Compaction of Sand and Clay - Constraints from Experimental Compaction, Chemical Reactions and Fluid Flow During Burial - an Overview*

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Abstract

For petroleum exploration and production it is important to be able to make predictions about rock properties before of drilling. The rock properties depend on the distribution of sedimentary facies and primary mineralogical and textural composition, both for sand and mud. This is the starting point before the mechanical and chemical changes which have occurred during burial after deposition (diagenesis) (Figure 1). Reconstructions of sedimentary facies based on seismic interpretation and well log data have been highly successful in petroleum exploration and has developed into a well-established discipline presented in several textbooks. The emphasis has, however, been mostly on sand and less on clay. The post depositional processes (diagenesis) has been considered a discipline for specialists in the petroleum industry and at many universities sedimentology does not include sedimentary petrology, mineralogy and diagenesis. The composition and properties of shales has also been an area for specialist in clay mineralogy. The properties of clay, mudstones and shale vary greatly depending on the mineralogy and the textural composition. This is important for the function of shales as source and cap rocks and also for compaction and basin subsidence.

There is a need for a more integrated basin analysis including provenance, transport, deposition and burial and compaction of sand and clay. Data from experimental sediment deformation (compaction) should also be included. Diagenetic reactions in sandstones and mudstones are in principle well understood, but it is difficult to make predictions about changes in rock properties as a function of burial in sedimentary basins. This is because of the great variations in the initial sediment composition, depositional environment and the burial history of the basins. Figure 2 shows the change in velocity during experimental compaction compared with log data showing natural mechanical and chemical compaction in Cenozoic and Cretaceous sediments from Haltenbanken, Offshore Norway. At the time of deposition, clastic sediments are composed of assemblages of grains which are mechanically unstable and in most cases also out of equilibrium thermodynamically. Diagenetic reactions are therefore driven towards reduced porosity (increased density) by mechanical compaction as a function of effective stress and chemical

compaction as a function of temperature and time. The chemical reactions involving solids, mainly minerals are driven towards a lower Gibbs Free energy for the mineral assemblage.

The mineralogy and grains size distribution is a function of the provenance, transport and the depositional environment. Biogenic components, mainly carbonate and silica may also be added causing important changes in the sediment composition. Post-depositional changes will take place due to mechanical compaction and reactions between minerals and pore water. In stationary pore water, transport of solids in solution is limited to diffusion. At shallow depth sediments may be flushed by meteoric water and cause dissolution of the most reactive minerals, such as feldspar, mica and carbonate. At greater depth (>10 -100 m) below the reach of meteoric water flow the transport of solids in solution is rather limited and the chemical reactions are nearly isochemical with limited potential for import or export of solids in solution by fluid flow. Transport by diffusion is controlled by the concentration gradients for the components dissolved in pore water which may approach equilibrium with common minerals like calcite. This reduces the potential for diffusive transport. The purpose of this presentation is to show how diagenetic processes depend on the primary mineralogy and textural composition of sand and clay and the effective stress and temperature during burial in sedimentary basins.

Sedimentary Facies and Provenance as a Starting Point for Burial Diagenesis

The sediment response to diagenetic processes depends on its initial mineralogical composition and primary textural characteristics. As these properties are inherited from the provenance regions, transport processes and the depositional facies, prediction of reservoir properties as a function of burial depth requires integration of sedimentology, sequence stratigraphy and facies analysis (Bjørlykke, 2015). The tectonic setting (e.g. stable craton interior, rift basin, converging plate margin) is therefore highly relevant when addressing reservoir quality in siliciclastic sandstones.

The sandstones of the Jurassic Brent Group from the North Sea Basin have a high content of feldspar and mica reflecting erosion from a granitic basement during rifting. Erosion of fine grained metamorphic rocks, basalts and gabbroid rocks is less likely to deposit reservoir rocks, but lithic fragments from such sources can still make up a significant fraction of the deposited minerals (e.g. in the southern North Sea and in the Barents Sea). The composition of mud also varies greatly ranging from clays with a high content of smectite, chlorite. Clays from granitic rocks contain mostly kaolinite which has different compaction properties. Sandstones with high concentration of stable framework grains (e.g. quartz) will likely preserve higher intergranular volume compared to sandstones with high content of lithic grains (e.g. mud rock fragments or ductile metamorphic or volcanic fragments). Grain sizes generally decrease with increasing transport distance from the provenance, resulting in textural variability along the source to sink profile (Figure 3). The distribution of feldspathic arenites and arkoses occur at relatively short distances from areas of faulting or tectonic uplift where erosion occurs rapidly relative to weathering.

Sedimentary facies will also control the distribution of sand and clay and their minerals and textural properties within a given sedimentary system. Lateral variability in reservoir quality within delta systems is related to the distribution of different facies. Moderately sorted, angular grains and little clay content generally characterizes the alluvial channel facies on the delta plain, whereas the delta front facies likely display higher degrees of sorting, rounding and mineralogical maturity due to reworking by waves. The dissolution capacity of the pore water is reduced when acidic freshwater is mixed with alkaline water from the ocean. Along this mixing zone, feldspar dissolution and kaolinite precipitation is reduced and the sediment might be more prone to early calcite cementation. The prodelta facies is typically more fine-grained and heterolithic in composition, which reduces the permeability of these deposits. Vertical variability in reservoir quality might relate to lateral shifts in delta lobes that develop over time (Figure 3). Understanding the characteristics and distribution of the different facies will aid in the development of reservoir models and in predicting reservoir quality prior to drilling.

Early (Near Surface) Diagenesis

Sediments potential to react with the atmosphere or water are highest close to the surface (0-10 m burial depth) both due to more efficient fluid flow and diffusion processes near the surface. A transition to reducing conditions (red/ox boundary) is usually found below the uppermost few centimeters of the seabed. Here sulfate in sea water will be reduced to H_2S which may dissolve metals and may be precipitated as sulfides. Near-surface diagenesis comprises reactions in sediments at relatively shallow depth and low temperatures. The sediment composition may in a marine environment change due to addition of biogenic carbonate and silica and authigenic minerals may precipitate near the sea floor. This may include carbonates, phosphates, glauconite, chamosite, sulphides and iron and manganese minerals. Organically produced carbonate or silica material is often unstable chemically and may later be sources of cement within the sediment during burial. Siliceous organisms composed of amorphous silica (opal-A) will recrystallize first to opal CT (opal consisting of clusters of stacked microcrystalline blades of Cristobalite and Tridymite) and then to quartz at higher temperatures. Organisms are to some extent incorporated within the deposited sediments, particularly in clay and mud, but sand may also contain significant amounts of biogenic material. Decomposition of organic matter consumes oxygen and strongly reducing conditions are usually found few centimeters below the sea (or lake) floor. Oxic respiration, denitrification and Mn and Fe reduction will take place quite fast during the upper few centimeters/decimeters of the sediment before most of the sulfate reduction and subsequently methanogenesis takes place. Sulfate reduction is limited by the diffusion of sulfate from sea water while methanogenesis in sands will continue further down as long as there is reactive material (organic matter) present in the sediment.

Meteoric Water Flushing

On land sediments are always exposed to air and rain water, but meteoric water may also reach marine sediments at considerable distances away from the shoreline. This depends on the rain fall and infiltration on land and surface, the permeability and the hydraulic head. The exposure to fresh water flushing (total flux) has an inverse relation to the sedimentation rate. Meteoric water is initially under-saturated with respect to all minerals when it infiltrates the ground. Meteoric water therefore causes dissolution of the most reactive minerals, such as feldspar, mica and carbonate. The degree of meteoric water flow is a function of climate (rainfall), the slope of the ground water table and the lateral continuity of sandstone aquifers. High flow rates causing significant leaching are mostly limited to shallow depth (<10 -100 m). Reactions dissolving feldspar and mica and precipitating kaolinite are driven by high H+/K+(Na+) ratios and low silica concentrations. Therefore, marine and brackish pore water will cause less dissolution compared to fresh water (Bjørlykke, 2014. Acid (H+) is provided in rainwater and decaying organic matter will contribute somewhat to acid production.

Meteoric Water Diagenesis as Evidence of Fluid Communication in Reservoirs

As meteoric water reacts with minerals like feldspar, mica and carbonate the rate of dissolution is reduced because the H+/K+(Na+) ratios in the pore water are decreased. In “normal” basins that are not affected by magmatic heat and hydrothermal fluid mobilization, feldspar dissolution and the presence of authigenic kaolinite is an indication of the meteoric water flux over time. The degree of dissolution, particularly feldspar and mica and precipitation of kaolinite, can therefore be used in the reconstruction of fluid flow pathways in reservoir sandstones. Dissolution caused by meteoric water flushing will cause dissolution (Figure 3 and Figure 6) and these secondary pores can in some cases be preserved to great depth (> 3 km) and form a significant part of the total porosity. Sand bodies deposited in a humid climate but with lower than normal feldspar dissolution and content of diagenetic kaolinite have been subjected to less meteoric water flow, and one explanation for this is limited fluid communication with the main sandstone bodies. This may be due to primary distribution of sand facies, but sand bodies may also have been detached by growth faulting at shallow depth avoiding much of the fresh water flushing. Permeable aquifers may be dead ends and effective groundwater flow must therefore be open to the seafloor. Deposition of sandstones in dry climates or deposition in a distal part of the sedimentary system without sufficient head to drive the meteoric water may also be reasons for limited meteoric water flushing.

Mechanical Compaction of Siliciclastic Sediments

Physical and elastic properties (e.g. porosity, density, velocity, bulk and shear moduli) of rocks change continuously from deposition through burial, uplift and reburial. At shallow burial depth, porosity loss is mainly mechanical in siliciclastic sediments due to grain reorientation, pseudo-plastic deformation of ductile grains and brittle grain fracturing (Figure 1). Mechanical compaction of siliciclastic sediments depends on mineralogy, grain size, shape, sorting and fluid composition (Chuhan et al., 2003; Mondol et al., 2007; Mondol et al., 2008; Fawad et al., 2011; Zadeh et al., 2016; Nooraipour et al., 2017).

Mechanical compaction is a loss of intergranular volume (IGV). The loss of IGV continues until quartz cementation prevents further mechanical deformation at about 60-80° C. From then on with further burial IGV is preserved and the compaction is mainly governed by pressure dissolution at stylolites and quartz cementation. Well data from Haltenbanken (mid-Norway) show that Pleistocene and Pliocene compact well because of poor sorting of partly glacially derived sediments, while Eocene mudstones with a high smectite content are characterized by high porosity and low velocity (Storvoll et al., 2006; Marcussen et al., 2009).

At 20 MPa effective stress (about 2 km burial depth) the porosity may vary by more than 10-15% before the onset of quartz cementation causing further porosity reduction. Chemical compaction results in a reduction in rock volume, but will unlike mechanical compaction not alter the intragranular volume much since cementation strengthen the grain framework more than required to compensate for the increase in lithostatic stress during deeper burial. The grains size and sorting can be linked to sedimentary facies and provenance. The composition of interbedded clay and mud can be predicted based on the depositional environment and on information from sandstone lithology. The degree of differential loading between sandstones and shales depend very much on the clay mineralogy.

Mechanical compaction can be reproduced in the laboratory and the porosity loss, velocity and density variations as a function of effective stress can be measured by varying the grain size, shape, sorting and mineralogy (Chuhan et al., 2003; Mondol et al., 2007; Fawad et al., 2011) (Figure 4). Compaction of clay in the laboratory shows that fine-grained clay like smectite has a very low compressibility compared to coarse grained clay like kaolinite (Figure 3c). Also, coarser-grained siliciclastic composites (both sands and clays) are more compressible compared to finer grained composites (Figure 3d-e). The coarser grains composites have lowest grain-to-grain contact compared to finer-grained composites and resulted in more compaction due to highest stress concentration per grain contact. Results of the experimental mechanical compaction compare well with natural compaction observed in sedimentary basins (Figure 5) (Mondol et al., 2008b; Marcussen et al., 2010, Fawad et al., 2010; Zadeh et al., 2016a.b).

Prediction of Rock Properties for Sandstones and Shales in Mechanically Compacted Domain

Velocity depth trends are the key to seismic modeling and also important in seismic processing (velocity analysis and depth conversion, NMO and DMO corrections) and inversion of seismic data. A Comparison of a standard velocity-depth trend with an actual velocity-depth trend may lead to a variety of new information. For example, the deviation of an actual velocity-depth trend compared to a theoretical or an empirical trend may indicate uplift/erosion, cementation, over-pressure, and/or presence of hydrocarbons (Figure 4). By utilizing a large sand experimental compaction database (Fawad et al., 2010; Fawad et al., 2011) we presented new empirical equations (Eqs. 1-3) for porosity and velocity (both Vp and Vs) prediction as functions of sand grain-size, shape, sorting and mineralogy.

$$\varphi = 0.4807-0.0023\sigma/v-0.1060\alpha-0.0098\beta-0.0812\gamma+0.0079\delta \ldots\ldots\ldots\ldots \quad (1)$$

$$Vp=2.9043+0.0073\sigma/v-2.2596\varphi+0.0764\alpha-0.0500\beta-0.0257\gamma-0.1271\delta \ldots\ldots \quad (2)$$

$$Vs=1.0461+0.0085\sigma/v-1.0553\varphi+0.00382\alpha-0.0481\beta-0.0157\gamma-0.0933\delta \ldots\ldots \quad (3)$$

where Vp is P-wave velocity, Vs is S-wave velocity, φ is porosity (fraction) calculated at an effective stress σ/v (MPa), α is the grain size (mm), β is the grain shape (fraction), γ is the sorting (phi scale), δ is the quartz percentage (fraction). These equations can help to predict sandstone properties (e.g. φ, Vp, Vs) at a certain depth as a function of provenance and sedimentary facies (Fawad et al., 2010).

Chemical Compaction at Deeper Burial

Chemical compaction involves dissolution and precipitation of minerals or other solids. The primary clastic and biogenic mineral assemblage is largely out of equilibrium with the pore-water while the driving force is due to thermodynamic instability (chemical affinity), the reaction rate (or kinetics of reaction) depends on mineral-water interactions (pore-water composition, minerals and mineral surfaces (reactive surface area), kinetic parameters (rate expression, rate constants, activation energy, pre-exponential factors, etc.), and temperature (Aagaard and Helgeson, 1982). Mineral nucleation is also prevented due to an activation barrier, dependent on substrate and temperature. Phases such as illite can therefore grow at lower temperatures (from e.g. smectite), but will only form at elevated temperatures (>120-130° C) if kaolinite is the substrate. Nucleation typically has higher activation energy than growth, and will be prevented to higher temperatures. The mineral reactivity is due to the primary mineral composition which depends on the provenance, transport and depositional environment. If the chemical composition of the bulk sediment volume remains constant, diagenetic processes can be written as balanced chemical equations (closed system). The reaction between K-feldspar and kaolinite precipitating illite, is an example of a closed system reaction:

$KAlSi_3O_8 + Al_2Si_2O_5(OH)_4 = KAl_2Si_2O_{10}(OH)_2 + 2SiO_2 + H_2O$ ………… (4)

The pore filling illite and the precipitated quartz has an important effect on reservoir quality, causing reducing the permeability (Figure 6). Dissolution of grains (mainly minerals) and precipitation of new minerals (cement) will be controlled by the kinetics (temperature) integrated over geologic time, restricted by the supply from the dissolving minerals (Bjørlykke and Egeberg, 1993). The rate of quartz cementation is controlled by temperature, time and the surface area available for quartz cementation (Walderhaug and Bjørkum, 1996). The source of quartz cement may be pressure dissolution at stylolites, or dissolution of unstable biogenic silica or volcanic fragments. Quartz cementation however also need surface area for nucleation and/or for syntaxial overgrowth, and grain-coating chlorites have been found to prevent both of these and preserve porosity to great depths (Jahren et al., 1998). A net increase in the porosity (secondary porosity) is only possible if the reaction products are removed by diffusion or fluid flow. The pore water will gradually approach equilibrium with most of the reactive minerals like carbonates and quartz, limiting the potential for mass transport by diffusion and fluid flow (advection) (Bjørlykke and Jahren 2012). Thermal convection is very limited in sedimentary basins because the critical Rayleigh number is rarely obtained. This is because shale layers will disrupt potentially large-scale circulation (Bjørlykke et al., 1988).

Constraints on Mass Transfer During Burial Reactions - Closed System Diagenesis

Below the reach of meteoric water flow and in “normal” basins that are not affected by magmatic heat and hydrothermal fluid mobilization, the flow of pore water is limited by the supply of pore water which is released by compaction and dehydration of minerals. The upwards component of the compaction driven flow is limited by the rate of loss of porosity in the underlying sediments and water from mineral dehydration. Normally this is smaller than the sedimentation rate and advective transport of solids in solution is therefore rather limited. The upwards flow of pore water may in many cases approach the sedimentation rate so that the pore water may remain at nearly constant depth (Bjørlykke, 1993). Pore water with high salinity and high density in the viscosity of evaporates tend to remain at a certain depth and is normally not driven towards the surface (Gran et al., 1993). Salinity gradients in pore water may therefore put important constraints of fluid flow in sedimentary basins.

As the pore water approach equilibrium with mineral phases such as carbonate and silicate, low concentration gradients limits element transport by diffusion. Creation of significant secondary porosity during burial diagenesis at 2-3 km burial depth is therefore not possible in the absence of fluid fluxes and because the pore water approach equilibrium with the reactive mineral phases. Chemical reactions at these depths are therefore nearly isochemical in the absence of hydrothermal intrusions. The requirement for large scale mass transfer and changes in the bulk composition as proposed by Day-Strirrat (2010) has not been quantified. When pore water is in in equilibrium with quartz at, i.e. 3.5 km depth and 120° C the silica concentration in the pore water is about 80 ppm and the silica concentration gradient is 3 ppm/100 m. An import of silica corresponding to 1% quartz cement by vertical advective flow through 100 m thick sandstone would then require a total pore water flow of 660 m^3/m^2 (Bjørlykke and Jahren, 2012). Import of silica as explanation for large-scale quartz cementation in sandstones is therefore unrealistic. In mudstones the potential for import of silica during burial is even smaller than for sandstones and quartz is precipitated from silica released by clay mineral reactions (Thyberg et al., 2009). Significant amounts of authigenic micro-quartz crystals form between about 65° C and 90° C and laterally extensive quartz platelets form above about 90C has been detected in mudstones (Thyberg and Jahren, 2011) Figure 6. The quartz cement platelets form preferentially oriented normal to the overburden reflecting incipient quartz schisocity in fine-grained mudstones buried to temperatures above about 90° C. The authigenic quartz cement is sourced from silica released from clay mineral reactions (smectite to illite above about 65° C and kaolinite to illite above about 130° C). Formation of authigenic quartz in mudstones and shales is consistent with chemically closed system burial diagenesis. Quartz is precipitated as sheet-like cements producing a pronounced cleavage (Thyberg and Jahren, 2011). Quartz cementation in mudstones change the rock properties from soft mud to hard shales with higher velocity and lower density (Thyberg et al., 2010).

Diagenesis During Uplift and Erosion (Unloading)

During progressive burial chemical compaction due to quartz cementation is an exponential function of temperature integrated over geologic time (Walderhaug and Bjørkum, 1996). During uplift the reactions like quartz cementation will continue at a reduced rate and cooling as long at the temperature remains above about 70° C. Mechanical compaction is not important for well cemented rocks during uplift, and the solids (mainly minerals) will shrink slightly due to cooling and this is also the case for the fluids (water and HC). In a hydrocarbon phase a temperature drop may result in condensation of gas which could result in reduced fluid pressure, in some cases under-pressure. At shallow depth, extensional fractures may develop depending on the horizontal stress distribution, but they are unlikely to be conduits for downwards flow of meteoric water unless there is a hydrothermal drive for fluid flow.

Conclusions

Rock properties at a certain burial depth and temperature can to a large extent be predicted if the primary depositional environment and the provenance as a part of a basin analysis. Unfortunately, many sedimentological descriptions of outcrops and cores do not include detailed analyses of the mineralogy and textural relationships. This is particularly true of mudstones and shales. Prediction of reservoir properties before drilling requires facies models for the distribution of sandstones and their primary properties such as mineralogy, grain size and sorting. We also need to understand changes in rock composition during early diagenesis and the reactions during deeper burial. Below the reach of meteoric water flow the pore water flow limited constraining transport of solids in solution and the potential for creating net secondary porosity or cementation. Accepting nearly isochemical burial diagenesis, predictions about rock properties as a function of burial depth can be made based on assumptions about the composition at shallow depth.

The initial sediment composition is a main factor when modeling diagenesis and porosity. An integrated basin analysis of sediments from source to deposition, burial and uplift is therefore required. This means that sedimentological studies should have more emphasis on the mineral composition of both sand and clay. Experimental compaction of sand and clay is important in the prediction of compaction processes (porosity) in sedimentary basins. Chemical compaction is a function of thermodynamics and kinetics and temperature integrated over geologic time and is therefore difficult to study experimentally. Log data and mineralogical studies of samples from subsiding sedimentary basins like the North Sea provides a good data base for calibrating burial diagenetic reaction for different initial lithologies.

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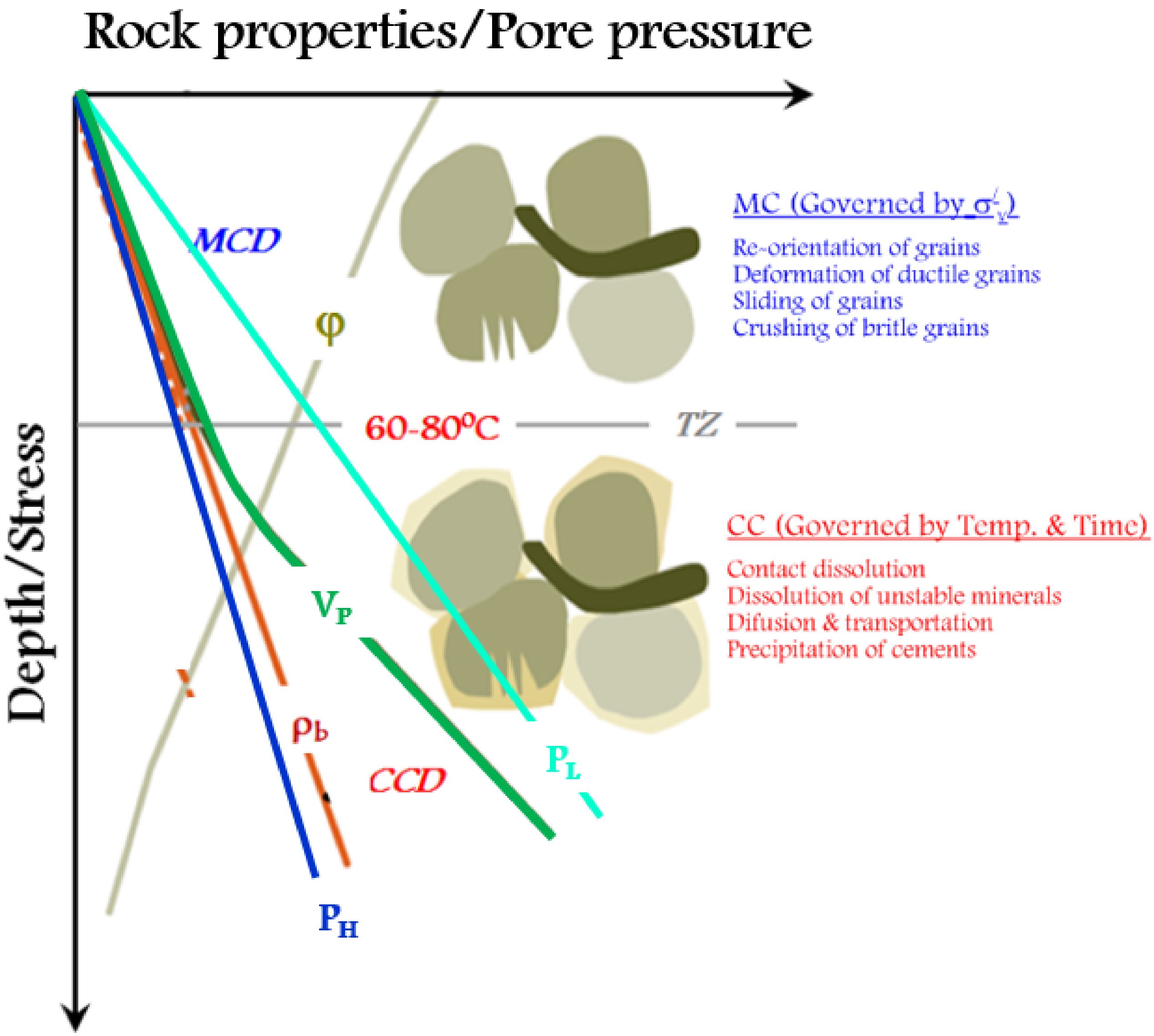

Figure 1. Schematic representation of compaction processes of siliciclastic sediments. Sediment compaction in mechanical compaction domain (MCD) is mainly governed by the effective stress where mineralogy, size, shape and sorting play important roles to control rock properties such as porosity (ϕ), density (ρb) and P-wave velocity (Vp). In normally compacted basin both lithostatic pressure (PL) and hydrostatic pressure (PH) increase as a function of depth. In the chemical compaction domain (CCD), temperature and time are the main driving forces to compact sediments. The two compaction domains are separated by a transition zone (TZ) at temperatures ~60-800° C where both mechanical and chemical compaction processes can work simultaneously.

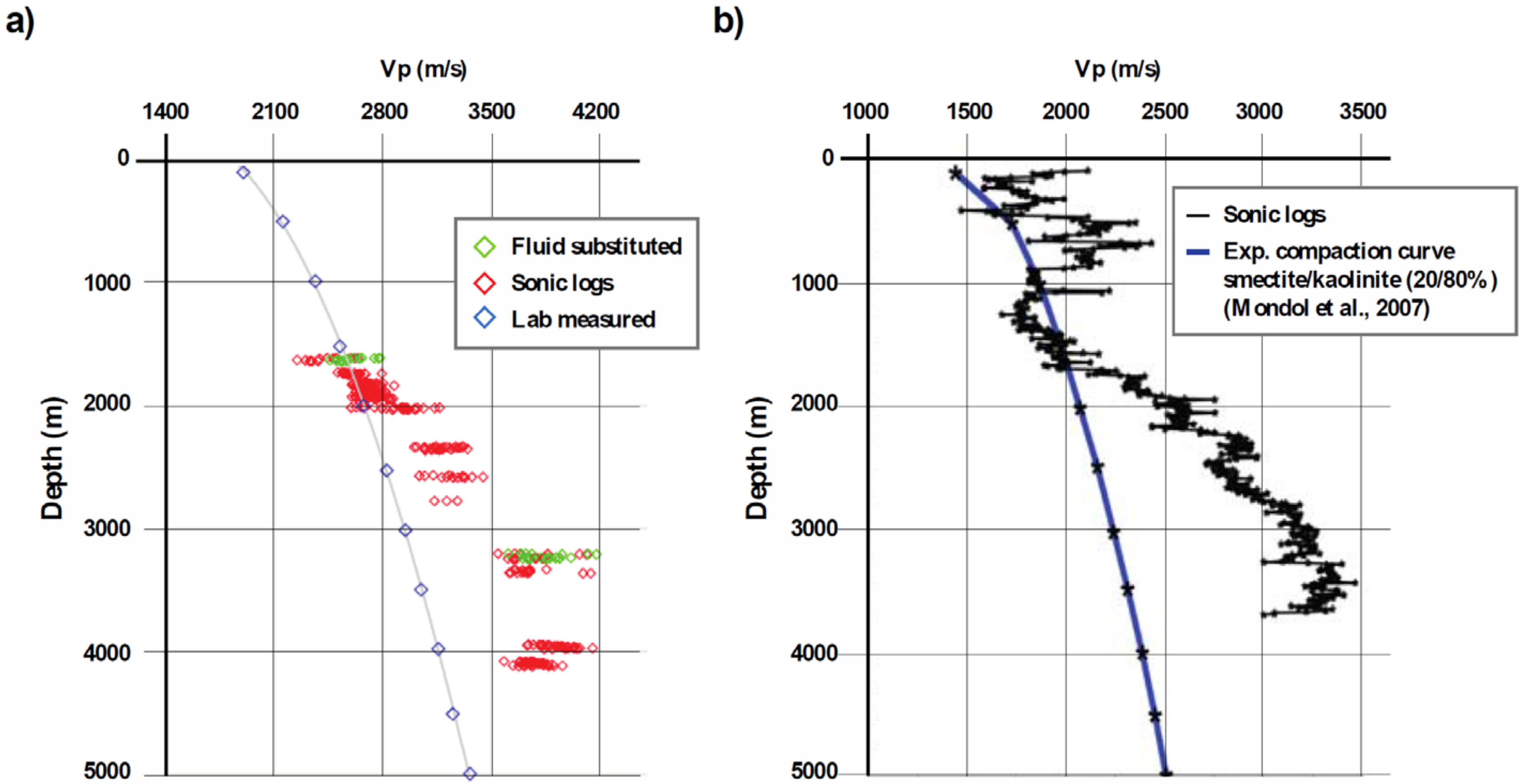

Figure 2. (a) Experimental compaction of loose sand from the Etive Formation, North Sea (blue curve, after Marcussen et el., 2010), compared to well log data (red dots). Deviations from the mechanical compaction trend are interpreted to represent the effect of chemical compaction (quartz cement). (b) The experimental compaction curve (blue line) documents the gradual changes in petrophysical properties in mudstones (75% kaolinite/25% silt, after Mondol et al., 2008) as a function of increased stress. Petrophysical well log deviation from this trend is interpreted to reflect evidence of chemical compaction in naturally occurring mudstones from the Haltenbanken, Offshore Norway (6305/1-1) (after Peltonen et al., 2008). Glacially influenced sediments compact to high velocities and shallow depth, while smectitic clay at about 1.5 km shows less compaction and low velocities.

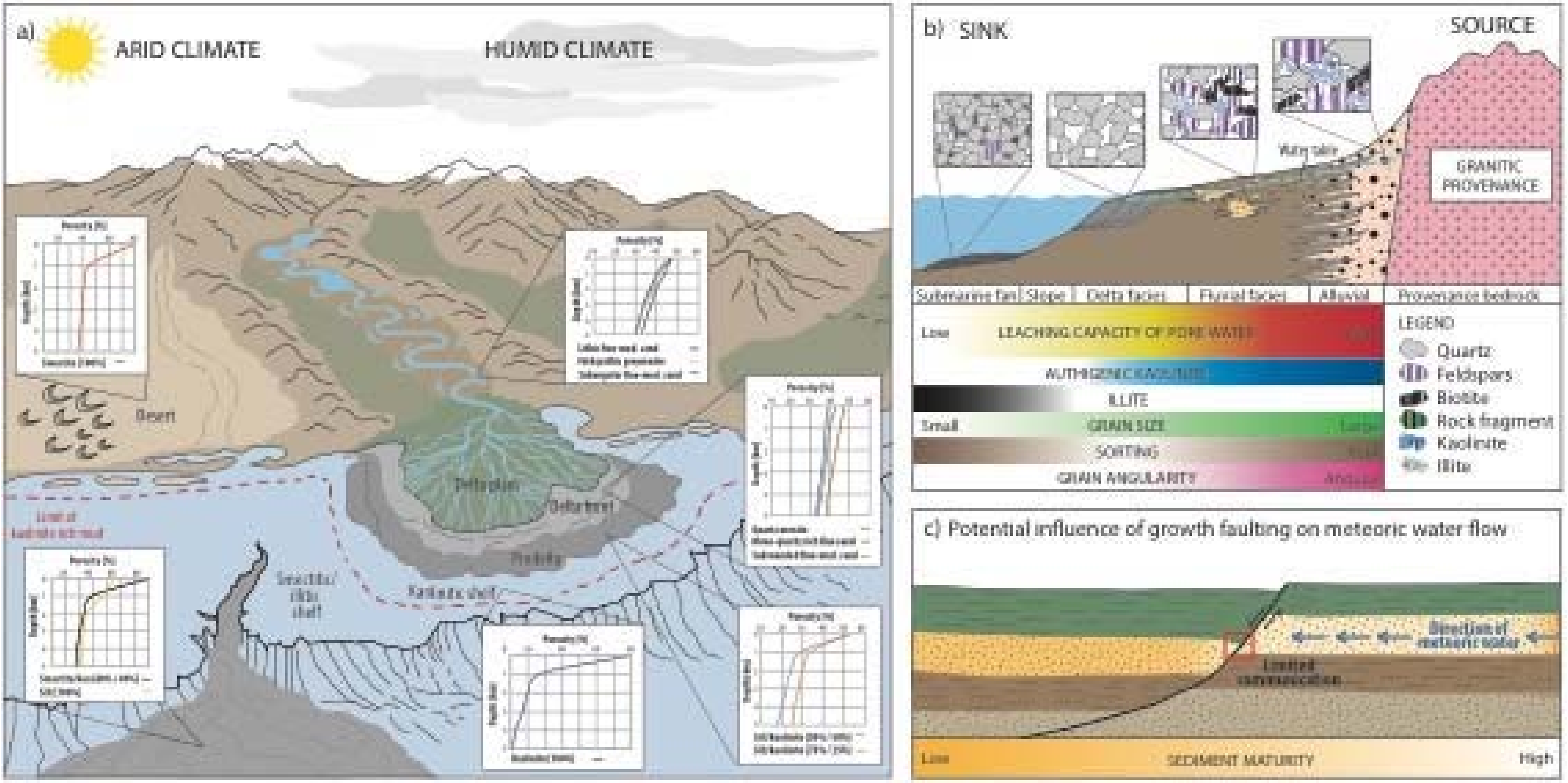

Figure 3. (a) Sediment properties and potential for meteoric water flow as a function of sedimentary facies and climate. Mechanical compaction trends vary as a function of sediment texture (grain size, sorting, and angularity) and composition. Distributions of these properties are governed by the provenance, climate and depositional facies. (b) Conceptual source-to-sink profile showing the lateral variation in leaching capacity of meteoric water and the resulting distribution of clay minerals as well as the textural properties of the sediment. Kaolinitic clay is most abundant in deltaic and proximal environments and smectite/illite in more distant facies. (c) The lateral connectivity between porous strata may be affected by growth fault tectonism and thus limit the flow of meteoric water though some of the sand bodies.

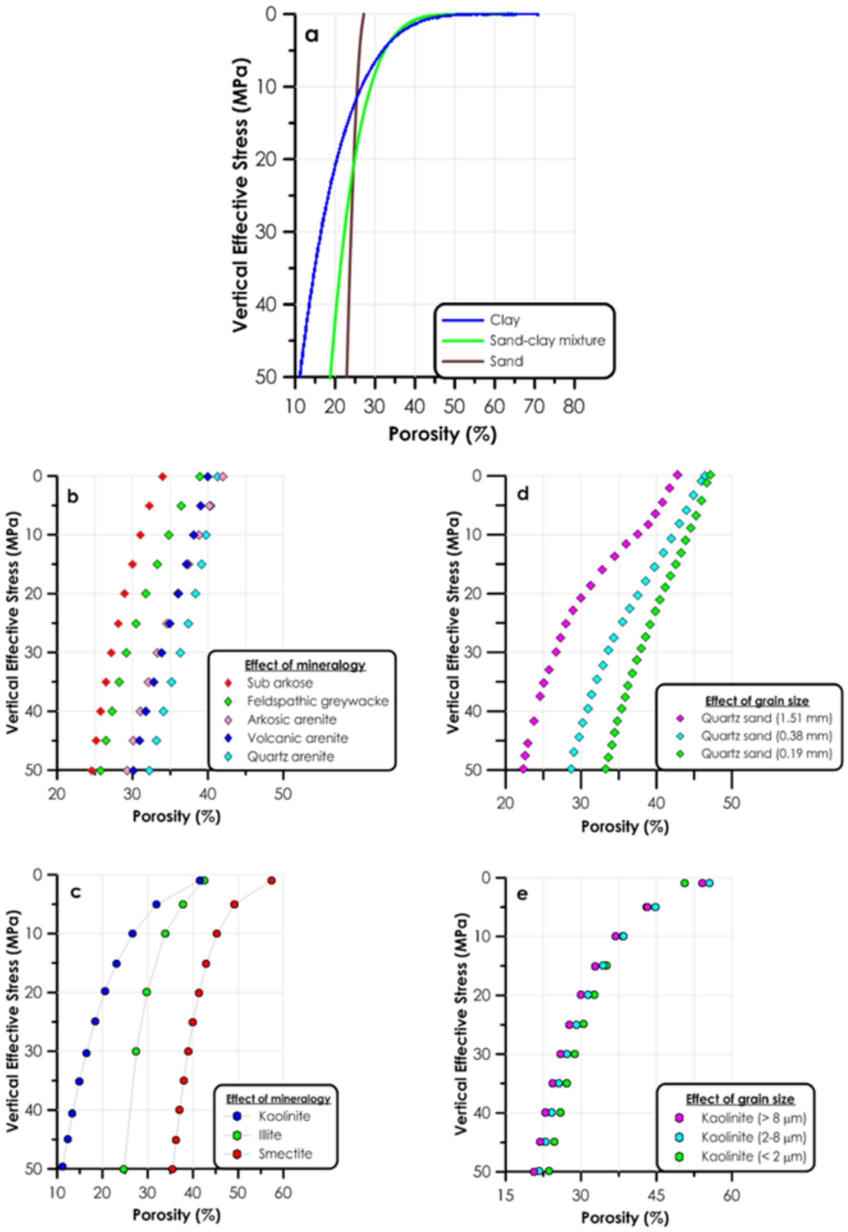

Figure 4. (a) Experimental mechanical compaction (i.e. porosity reduction as a function of effective stress) of sand, clay and sand-clay mixture. Usually clayey sediments have high initial porosity (~80%) at the time of deposition compared to sands (~40%) and sand-clay mixtures (50-60%). Mechanical compaction and rock properties of sands (b) and clays (c) are controled by mineralogy. Grain sizes also play significant a role in controling rock properties of sands (d) and clays (e) in the mechanical compaction domain.

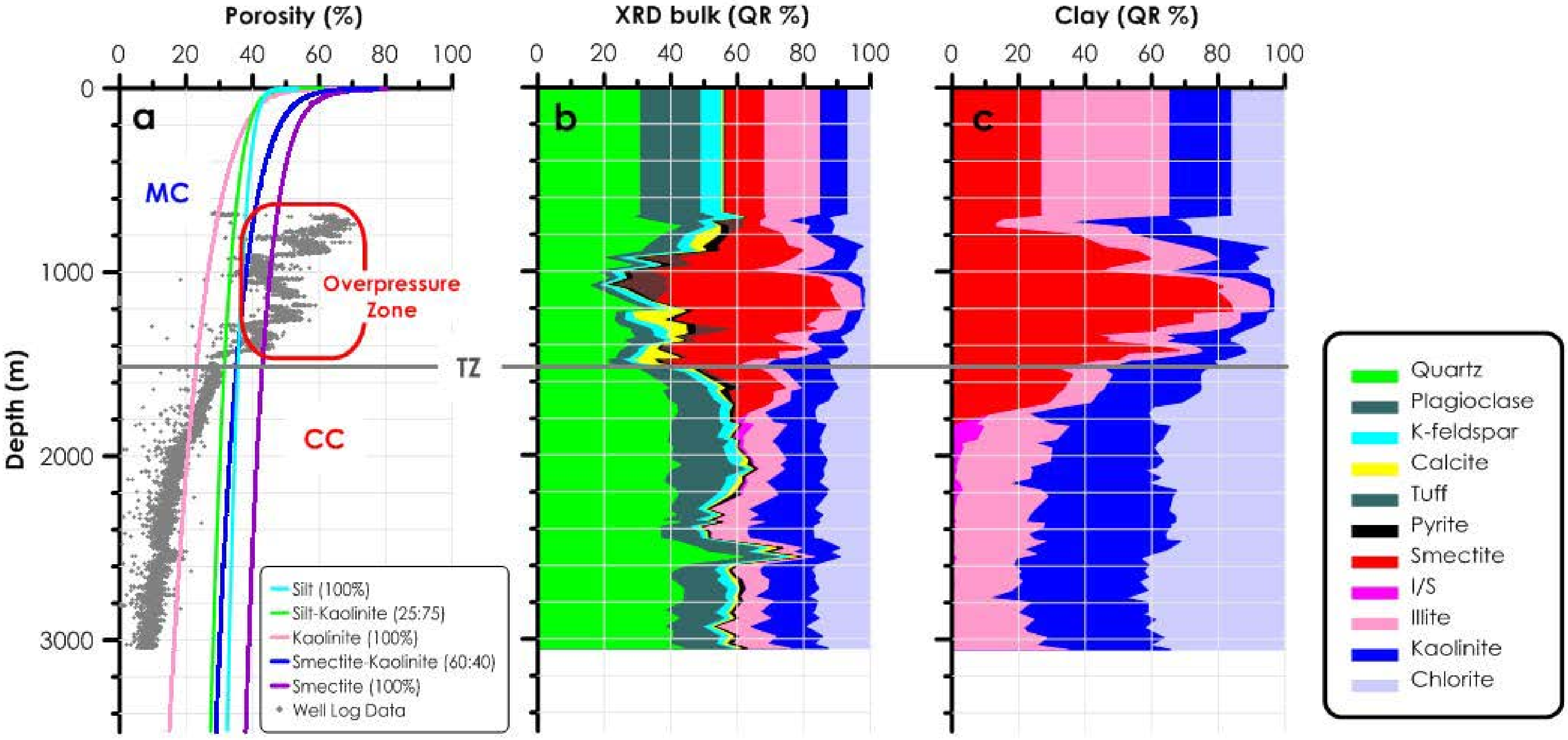

Figure 5. (a) Possible zone of overpressure development in a well indicated by the anomalous high porosity compared to the established range of compaction trends of synthetic mudstone. The high porosity zone overlaps the high smectite interval documented by (b) XRD bulk data, and (c) the clay fraction. Samples of cuttings are available only at depths from 700 m to 3100 m (Modified after Mondol et al., 2008b).

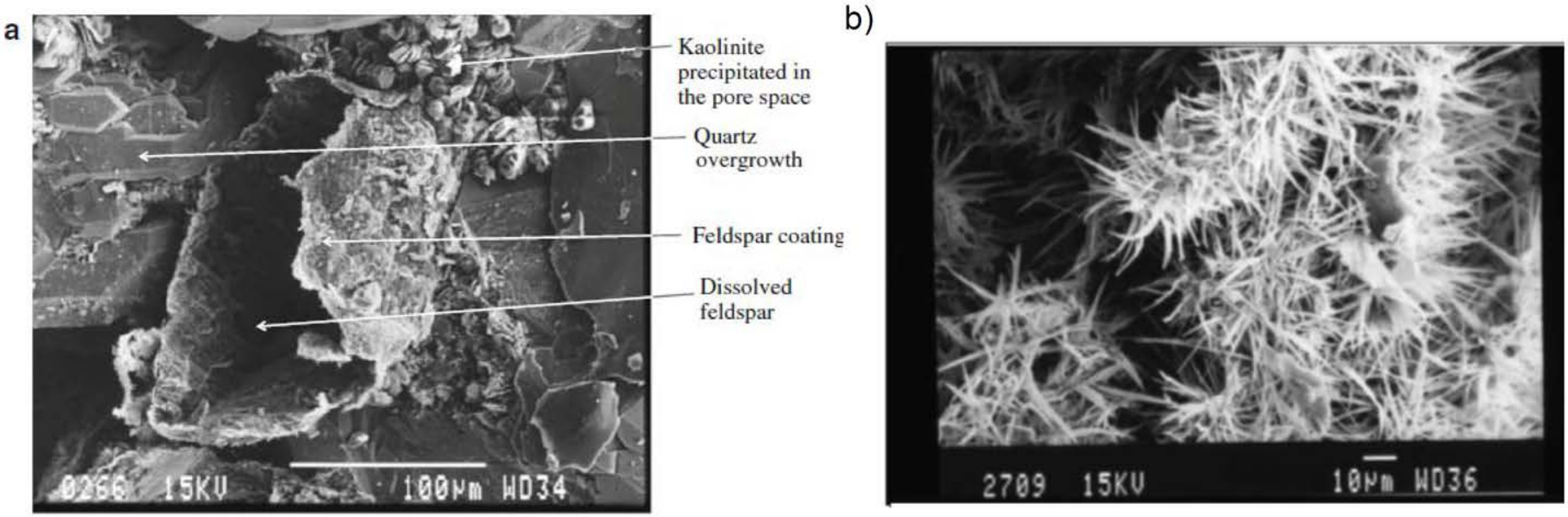

Figure 6. (a) Scanning electron microscope image of a sandstone (Brent Group) from the North Sea (from Bjørlykke, 2015). The scale is 0.1 mm (100 µm). In the center of the picture we see a cavity left by a dissolved feldspar grain. A clay rim around the feldspar remains undissolved, outlining the primary grain morphology. In the upper part, authigenic kaolinite crystals are forming small (10-20 µm) booklets. They have formed from the silica and aluminium released when the feldspar was dissolved by meteoric water. To the left, authigenic quartz is growing on clastic quartz. Note the relatively large pores between quartz and feldspar grains and the small pores between kaolinite crystals. (b) Pore filling authigenic illite from a Jurassic reservoir at 4.2 km depth at Haltenbanken, offshore Norway (Ramm et al., 1994). The transition from pore filling kaolinite to illite causes a marked reduction in permeability.

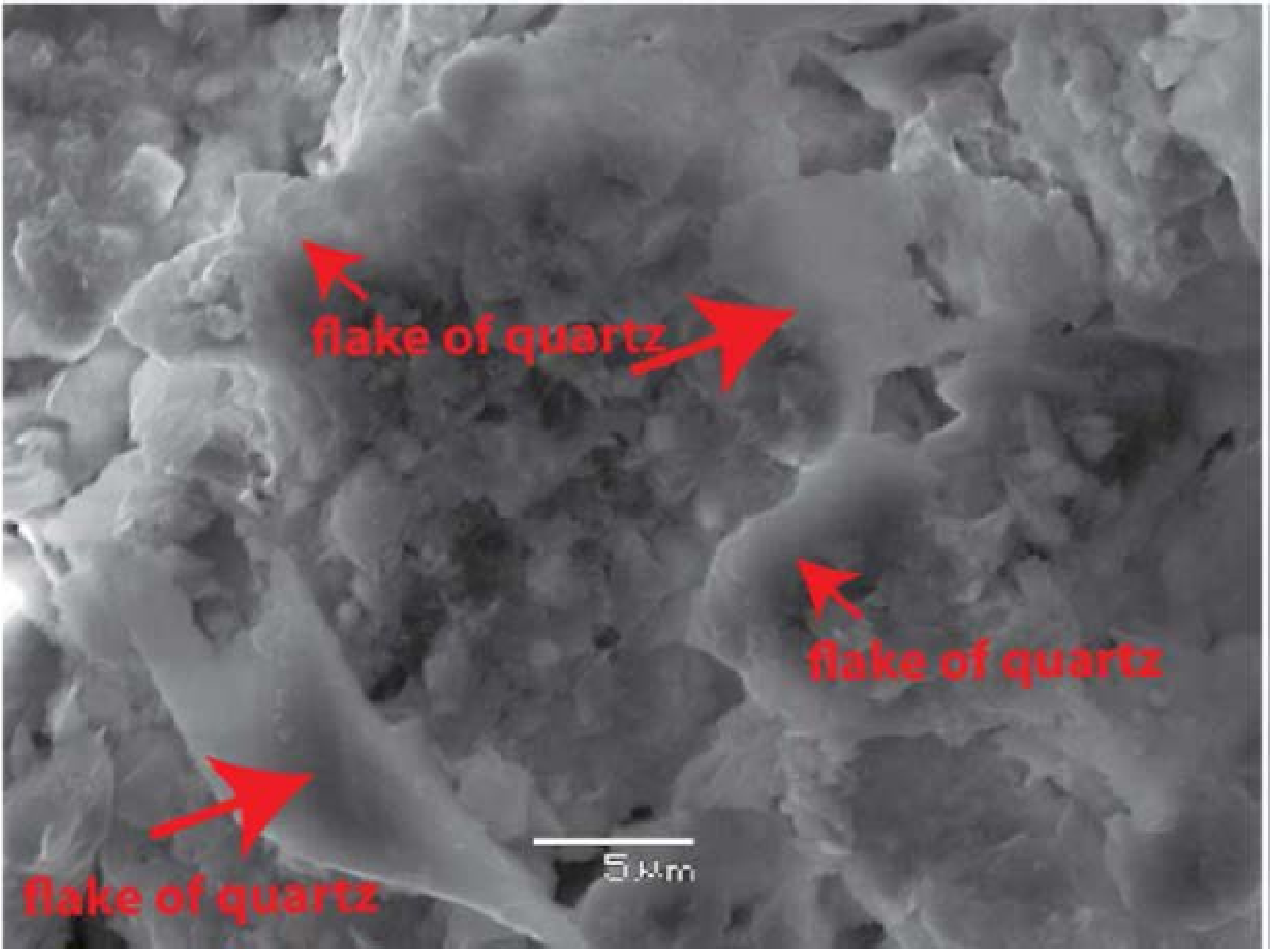

Figure 7. Evidence of continuous quartz cement growth in Late Cretaceous mudstones at 2610 m depth (90-95° C), well 6505/10-1, Haltenbanken, Offshore Norway (from Thyberg and Jahren, 2010).