarbon saturation. Computed results indicate that high degrees of porosity are favored in younger rocks, in cool basins, and for sandstones having low stylolite densities; nevertheless, significant amounts of cementation can occur in young sandstones if they are rapidly buried in basins with high thermal gradients. The presence of hydrocarbons was found to preserve porosity only in sandstones having atypically low stylolite densities.

The model presented in our study offers numerous advantages over previous empirical, fluid advection, or pressure solution models. First, our model is consistent with recent estimates of the quantity of advective flow in sedimentary basins, and petrographic observations demonstrating that pressure is not the rate controlling parameter of quartz dissolution and cementation in sedimentary sandstones. Because it is process based, our model is generally applicable to a wide variety of sandstones and sedimentary basins having different temperature/depth histories. Because it requires no prior knowledge of either fluid flow rates or pore fluid pressures during basin evolution, our model can be used to make porosity estimates in potential petroleum reservoirs during the initial stages of exploration, when only limited information and data are available. As further basin specific information becomes available, it can be readily incorporated.
into model calculations for improved porosity predictions.

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