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# DIAGENETIC REGIMES AND THE FORAMINIFERAL RECORD IN THE BEAUFORT–MACKENZIE BASIN AND ADJACENT CRATONIC AREAS

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**Abstract:** Terrigenous clastic, deltaic-dominated, sedimentary basins, such as the Beaufort–Mackenzie Basin of Arctic Canada, are thick, areally extensive bodies of sediment that undergo a wide variety of diagenetic processes in a wide variety of physical and chemical environments. Diagenetic processes and effects are unique to various parts of the basin, but four diagenetic regimes can be isolated to encompass many of the diagenetic processes affecting sedimentary basins. These are the early, burial, overpressured, and meteoric diagenetic regimes. Foraminiferal preservation may be affected by chemical, physical, and biological activities in the diagenetic regimes.

Early diagenetic conditions are generally favourable for the preservation of foraminifers, but bacterial production of $CO_2$ could lead to the dissolution of calcareous foraminifers and early diagenetic pyrite often infills foraminiferal tests.

The burial diagenetic regime is dominated by increasing temperature, pressure, and compaction. Increased temperature is reflected by thermal, colour alteration of organic cement in the test of agglutinated foraminifers. Thermally controlled mineralogical changes are also evident in the burial diagenetic regime and include silicification of the agglutinated test, chloritization of calcareous foraminifers, and precipitation of secondary clay minerals such as kaolinite, smectite, illite, and chlorite within the foraminiferal test. Thermal maturity can be assessed by application of the Foraminiferal Colouration Index (FCI). Mineralogical changes in foraminifers allow for the establishment of four broad burial diagenetic zones (A to D).

The overpressured regime may be responsible for a retardation in thermal alteration of agglutinated foraminifers and silicification of agglutinated foraminifers may be a precursor as well as an effect of overpressured fluids in sedimentary basins.

The meteoric regime is significant during periods of extensive erosion. Interactions between organic and inorganic detritus and meteoric waters may lead to dissolution of calcareous foraminifers and precipitation of secondary minerals such as kaolinite and minor amounts of quartz.

**Abstrakt:** Klastyczne, przeważnie deltowe, osady o dużej miąższości, deponowane w basenie sedymentacyjnym Beauforta–Mackenzie (arktyczna część Kanady) podlegały różnego rodzaju procesom diagenetycznym w warunkach szerokiej zmienności parametrów fizycznych i chemicznych. Procesy te i ich efekty są inne w każdej części basenu, jakkolwiek można wyróżnić cztery środowiska, które charakteryzują diagenezę w basenie sedymentacyjnym. Należą do nich procesy związane z: (i) wczesną diagenezą, (ii) pogrzebaniem osadów, (iii) nadciśnieniem płynów porowych oraz (iv) działalnością wód meteorycznych. Stan zachowania otwornic może zależeć od oddziaływania czynników chemicznych, fizycznych i biologicznych w czasie diagenezy osadów.

W warunkach wczesnej diagenezy, stan zachowania otwornic jest dobry, jakkolwiek biologiczna produkcja $CO_2$ może spowodować rozpuszczanie ich węglanowych skorupek. Skorupki są wtedy wypełniane wczesnodiagenetycznym pirytem.

W trakcie pogrzebania osadów wzrasta temperatura, ciśnienie i kompakcja. Wzrost temperatury powoduje zmiany składu mineralnego skorupek otwornic. Należą do nich sylifikacja skorupek otwornic aglutynujących, chlorytyzacja skorupek otwornic węglanowych i wytrącanie wtórnych minerałów ilastych (kaolinit, illit i chloryt) wewnątrz skorupek. Dojrzałość termiczną można określić stosując wskaźnik zmiany barwy otwornic (Foraminiferal Colouration Index). Zmiany mineralogiczne skorupek otwornic pozwoliły na ustalenie czterech stref przeobrażeń diagenetycznych (A–D).

Procesy związane z nadciścieniem wód porowych mogą być odpowiedzialne za spowolnienie procesów przemian termalnych w skorupkach otwornic aglutynujących. Sylifikacja skorupek może być zarówno prekursorem jak i pierwszym z efektów tych procesów.

Wpływ wód meteorycznych jest znaczący w czasie procesów silnej erozji osadów. Wzajemne oddziaływanie detrytusu pochodzenia organicznego i nieorganicznego z wodami meteorycznymi może prowadzić do rozpuszczania węglanowych skorupek otwornic oraz krystalizacji wtórnych minerałów (kaolinitu i podrzędne ilości kwarcu).

**Key words:** diagenesis, benthic foraminifers, sedimentary basins, thermal colour alteration, silicification, clay minerals, geopressure, meteoric water.

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# INTRODUCTION

Diagenesis refers to the physical and chemical changes undergone by a sediment, including its biological components, after initial deposition and continuing through lithification, exclusive of weathering and metamorphism. Diagenetic effects have been studied extensively by sedimentologists and geochemists in order to understand the modifications that sediments undergo on their path to becoming rocks, in particular as reservoirs for hydrocarbons. Paleontologists have studied early diagenesis mostly in a taphonomic context, being concerned with the early diagenetic history of an organism and effects on preservation. In contrast, the effects of late diagenesis (burial diagenesis) have received very little attention from micropaleontologists. Burial diagenetic effects, however, are conspicuous and accumulative in fossil foraminifers and provide valuable information on the thermal history of sedimentary basins (McNeil *et al.*, 1996).

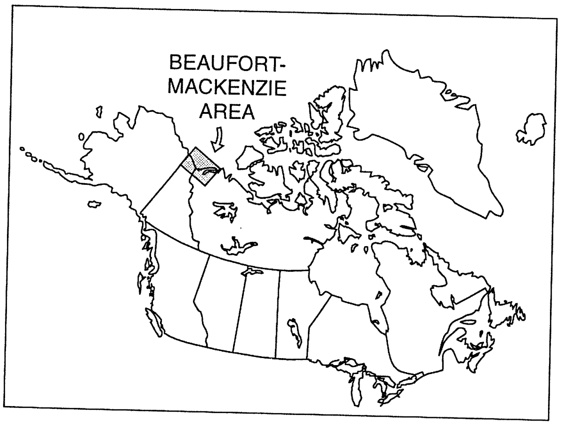



**Fig. 1.** Location of Beaufort–Mackenzie area

In this paper, diagenesis is reviewed from numerous perspectives based largely on information culled from the diagenetic literature and on observations and research into the distributions of benthic foraminifers in the Beaufort–Mackenzie Basin of Arctic Canada (Fig. 1). This basin is of interest for several reasons, and generalizations can be inferred to other similar basins. It has been the site of thick accumulations of deltaic influenced terrigenous clastic sediments (10 to 15 km preserved) through most of the Cenozoic and is currently receiving sediment and actively subsiding. It is a petroleum exploration basin that has received a significant amount of exploration, but still holds much potential for future exploration. Its biostratigraphy relies heavily upon benthic foraminiferal assemblages which have been recovered through a wide spectrum of terrigenous clastic sedimentary environments.

It was observed at an early stage in the exploration of this sedimentary basin that two aspects of the foraminiferal record were abundantly clear. The first centred around the conspicuous reciprocal relationship between the distributions of agglutinated versus calcareous benthic foraminifers in various parts of the basin and at various times through the history of the basin (Schröder-Adams & McNeil, 1994). The second stemmed from the observation that the mineralogical, textural, and overall visual appearance of the foraminiferal assemblages changed progressively (through burial diagenetic processes) as depth of burial and temperature increased (McNeil *et al.*, 1996).

Empirical data from the Beaufort–Mackenzie Basin indicated that diagenetic processes had affected foraminiferal assemblages in different parts of the basin in a variety of ways. These differing diagenetic affects can be described in the context of diagenetic regimes (Fig. 2) previously out-

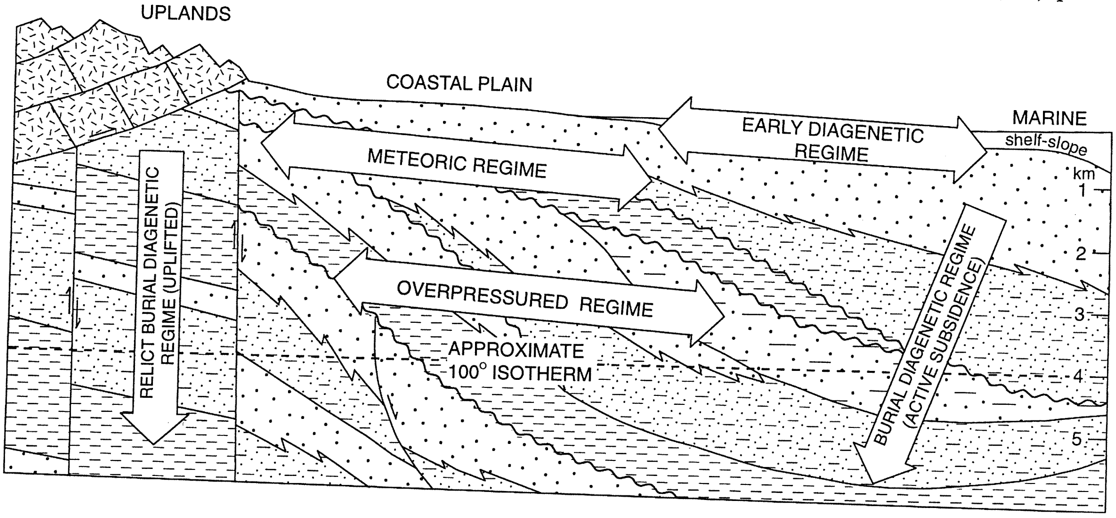



**Fig. 2.** Generalized distribution of diagenetic regimes in a continental margin, terrigenous clastic sedimentary basin such as the Beaufort–Mackenzie Basin. *Coarsely stippled areas* represent coarse clastic sediments; *darker patterns* represent finer grained mudstones and shales

lined in the diagenetic literature, for example, Bjørlykke (1983) and Galloway (1984). The diagenetic regimes exclude pre-diagenetic processes such as dissolution by undersaturated sea water, abrasion, predation, borings, etc. The first regime, or early diagenetic regime, controls the initial fossilization of benthic foraminifers in the uppermost several metres of sediment immediately below the sediment-water interface. The second regime is that of burial diagenesis which is controlled by increasing temperature at depth. In this regime, organic matter is progressively matured, diagenetic fluids are mobilized, and secondary mineralization and recrystallization occurs. The third regime is the overpressured regime in which fluids, trapped in rapidly deposited sediments, attain abnormally high pore pressure. The overpressured regime is important because of potential affects on the maturation of organic matter and affects on other chemical reactions in an abnormally pressured fluid environment. A fourth diagenetic regime is referred to as the meteoric regime defined by the flow of meteoric waters into the subsurface potentially to depths of 2 km controlled by hydraulic pressure and permeability pathways. This regime may have a limited affect on the preservation of calcareous foraminifers and also a minor influence on the mobilization and precipitation of silica and clay minerals.

Diagenetic regimes in sedimentary basins have been reviewed by numerous authors. Broad overviews of the topic have been presented by Galloway (1984), Chilingarian and Wolf (1988), Bjørlykke (1989), and Harrison and Tempel (1993). The chemical-biological environment of the early diagenetic zone has been reviewed by Hesse (1990). Numerous accounts on taphonomy and the fossil record are compiled in the publications of Allison and Briggs (1991) and Donovan (1991). Compilations on a wide variety of diagenetic topics can also be found in publications by Marshall (1987), and McIlreath and Morrow (1990). Specific studies on the geochemistry and petrography of the Beaufort–Mackenzie Basin can be found in Bloch and Issler (1996), Ko (1992), Dixon (1996), and Schmidt (1987).

## THE EARLY DIAGENETIC REGIME

In normal marine environments, the early diagenetic regime begins in subaqueous sediments and comprises the uppermost 10 metres or so of sediment consisting of a brownish-yellowish oxidized layer, usually less than 10 cm thick, underlain by a much thicker greyish-blackish reducing layer (Fig. 3). In anoxic bottom water environments, the oxidation zone is absent and early diagenesis occurs entirely in the reducing zone. The majority of benthic foraminifers live at the sediment/water interface and within the oxidizing zone, but some species can tolerate anaerobic living conditions and burrow a short distance into the reducing layer (Moodley & Hess, 1992; Sen Gupta & Machain-Castillo, 1993; Kitazato, 1994).

Early diagenesis is dominated by chemical and biological interactions dependent on the nature of the pore water, the amount and type of organic matter, the abundance of reactive iron, the sedimentation rate, the level of bioturbation, and most importantly the activity of either aerobic or anaerobic bacteria (Hesse, 1990). Foraminifers are involved directly in early diagenetic processes as contributors of calcareous and siliceous tests, as infaunal inhabitants and bioturbators of the oxidized zone, and as suppliers of organic matter and traces of reactive iron.

In normal marine sediments, pore waters at shallow burial depth in the oxidizing zone are typically oxygenated, calcium carbonate saturated, and pH neutral. Berner's (1981) measurements of hundreds of estuarine and normal marine sediments indicated that 90% were between pH 6.5 and 7.5. Under these conditions calcareous tests will generally be stable, particularly because of the affects of natural buffers (clays, dissolved gases, organic acids, etc.). There are, however, numerous processes capable of producing acids, carbonic acid in particular, that are potentially destructive for foraminifers. In the aerobic zone, bacterial decay of organic matter produces $CO_2$ through the simplified equation $CH_2O + O_2 = CO_2 + H_2O$. Hydration of $CO_2$ leads to the highly dissociated carbonic acid ($H_2CO_3$) in the reaction $CO_2 + H_2O = H^+ + HCO_3^-$ (Golubic & Schneider, 1979). If pH is not buffered (an unlikely situation), a build-up of carbonic acid would lead to the dissolution of calcareous foraminifers. Fluid exchange with overlying marine water also negates the potential for carbonic acid buildup (Kidwell & Bosence, 1991).

Aerobic bacterial activity acting directly on foraminifers can play a significant role in diagenetically altering the foraminiferal record. Freiwald (1995) documented that bacterial degradation of organic material within *Cibicides lobatulus* created minutely localized buildups of $CO_2$. Microscopic-scale dissolution occurred, even though surrounding pore waters were saturated with respect to calcium carbonate, because $CO_2$ was trapped under a biofilm secreted by the bacteria.

Bacterial decomposition of organic material in the oxidation zone is generally so active that it depletes available oxygen below the upper few centimetres of sediment as sediments become buried and pass into the anaerobic reducing zone. Decomposition of organic matter in this zone occurs mainly through anaerobic, bacterially mediated, sulphate reduction although nitrate, iron, and manganese reduction may be important locally. A simplified equation for this process is $2CH_2O + SO_4^{2-} = H_2S + 2HCO_3^-$ (Berner, 1985; Hesse, 1990). A consequence of this generalized reaction is an overall increase in $CO_2$ (carbonic acid and its dissociated species bicarbonate $HCO_3^-$ and carbonate $CO_3^{2-}$; Hesse, 1990, p. 283) and potential dissolution of calcareous fossils in some facies such as black shales (Curtis, 1980). In most shelf sediments, however, the sulphate reduction process leads also to the reduction of iron which raises pH (Curtis, 1980; Boudreau & Canfield, 1988; Abercrombie, *personal communication*). Iron is generally abundant in terrigenous clastic shelf sediments in the form of oxidized coatings on iron-bearing detrital grains.

An important aspect for both calcareous and agglutinated foraminifers in the reducing zone is the formation of diagenetic pyrite. Pyritization occurs in the sulphate reducing zone under anaerobic conditions. Organic matter is metabolized by anaerobic sulphate reducing bacteria. Hydrogen sulphide produced in the reduction of sulphate either

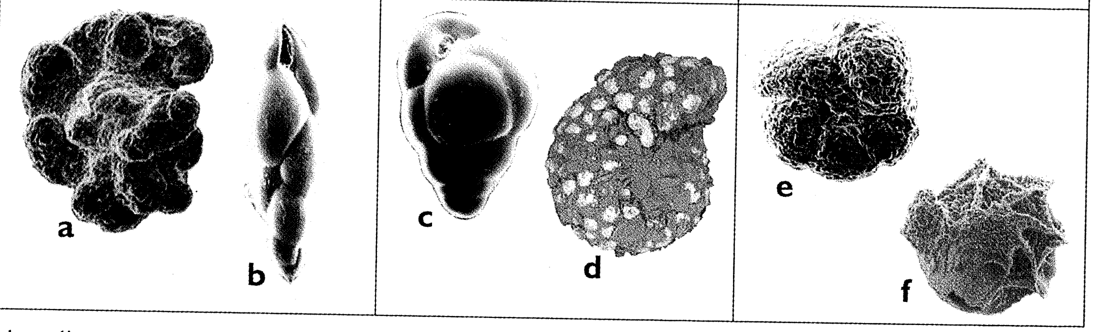
| | EARLY DIAGENETIC REGIME | | |
|---|---|---|---|
| | STAGNANT MARINE | NORMAL MARINE | DELTAIC |
| BOTTOM WATER | - water stratification<br>- limited circulation<br>- dysaerobic to aerobic<br>- limited benthos<br>- marine organic rich<br>- sulphate rich | - active circulation<br>- oxygenated<br>- diverse, abundant benthos<br>- marine/terrestrial organics<br>- sulphate rich | - rapid,turbid, sedimentation<br>- oxygenated<br>- abundant benthos, low diversity<br>- terrestrial organics<br>- iron rich, sulphate poor |
| OXIDIZING LAYER mm - cm | - oxidizing layer absent<br>- if bottom water is dysaerobic, oxidizing layer is thin, as oxygen is quickly depleted by aerobic bacteria | - aerobic bacteria feed on marine organic material, depleting oxygen<br>- pH variable within limited range (~7.0-7.5) | - sediments are well oxygenated<br>- aerobic bacteria may deplete organic matter |
| REDUCING LAYER cm - 10 m | - anaerobic bacteria reduce and deplete sulphate<br>- principal products are bicarbonates, hydrogen, and bisulphide ion<br>- pH may be acidic causing dissolution of calcareous foraminifers<br>- pH could be buffered by silicate hydrolysis | - anaerobic bacteria reduce sulphate and ferric iron<br>- pyrite is formed from reduced iron and dissolved $HS^-$<br>- ferric iron in agglutinated foraminifers is reduced, possibly leading to disaggregation of agglutinated test | - anaerobic activity may be limited by lack of marine organic matter<br>- agglutinated foraminifers are well preserved and calcareous species can be dissolved by acidic meteoric water |
| FOSSIL RECORD |  | | |



**Fig. 3.** Summary of fluids, sediments, chemistry, bacterial activity, and foraminiferal preservation in the oxidizing and reducing layers of the early diagenetic regime. **a.** *Trochammina*, GSC 109450, aberrant growth in low oxygen conditions; **b.** *Stainforthia*, GSC 89548, adapted for survival in low oxygen environment; **c.** *Turrilina*, GSC 89524, normal marine habitat; **d.** *Reticulophragmium*, GSC 109537, specimen replaced partly by pyrite (light areas); **e.** *Portatrochammina*, GSC 89532, excellent preservation in deltaic habitat; **f.** *Asterigerina*, GSC 109538, partial dissolution, iron oxide encrusted

migrates upward to be re-oxidized to sulphate or reacts with detrital iron minerals to form a series of sulphides with the end product being pyrite (Berner, 1985). In anoxic marine environments, the oxidation zone is absent and $H_2S$ is naturally abundant in bottom waters. Pyrite formation may be limited by the supply of reactive iron and/or sulphate or organic material (Berner, 1985).

In normal marine shelf settings, intense bioturbation may disrupt the above cited processes and lead to the dissolution of calcareous fossils (Lewy, 1975; Aller, 1982) in the following manner. Bioturbation inhibits an increase in pH, despite high rates of sulphate reduction and bicarbonate ion production, by the addition of carbon dioxide through respiration. Bioturbation sustains a supply of sulfate-rich bottom water, and in the presence of reactive detrital iron (abundant in terrigenous clastic sediments), pyrite forms locally by sulphate reduction. Further bioturbation exposes the pyrite to oxidation and produces sulfuric acid microenvironments capable of dissolving calcareous tests. In Baltic Sea sediments, Lewy (1975) noted that bioturbation in the upper 20 cm of sediment could recycle calcareous foraminifers to the surface where they were dissolved by sea water. Below the level of bioturbation, anaerobic decomposition of organic matter, sulphate reduction, and reduction of iron supports good preservation of calcareous fossils. Lewy (1975) and Aller (1982) have both documented the preservation of deep burrowing calcareous shelly organisms.

Another potentially destructive early diagenetic mechanism affecting agglutinated foraminifers was proposed by Schröder-Adams and McNeil (1994). They speculated that the intense activity of sulphate reducing bacteria in fine-grained marine muds with abundant organic matter could lead to the disaggregation of agglutinated foraminifers. Disaggregation occurs because the test-stabilizing reactive ferric iron in the organic cement of agglutinated foraminifers is reduced to ferrous iron and utilized in the formation of pyrite. Evidence to support this hypothesis came from sediments containing abundant pyrite associated with rare pyritic steinkerns of agglutinated foraminifers in the midst of well preserved assemblages of calcareous foraminifers (Fig. 3c,d). The pyritic steinkerns were interpreted as sparse representatives of a disaggregated assemblage of agglutinated foraminifers.

Sedimentation rate is an important factor in early diagenesis (Hesse, 1990, p. 283). In high depositional rate environments, such as the Beaufort–Mackenzie Basin and many continental shelf areas, more organic matter survives the oxidation zone and is preserved in a reactive state in the reducing zone. The increased sediment thickness also traps the products of early diagenesis because diffusion and advection is decreased. The combination of these factors leads to more intense diagenetic reactions such as sulphate reduction in rapidly buried sediments. Sulphate reduction is the initial step in the formation of pyrite and this process impacts on foraminifers in both constructive and destructive ways.

Formation of early diagenetic pyrite can be constructive in the preservation of foraminifers, particularly for agglutinated species, by reinforcing the test interior against collapse from burial pressure. Pyritization in foraminifers could also be destructive if pyrite is broken down later by oxidation from weathering processes. This process of oxidation can be simulated in laboratory processing if calcareous foraminifers with pyritic moulds are processed with hydrogen peroxide. Hodgkinson (1991) has cautioned micropaleontologists against using hydrogen peroxide to process pyritized microfossils, but experience has shown that it can be used if diluted (10%) and processing time is minimized.

## THE BURIAL DIAGENETIC REGIME

The burial diagenetic regime occurs at depth in sedimentary basins as temperatures and pressures increase and there are no direct links to the meteoric or the marine hydrosphere (Fig. 2). Hesse (1990) noted that burial diagenetic reactions begin at about 75°C as bacterially mediated decomposition reactions in organic matter give way to thermally controlled (kinetic) reactions, such as the thermal maturation of kerogen. Depending on the geothermal gradient and burial rate, kerogen maturation begins taking place as rising temperatures generate the activation energy necessary for breaking organic molecules. In a normally compacting continental shelf basin, these thermochemical diagenetic processes begin operating at depths of 2–3 km. The burial diagenetic regime applies through a temperature range of about 75°C to about 200–250°C and burial depths up to about 8 km where mineral transformations of the early metamorphic regime begin.

Fossil foraminifers exhibit many of the classical thermal effects that occur progressively in organic matter and various mineral phases through the burial diagenetic regime. For example, the Foraminiferal Colouration Index (FCI) recently developed from Beaufort–Mackenzie Basin data by

**Table 1**

Foraminiferal Colouration Index (FCI) of McNeil *et al.* (1996) relative to standard colours of the Munsell Soil Colour Chart

| FORAMINIFERAL COLOURATION INDEX (FCI) | | |
|---|---|---|
| FCI | MUNSELL COLOUR STANDARD | |
| 0 | 7.5YR6/6 | reddish yellow |
| 1 | 10YR8/1, 10YR8/2 | white |
| 2 | 10YR7/1, 10YR7/2 | light grey |
| 3 | 10YR6/1, 10YR6/2 | light brownish grey to grey |
| 4 | 10YR5/1, 10YR5/2 | grey to greyish brown |
| 5 | 10YR4/1, 10YR4/2 | dark grey to dark greyish brown |
| 6 | 10YR3/1, 10YR3/2 | very dark grey to very dark greyish brown |
| 7 | 10YR2/1, 10YR2/2 | very dark brown to brownish black |
| 8 | 10YR2/1, N2/0 | very brownish black |
| 9 | N2/0 | black (partially translucent) |
| 10 | N2/0 | black (opaque) |

McNeil *et al.* (1996) documents thermally controlled colour alteration of the organic cement in agglutinated foraminifers (Table 1). FCI data and calculated borehole temperatures in the Beaufort–Mackenzie Basin indicate that thermal maturation of foraminifers begins at temperatures of 60–70°C and that a fairly rapid colour change to brownish black occurs as temperatures rise to 100–140°C, making foraminifers sensitive thermal indicators in the early stages of petroleum generation.

Foraminifers are also sensitive indicators of thermally controlled diagenetic mineralization that occurs in the burial diagenetic regime involving silica, carbonates, and clay minerals (McNeil *et al.*, 1996). From a micropaleontological perspective, the most widespread and readily recognizable of these mineralogical changes is the silicification of agglutinated foraminifers (Fig. 4). Silicification occurs by precipitation of secondary quartz, as overgrowths on quartz grains in the foraminiferal test (McNeil *et al.*, 1996). Mineral assemblages and temperatures control the amount of pore-fluid silica available for precipitation. Initially, during early diagenesis, pore fluids are oversaturated with respect to quartz, but precipitation does not generally occur until kinetic limitations on quartz precipitation are overcome in the burial diagenetic realm (Abercrombie *et al.*, 1994). In the burial diagenetic realm, the silicification of agglutinated foraminifers is progressive with increasing temperature and burial (McNeil *et al.*, 1996). Mapping of the horizon at which silicification initially occurs provides a readily recognizable datum for the upper limit of burial diagenetic processes affecting foraminifers. This is a potentially important datum in hydrocarbon basins since it marks the point at which porosity and cementation in reservoirs are also likely to be affected. Bjørlykke and Egeberg (1993) have noted that in normally subsiding basins, most quartz cement forms at temperatures above 90–100°C.

Thermal maturation of agglutinated foraminifers occurs as the organic cement (glycosaminoglycan) that coats all grains within and around the agglutinated test wall is heated and volatile components are driven off. Glycosaminoglycans, previously referred to as mucopolysaccharides, are unbranched polysaccharide chains of proteoglycans composed

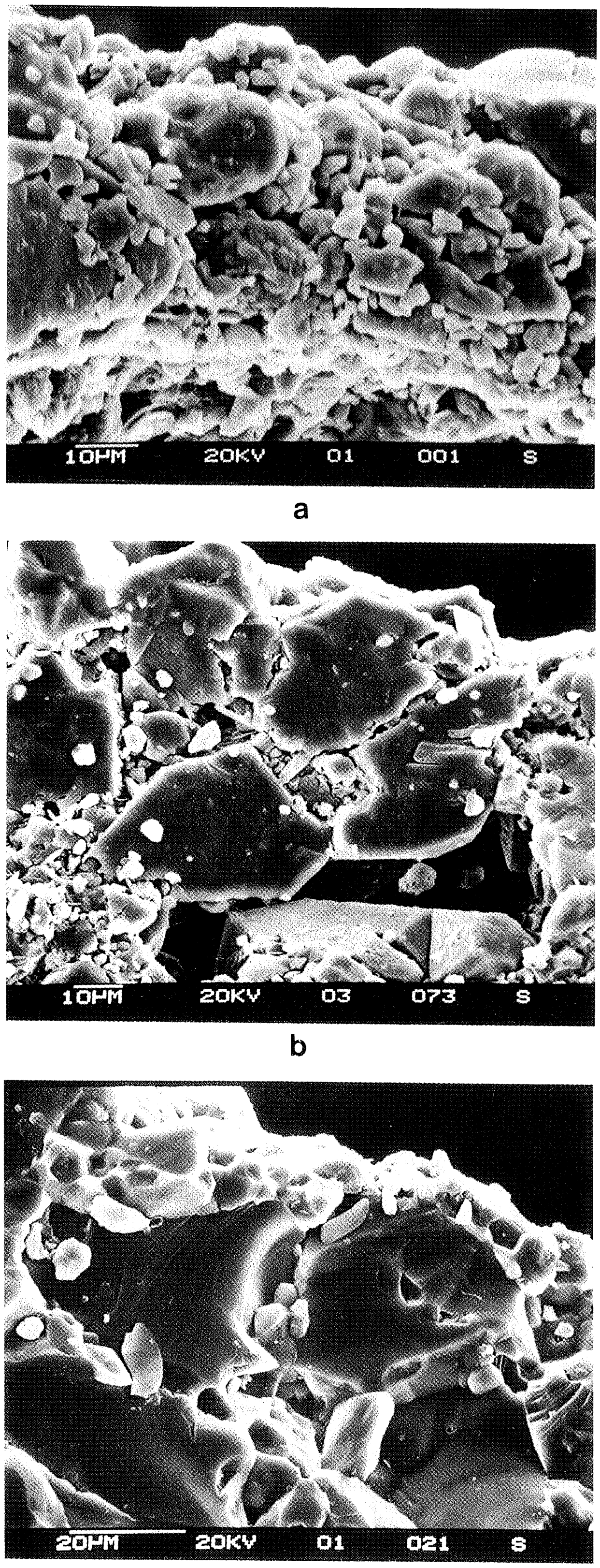



**Fig. 4.** Progressive silicification in the wall of agglutinated foraminifers. **a.** Recent, unaltered specimen of *Recurvoides*, GSC 112234; **b.** well developed quartz overgrowths in a silicified specimen of *Reticulophragmium*, GSC 112235, burial depth 4115 m; **c.** solid overgrowth cementation in silicified specimen of *Haplophragmoides*?, GSC 112236, estimated burial depth 6–8 km

of repeating disaccharide sequences of amino sugars (Langer, 1992). Little is known at present about the detailed chemical reactions involved in the maturation of organic cement, but heating experiments using a pyrolysis apparatus have simulated geologically observed colour changes in both Recent and fossil foraminifers (McNeil *et al.*, 1996).

The preservation of organic cement in fossil foraminifers is a fundamental prerequisite for their use as thermal maturation indices. Colour changes through pyrolysis experiments provide visual evidence of thermal maturation. SEM examination of the microstructure of agglutinated foraminifers provides direct evidence of the occurrence and distribution of fossil organic cement. In order to separate organic cement from original detrital grains, usually quartz, in the test wall, fossil specimens were etched briefly in hydrofluoric acid. Figure 5 illustrates fossil organic cement separated from an agglutinated grain of quartz.

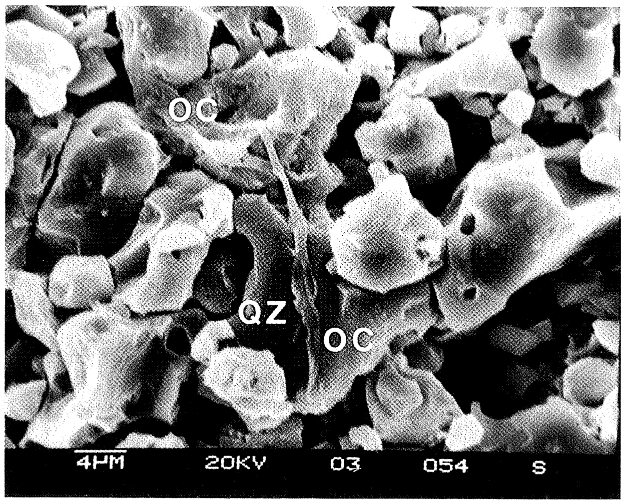



**Fig. 5.** Quartz grains (QZ) with remnants of organic cement (OC) in the wall of *Bathysiphon*, GSC 109545, after etching in dilute hydrofluoric acid

The application of FCI in assessing thermal maturity in sedimentary basins has been dealt with thoroughly by McNeil *et al.* (1996). Data from one well in the Beaufort Sea (Amauligak J-44) are presented here to illustrate the application of FCI (Fig. 6; Table 2). Quantification and accurate determination of fossil colour is initially achieved by visual comparison of individual specimens against the standard Munsell Colour Chart. The statistical mean and standard deviation of measurements for each sample then provides a reliable assessment of the thermal maturity for the entire sample. Anomalies in the distribution of FCI values within any one sample are usually explained either by reworked microfossils or by contamination through caved well cuttings, as the data from sample 3797 m in Amauligak J-44 illustrates (Fig. 6; Table 2). FCI therefore provides a quantitative measure of well sample quality.

In addition to conspicuous colour changes, burial diagenesis is responsible for textural and mineralogical changes in foraminifers (McNeil *et al.*, 1996). As already noted, silicification of agglutinated foraminifers (Fig. 4) is

**Table 2**

FCI data from well cuttings samples of the Amauligak J-44 well in the Beaufort Sea (from McNeil *et al.*, 1996)

| Depth (m below KB) | Average TVD (m below SL) | FCI data 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | Average FCI |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 2200-2210 | 2150 | | 1 | | | | | | | | | 2 |
| 2395-2405 | 2345 | 8 | 130 | 34 | 1 | | | | | | | 2.16 ± 0.49 |
| 2597-2603 | 2547 | 23 | 42 | 62 | 2 | | | | | | | 2.33 ± 0.78 |
| 2780-2804 | 2730 | 34 | 56 | 90 | 6 | | | | | | | 2.37 ± 0.82 |
| 2999-3005 | 2949 | 8 | 3 | 22 | 23 | | | | | | | 3.07 ± 1.02 |
| 3197-3203 | 3147 | 5 | 2 | 17 | 7 | 1 | 2 | | | | | 3.09 ± 1.24 |
| 3398-3404 | 3348 | 3 | 25 | 133 | 101 | 42 | 11 | | | | | 3.59 ± 0.97 |
| 3599-3605 | 3549 | 9 | 6 | 4 | 16 | 31 | 14 | | | | | 4.20 ± 1.56 |
| 3797-3803 | 3747 | 14 | 19 | 22 | 23 | 14 | 8 | | | | | 3.28 ± 1.34 |
| 3998-4002 | 3948 | | 5 | 9 | 5 | 18 | 52 | 44 | 15 | | | 5.99 ± 1.44 |

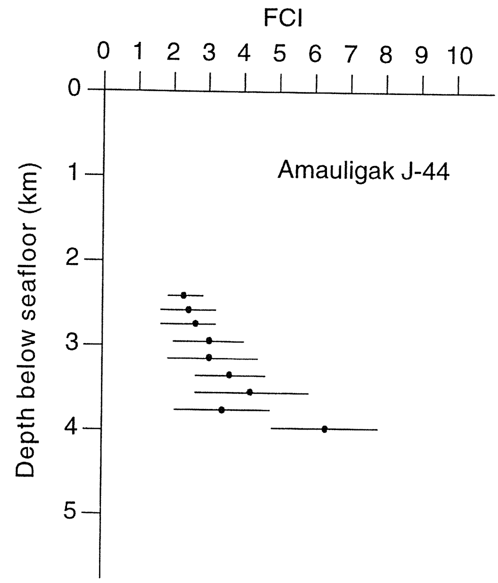



**Fig. 6.** Plot of FCI data versus depth in the Amauligak J-44 well, see Table 2 for raw data. (from McNeil *et al.*, 1996) Note: sample at 3797 m with anomalously low FCI average value of 3.28 represents caved well cuttings

one of the most obvious of these diagenetic changes, but other diagnostic minerals commonly found in fossil foraminifers include kaolinite, smectite, illite, chlorite, and recrystallized calcite. A zonation of mineralogical and textural changes observed in fossil foraminifers has been established by McNeil *et al.* (1996). Figure 7 summarizes briefly the main features of this zonation. Mineralogical and textural features of the diagenetic sequence are easily observed through SEM techniques.

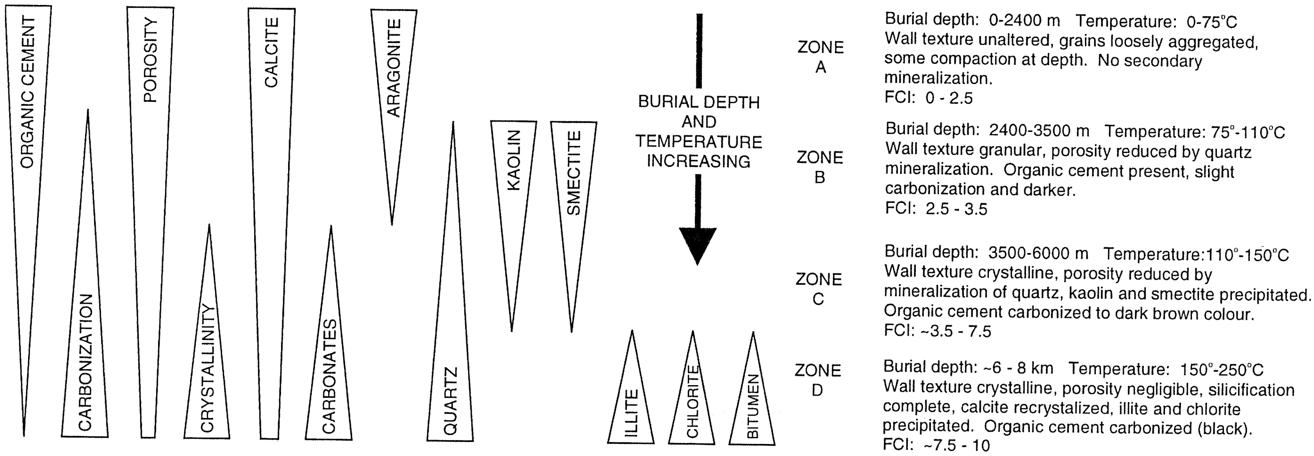



**Fig. 7.** Zonation of burial diagenetic trends in foraminifers related to approximate temperatures, burial depths, and Foraminiferal Colouration Index (FCI) (modified from McNeil *et al.*, 1996)

Clay minerals are a distinctive secondary mineralization feature in both agglutinated and calcareous foraminifers (Figs. 8, 9). Their distribution follows predictable trends in clay mineralogy through increasing temperature regimes. At low levels of thermal maturation, kaolinite and smectite are the stable clay minerals and their more or less simultaneous precipitation can be observed within the wall of silicified, agglutinated foraminifers (Fig. 8). At higher levels of thermal maturation, illite and chlorite are the stable clay minerals (Fig. 9). The occurrence of smectite and illite in hydrocarbon basins has been the focus of much research based on the increase in the illite/smectite ratio at burial depths greater than approximately 2 km and the relationship of the smectite/illite reaction to the generation and migration of petroleum (Burst, 1969). Generally, this transition is related to temperature, burial rate, geothermal gradient and pore fluid composition. Abercrombie *et al.* (1994) have shown that the reaction may be linked to aqueous silica activity and occur between temperatures of 50°C to 150°C depending on relative burial rates, i.e. time-temperature relationships. McNeil *et al.* (1996) recognized illite as a diagnostic mineral in foraminiferal burial diagenetic zones C and D (Fig. 7).

The development of significant amounts of chlorite (Fig. 9) was observed in recrystallized calcareous benthic foraminifers. Chloritization of calcareous foraminifers is not a localized phenomenon as it has been observed in both the Beaufort–Mackenzie Basin and the Western Canadian Sedimentary Basin. Chloritization is pervasive and can alter the entire calcareous foraminiferal assemblage. In the foramini-

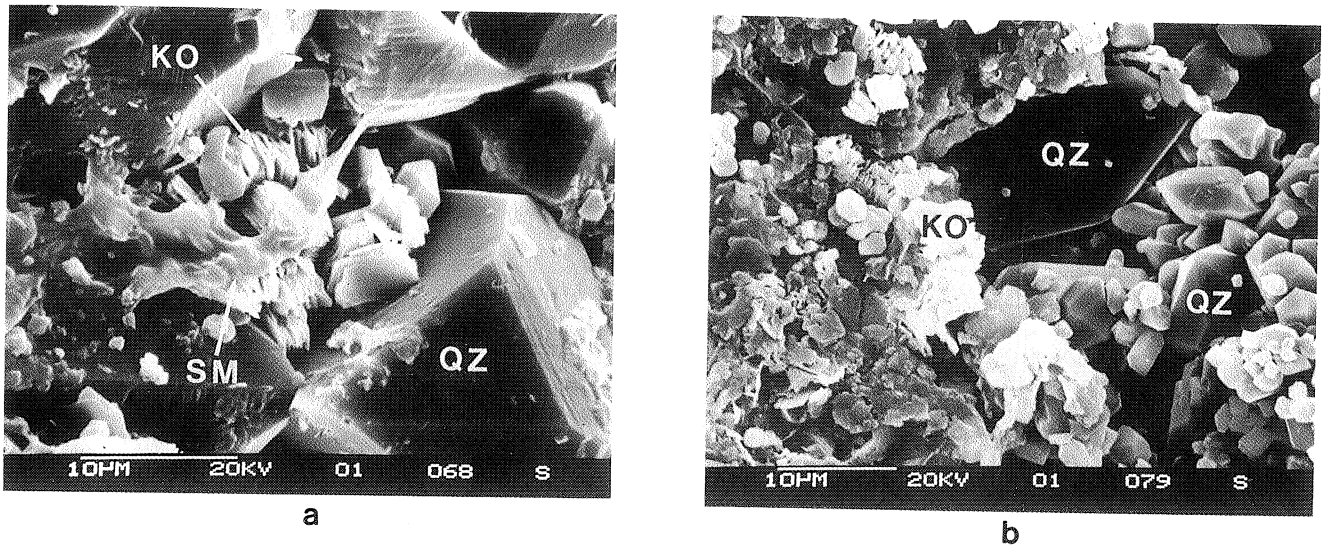



**Fig. 8.** Diagenetic quartz and clay minerals in the tests of agglutinated foraminifers. **a.** Diagenetic quartz crystals (QZ), kaolinite (KO), and smectite (SM) in the interior of the agglutinated wall of *Labrospira*, GSC 112290, from Amauligak J-44, 3599 m; **b.** Diagenetic quartz (QZ) and kaolinite (KO) filling chamber lumen in *Reticulophragmium*, GSC 109544, from Amauligak J-44, 3197 m

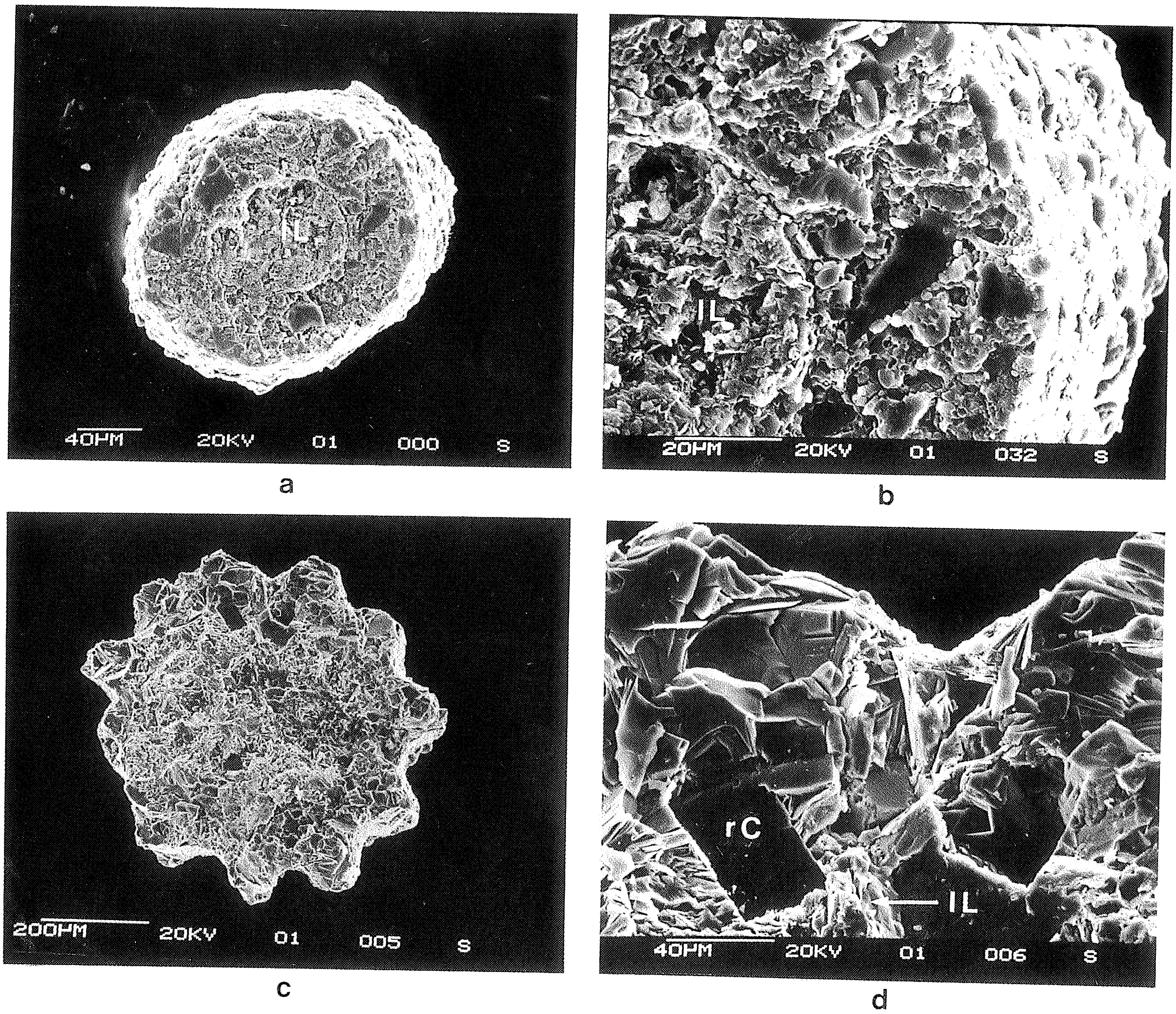



**Fig. 9.** Clay mineralization in agglutinated and calcareous benthic foraminifers. **a, b.** Illite (IL) filling the interior of *Bathysiphon*, GSC 112293, estimated burial depth approximately 5 km; **c, d.** Chlorite (CL) and recrystalized calcite (rC) in the wall of *Marginulina*, GSC 112294, estimated burial depth of 6–8 km

feral burial diagenetic zonation of McNeil *et al.* (1996), chlorite is diagnostic of zone D signifying burial temperatures in the range of 150–250°C and burial depths of 6 to 8 km. Chlorite is additionally significant as a geothermometer, since it crystallizes in a series of temperature controlled polymorphs (Hayes, 1970).

Another temperature controlled burial diagenetic effect in foraminifers was described by Reiser (1988) who recognized that the aragonitic calcite of the family *Robertinacea* was stable at low temperature and burial depth, but converted to calcite by means of thermal alteration. In Lower Cretaceous shales of northwest Germany, this transformation occurred at depths of about 3 km and paleotemperatures of 100°C and thus could serve as a useful tool for paleotemperature reconstruction.

## THE OVERPRESSURED REGIME

The overpressured regime occurs in the subsurface of sedimentary basins where fluid pressures can significantly exceed normal hydrostatic pressure (Fig. 2). Overpressured fluids have long been of interest to the petroleum industry for a variety of reasons including their relationship to the generation and migration of hydrocarbons, effects on the porosity and permeability of reservoirs, and safety factors during drilling.

The overpressured condition develops as fluids become trapped within or beneath low permeability sediments undergoing compaction and burial. Several factors are usually cited in the development of overpressured fluids. These include rapid deposition of fine grained sediment, tectonic stress, expansion of heated fluids, mineral transformations, and hydrocarbon generation (Sharp *et al.*, 1988; Snowdon, 1995). Of these processes, the most significant is thought to be rapid deposition of fine-grained sediments leading to compaction disequilibrium and the entrapment of increasingly pressurized pore fluids.

In the Beaufort–Mackenzie Basin, the main overpressured zone occurs at depths below 2 km, but occurs progressively deeper (3 or 4 km) in offshore areas where the Pliocene–Pleistocene Iperk Sequence attains thicknesses of 2 to 4 km. Hitchon *et al.* (1990) considered that the main overpressured zone in the Beaufort–Mackenzie Basin developed during the mid-Cenozoic and was depressed later by relatively rapid deposition of the Iperk Sequence in the late Cenozoic. Overpressured fluids, trapped since the mid-Cenozoic, are now buried much deeper in an offshore direction as a result of subsidence under the thick Pliocene–Pleistocene Iperk Sequence. The geothermal gradient, on the other hand, has adapted to this downwarping as the 100°C isotherm (Fig. 2) is fairly uniform at a depth of approximately 3.8 km (Hitchon *et al.*, 1990). The overpressured zone thus occurs through a range of geothermal environments.

Hitchon *et al.* (1990) reported that the salinities of formation waters average 10‰ in geopressured zones and 21‰ in normal hydrostatically pressured zones. They concluded that the low salinities of geopressured waters is a shale membrane effect filtering fluids that moved through muddy sediments, but this interpretation has been rejected generally as a viable mechanism for producing low salinity subsurface waters in compactional regimes (Hanor, 1994). It is more likely that low salinity waters of meteoric origin were trapped in porous units and then became overpressured. In contrast, a similar occurrence of low salinity water concentrated in the upper part of the geopressure zone in the Gulf of Mexico, has been attributed to water released from smectite/illite conversion (Morton & Land, 1987).

A complete understanding of diagenetic phenomenon within overpressured zones is difficult at best, and is certainly beyond the scope of this paleontologically oriented overview. In the context of agglutinated foraminifers, however, two aspects of diagenesis in the overpressured regime are of particular interest – silicification and thermal colour alteration.

Silicification of agglutinated foraminifers is a widespread diagenetic phenomenon. In the paragenetic zonation of McNeil *et al.* (1996), silicification occurs progressively through zones B to D representing burial depths of 2.4 to about 8 km and temperatures of 75 to 200°C or more. A preliminary examination of a number of wells in the Beaufort–Mackenzie Basin indicates that silicification begins to occur above or nearly coincident with the main overpressure zone. Figure 10 illustrates a typical example. This empirical evidence is important because it implies that the silicification of foraminifers may be used a predictive tool for recognizing overpressured zones in exploration drilling. In offshore areas, where the rapidly deposited Iperk Sequence has depressed the main overpressured zone, the onset of silicification also occurs deeper. This is a kinetic effect (time/temperature) documented by FCI trends (McNeil *et al.*, 1996) and is consistent with Abercrombie *et al.* (1994) who showed that onset of quartz precipitation may be delayed in

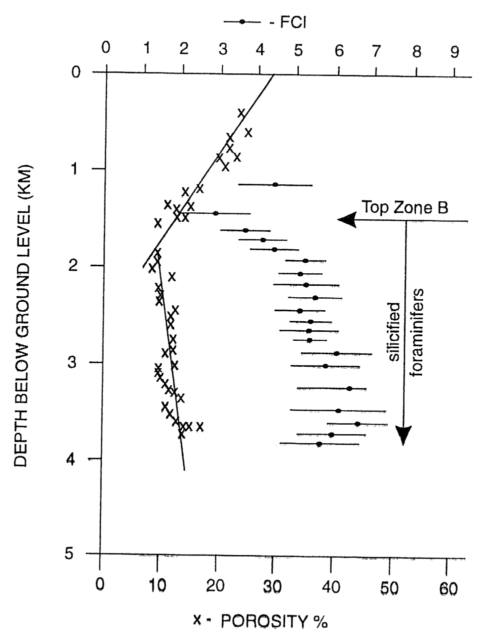



**Fig. 10.** Distribution of FCI and shale porosity trends in the Reindeer D-27 well of the Mackenzie Delta (from McNeil *et al.*, 1996). Silicification of agglutinated foraminifers (top of burial diagenetic zone B) begins at 1624 m; overpressured zone begins at approximately 2000 m as indicated by reversal in porosity trend

basins undergoing rapid burial and high heating rates.

The interior of the agglutinated test appears to be a remarkably receptive environment for the formation of quartz overgrowths (Fig. 4). Quartz overgrowths are confined to the interior of the test wall and do not extrude from the outer wall of the foraminiferal test, apparently inhibited by the thick outer organic layer. McNeil *et al.* (1996) concluded that silicification in foraminifers resulted from quartz overgrowths precipitated from quartz-saturated pore fluids controlled by temperature and mineral assemblages. Bloch and Hutcheon (1992) have described processes involving quartz mineralization in shale microenvironments. The most likely mechanism for the redistribution of quartz is the dissolution/pressure solution of detrital grains, particularly amorphous biogenic silica. The relative importance of potential sources for quartz cementation in sedimentary rocks is controversial (McBride, 1989); notably this includes silica released through the smectite/illite reaction.

In the North Sea Basin, fluid inclusion data from quartz cements indicate that most of the quartz cementation takes place at temperatures above 90–100°C (Bjørlykke & Egeberg, 1993) and at depths below 2.5 to 3.0 km. This compares broadly with depths greater than 2.4 km and temperature estimates of 75°C or more for quartz overgrowths in agglutinated foraminifers in the Beaufort–Mackenzie Basin (McNeil *et al.*, 1996; zones B–D). At and above these temperatures, quartz precipitation results when fluids migrate upward and cool (McBride, 1989, and Bjørlykke & Egeberg, 1993). Migration of fluids in shales and mudstones however is generally thought to be limited and insignificant as a mechanism for transfer of silica for distances greater than a few metres (Bjørlykke & Egeberg, 1993; Bloch & Hutcheon, 1992), so that silicification in foraminifers probably represents a remobilization of silica from local sources such as pressure solution or reactions between silicate minerals.

An intriguing potential source of silica for the silicification of foraminifers, and one that has a possible direct link to the development of overpressure, is the much researched and still controversial smectite/illite reaction. This reaction is a well documented phenomenon in sedimentary basins and occurs with increasing burial temperatures and depth (Powers, 1967; Hower *et al.*, 1976; Foscolos *et al.*, 1976; Elliot *et al.*, 1991). The reaction of smectite to illite consumes potassium and aluminium and produces silica and water, amongst other ions. Foscolos (1990) noted that quartz increases in abundance with burial depth and considered that the silica generated by clay reactions was precipitated as quartz overgrowths. This is a potential mechanism for the silicification of foraminifers since it generates both the silica and the water necessary to transport that silica into the foraminiferal test. Furthermore, the smectite/illite reaction may itself be a possible factor in causing overpressured conditions in shales. Freed and Peacor (1989) proposed that the smectite/illite reaction produced coalesced illite packets that decreased local permeability leading to a more efficient geopressure seal and a corresponding increase in pore fluid pressure. The fact that clay mineral reactions produce a significant amount of extra water contributes as well to the geopressured condition. The appealing aspect of the smectite/illite reaction is that it appears to be coincident with silicification trends and the development of the overpressured regime. Abercrombie (*personal communication*), however, believes that smectite is present only in small quantities in the Beaufort–Mackenzie Basin and that therefore its role in the generation and maintenance of overpressuring is questionable.

The other aspect of the overpressured condition that may be of relevance to foraminiferal diagenesis is its potential effect on the colouration of thermally altered foraminifers. In the subsurface of the Mackenzie Delta, McNeil *et al.* (1996) observed a correlation between overpressured zones and a retardation in thermal alteration colour as measured by FCI in agglutinated foraminifers in the Reindeer D-27 well (Fig. 10). Furthermore, near the base of the well, FCI actually decreased coincident with a marked increase in porosity caused by a peak in overpressure. These observations were preliminary, but will be followed up by further analysis and investigations in other wells. It is of interest, however, that recent investigations in the Tertiary basins of the South China Sea by Fang *et al.* (1995) have reached similar conclusions regarding the retardation of organic-matter maturation in overpressured environments. Fang *et al.* (1995) documented that the thermal maturity of normally pressured sediments was significantly higher than the thermal maturity of overpressured sediments as measured by vitrinite reflectance (%$R_O$). This organic maturity anomaly could not be explained by variations in activation energies, conductivity contrasts, or hydrological effects. Since the difference between the predicted and measured vitrinite reflectance level increased exponentially with increasing pore fluid pressure, Fang *et al.* (1995) concluded that increased pore fluid pressure increased the activation energies of organic matter maturation reactions.

Similar conclusions were reached from experimental work by Price and Wenger (1992) using aqueous pyrolysis techniques. These experiments attempted to approximate natural systems by generating thermal reactions under pressure in closed, water-wet systems. The reaction products (hydrocarbons) were measured by gas chromatography, and Price and Wenger (1992) concluded that increasing static fluid pressure strongly retarded the decomposition of kerogen, thus causing a decrease in the amount and nature of hydrocarbon products generated. Furthermore, they concluded that under natural geological conditions the effects of overpressure would be even greater because of the significant partial pressure contributions from hydrocarbon gases as opposed to helium which was used in the pyrolysis experiments.

In an attempt to analyse reservoir diagenesis and hydrocarbon migration, Swarbrick (1994) noted that the influence of high pressure on reaction kinetics is not well known, but that experimental studies (Enguehard *et al.*, 1990) showed that increasing pressure decreases the rate of thermal cracking of larger to smaller hydrocarbon molecules. High pressure also retards the carbonization of organic matter in a sealed aqueous system, as illustrated by experimental study on conodonts (Epstein *et al.*, 1977).

The overpressured regime is obviously a significant component in any basin analysis study. Price and Wenger

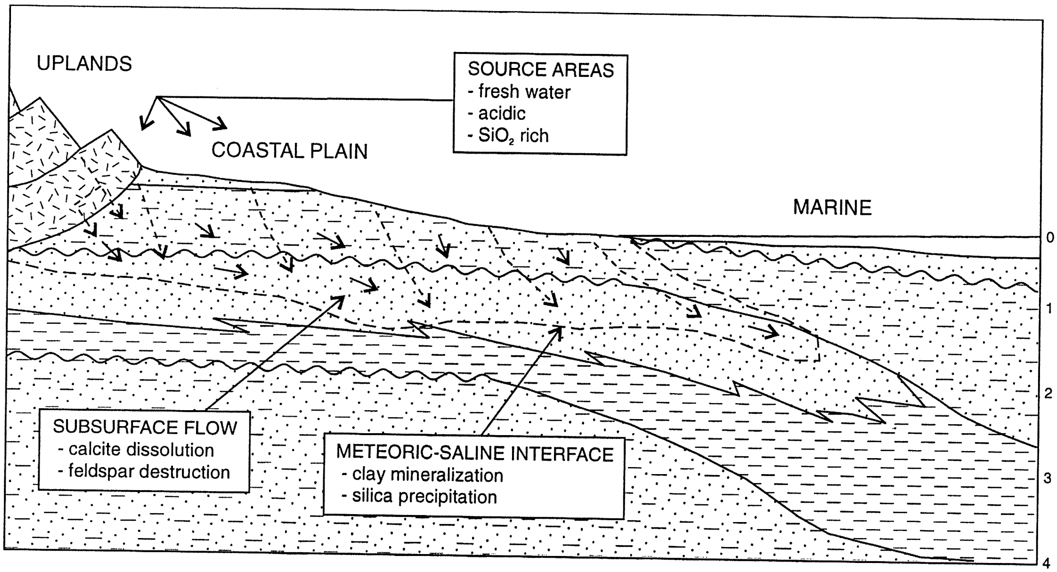



**Fig. 11.** Distribution and general characteristics of the meteoric regime in proximal sediments of a terrigenous clastic sedimentary basin

(1992) noted however that its effects are difficult to discern and document in nature due to the transient character of fluid pressures over time. If FCI measurements on agglutinated foraminifers provide a means of detecting overpressured conditions in the geological past, then they can make a significant contribution to understanding the evolution of sedimentary basins.

## THE METEORIC REGIME

The meteoric diagenetic regime plays an important role in numerous aspects of basin analysis including hydrocarbon migration and biodegradation, reservoir porosity, and secondary mineralization. Microfossils are generally not considered in investigations of meteoric diagenesis, but they are nonetheless potentially sensitive to the effects of dissolution and mineralization and therefore could provide important evidence of meteoric water diagenesis. Since interactions involving meteoric waters are capable of destroying microfossils and altering the fossil record, the meteoric regime is of fundamental importance to micropaleontology.

Meteoric water is typically oxidizing (positive Eh), acidic (about pH 5.5 or lower), silica over-saturated (10–20 ppm), and low in salinity (< 1‰). The flow of meteoric water from terrestrial environments into subsurface aquifers is driven by hydraulic head (Fig. 11) and generally is initiated by uplift in the proximal areas of sedimentary basins (Figs. 2, 11). At the basin margin, meteoric waters are involved in weathering processes, but as they flow into the subsurface of sedimentary basins through permeable pathways they become involved in diagenetic reactions through contact with more saline formational waters, unstable mineral species, and increasing temperatures. The meteoric regime can be very extensive and has been documented at depths of more than 2 km (Galloway, 1982; Hitchon *et al.*, 1990; Sharp *et al.*, 1988).

Diagenesis in the meteoric regime generally involves reactions with organic matter, carbonates, and silicates (Bjørlykke, 1989: Sharp *et al.*, 198). Silicates such as feldspar and mica are typically unstable in meteoric waters. As meteoric water evolves by interacting with subsurface pore water and sediments, authigenic silica and clay minerals, kaolinite in particular, are eventually precipitated (Fig. 11). To a limited extent, re-oxidation of earlier-formed iron minerals results in hematite and limonite deposits, particularly in sandstones. Meteoric waters that are initially oxidizing may be depleted of oxygen through diagenetic reactions with minerals and organic matter, thus eventually become reducing and gradually neutralized as well (Bjørlykke, 1989). Destruction of feldspars may increase the silica content of pore waters and lead to precipitation of quartz as silica rich fluids migrate into lower-temperature regimes (Bjørlykke, 1989).

In the Beaufort–Mackenzie Basin, formational waters have been flushed extensively by meteoric water to a depth of at least 2 km (Hitchon *et al.*, 1990). Snowdon (1988) also noted evidence of extensive meteoric effects indirectly through the occurrence of biodegraded oils in the onshore Richards Island area and to a lesser extent offshore. Biodegradation apparently occurred at several times during the Tertiary associated with periods of significant erosion. Very little is known however about the precise pathways, effects, and history of these meteoric waters.

Possible evidence of meteoric diagenetic activity comes from the microfossil record in the Issungnak O-61 and Isserk E-27 wells in the offshore Beaufort–Mackenzie Basin. These wells display good preservation of foraminiferal calcite except for conspicuously altered zones as indicated on Figure 12. In the Isserk E-27 well for example, within a sequence of well preserved calcareous benthic foraminifers, specimens from 1500 to about 1800 metres in the Mackenzie Bay Sequence are partly dissolved and coated with limonitic crusts (Fig. 12). Although there are no conspicuous meteoric water pathways into this mudstone section of the Mackenzie Bay Sequence, it is speculated that the calcite

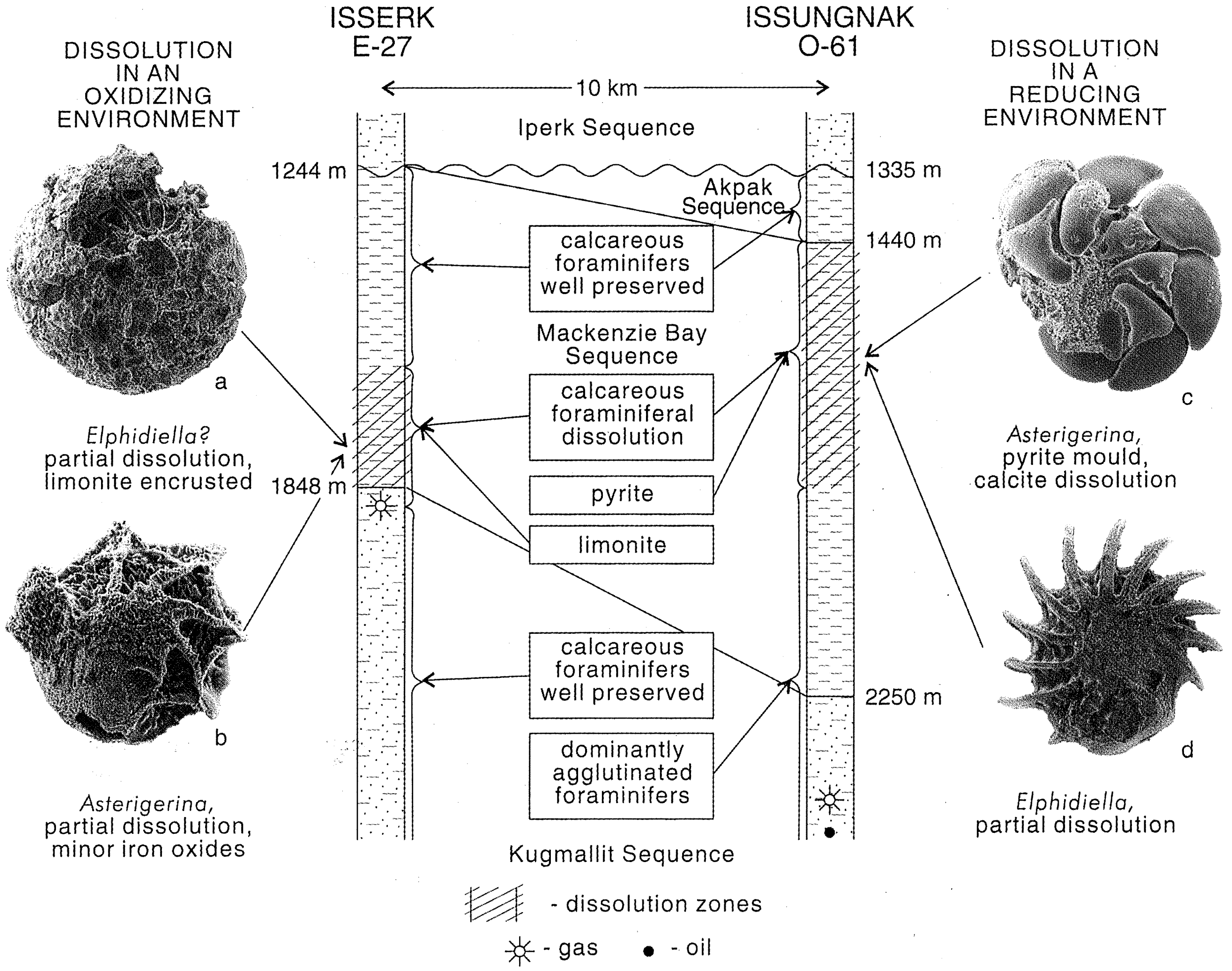



**Fig. 12.** Possible meteoric-water effects on the preservation of benthic foraminifers in contrasting oxidizing and reducing environments. **a.** *Elphidiella*?, GSC 109539; **b.** *Asterigerina*, GSC 109538; **c.** *Asterigerina*, GSC 109540; **d.** *Elphidiella*, GSC 109541

dissolution and limonitic coatings may have resulted from interactions with meteoric water. In the Issungnak O-61 well, an alternative circumstance for calcite dissolution appears to have occurred. In this well, pristine pyrite internal moulds from dissolved calcareous benthic foraminifers and partially dissolved calcareous benthic foraminifers are abundant (Fig. 12). The preservation of early diagenetic pyrite in these specimens appears to indicate dissolution in a reducing environment. This may have occurred as the result of dissolution from oxygen depleted meteoric waters.

A striking but enigmatic occurrence of secondary quartz overgrowths (Fig. 13a, b) was observed within the interior of tubular *Rhabdammina* in the Koakoak O-22 and Akpak 2P-35 wells by Schröder-Adams and McNeil (1994). Secondary quartz mineralization affects the entire foraminiferal assemblage and typically produces a mottled (light-dark) appearance on the test exterior. The exterior texture of the mottled surface is shown in detail on a specimen of *Haplophragmoides* in Figure 13(c, d). As in the Isserk E-27 and Issungnak O-61 sections cited above, these diagenetic effects occur anomalously within otherwise normal burial diagenetic trends at low maturity levels. Schröder-Adams and McNeil (1994) suggested precipitation of these quartz overgrowths as a function of burial diagenesis or the mixing of meteoric waters with formational pore waters. This anomalous quartz could also be explained by redistribution of amorphous silica from siliceous microfossils through thermally controlled burial diagenesis (Abercrombie, *personal communication*). Detailed isotopic analysis of silica isotopes might help to resolve the diagenetic history of this secondary quartz. Longstaffe (1993), for example, has recognized the diagenetic influence of meteoric water by the occurrence of low $^{18}O$ concentrations. Although such analyses have not been undertaken on the Beaufort–Mackenzie Basin material, this is a logical avenue for more detailed research.

A final note on diagenetic mineralization in the meteoric regime concerns the occurrence of kaolinite. Osborne *et al.* (1994) cited petrographic and isotopic evidence to indicate that kaolinite morphologies were diagnostic of specific temperatures and depths. Vermiform kaolinite was indicative of slow precipitation at shallow depths (571–1286 m) and low temperatures (25–47°C) in the meteoric regime and blocky kaolinite was characteristic of more rapid precipitation at greater depths ((1286–2143 m) and temperature (50–80°C) in the burial diagenetic regime. Kaolinite in planktonic foraminifers from the Turonian of western Canada has been recorded by Bloch *et al.* (1993, fig. 14c, d). They illustrated vermiform kaolinite filling chambers which

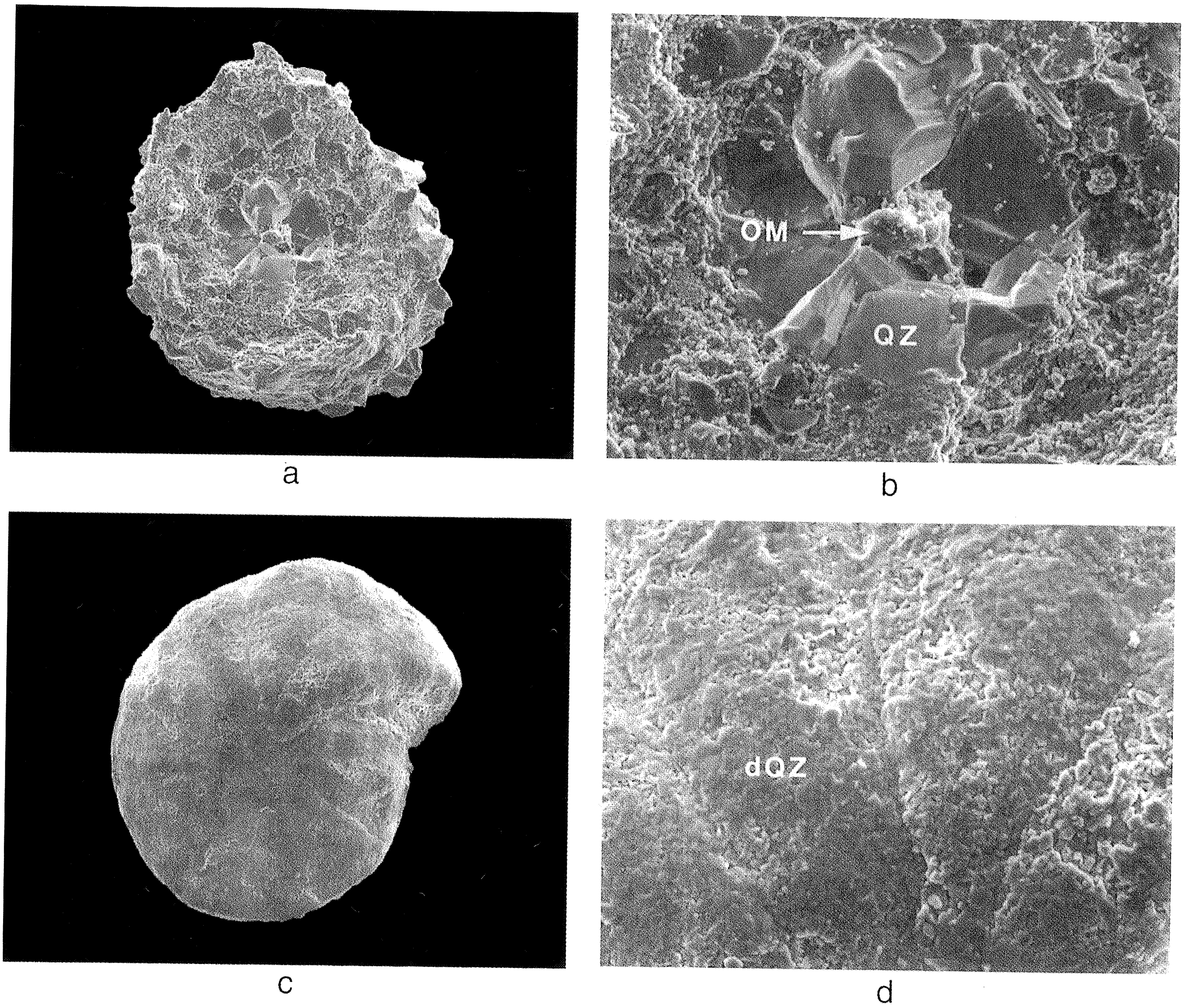



**Fig. 13.** Secondary quartz in agglutinated foraminifers. **a, b.** Quartz crystals (QZ) and organic matter (OM) filling the interior of *Rhabdammina*, GSC 109543; **c.** *Haplophragmoides*, GSC 190542, lateral view; **d.** detail of mottled texture on exterior of *Haplophragmoides*; smooth areas infilled by diagenetic quartz (dQZ)

were also lined on the interior by diagenetic calcite crystals. These rocks are at present day burial depths of about 400 metres and are undoubtedly immature from a low temperature regime which would be consistent with the vermiform style of low temperature kaolinite crystallization (Osborne *et al.*, 1994). In the Beaufort–Mackenzie Basin, kaolinite was also observed within the walls of agglutinated foraminifers at present day burial depths of 3197 m (vermiform?) and 3590 m (blocky) in the Amauligak J-44 well (Fig. 8). The original temperature of crystallization is unknown, but the present temperatures are approximately 100°C and the kaolinite crystals appear to have followed quartz overgrowths indicative of the burial diagenetic regime and temperatures in excess of 70° to 80°C.

## CONCLUSIONS AND SUMMARY

1) Foraminifers are affected by numerous chemical and physical phenomena as they become buried by sediments and enter the fossil record. Initially, they are affected by the processes of early diagenesis. Later in their burial history, they become affected by larger scale processes that are fundamental in the evolution of sedimentary basins and in some cases the generation of hydrocarbons.

2) In a broad perspective, many diagenetic processes in sedimentary basins can be categorized into four regimes: (i) early diagenetic, (ii) burial, (iii) overpressured, and (iv) meteoric.

3) The Beaufort–Mackenzie Basin of Arctic Canada contains well developed examples of each of the four diagenetic regimes. This basin is filled by up to 15 km of ter-

rigenous clastic sediments that are actively subsiding. It also contains assemblages of abundant calcareous and agglutinated benthic foraminifers that have been subjected to various physical and chemical conditions in each of the four diagenetic regimes. The basin is also flanked by uplifted rocks that preserve the relict effects of diagenetic processes in older sedimentary basins. For these reasons, it is an excellent natural laboratory for diagenetic analysis with applications to other terrigenous clastic basins.

4) Early diagenetic processes are dominated by the decay of organic material through the activity of aerobic and anaerobic bacteria in the upper 10 m of sediment. Production of $CO_2$ may lead to the dissolution of calcareous foraminifers. Bacterial reduction of sulphate and iron often results in pyrite infilling foraminiferal tests.

5) In the burial diagenetic regime, which occurs approximately from depths of 2 to 8 km and temperatures of approximately 75–200°C, foraminifers exhibit many of the classical burial diagenetic changes that affect organic matter, silica, clay minerals, and carbonate. The thermal alteration/colouration of the organic cement (glycosaminoglycan) in agglutinated foraminifers is illustrated by the Foraminiferal Colouration Index (FCI) which provides a standard for assessing thermal maturity. The silicification of agglutinated foraminifers by secondary mineralization of quartz occurs progressively beyond temperatures of about 80°C and burial depths of about 2.5 km. The initiation of silicification is an important diagenetic horizon in sedimentary basins, marking the point at which porosity changes, quartz cementation, and overpressured pore fluids are likely to be encountered. Foraminiferal tests can contain secondary clay minerals such as smectite, illite, and chlorite which are indicative of significant thermally controlled diagenetic events. The progression of mineralogical and textural changes in the foraminiferal test provides criteria for a generalized burial diagenetic zonation.

6) The overpressured diagenetic regime exists where pore fluid pressures exceed the normal hydrostatic pressure expected from the sedimentary column. Preliminary evidence suggests that the thermal alteration of foraminifers may be retarded by overpressured pore fluids. Preliminary evidence also suggests that there may be a relationship between processes that cause overpressure and the silicification of agglutinated foraminifers, and that initiation of silicification may be a precursor of overpressured zones in sedimentary rocks. At the very least silicification probably signals an environment where geopressures are more likely to be encountered.

7) The meteoric regime occurs in the proximal areas of the sedimentary basin and can reach to depths of 2 km. Meteoric waters are oxygenated, corrosive, over-saturated with respect to quartz, and low in salinity. They are capable of dissolving calcareous foraminifers and can produce secondary mineralization in the foraminiferal test in the form of clay minerals such as vermiform kaolinite and euhedral quartz. Secondary mineralization in foraminifers could prove to be a useful tool in recognizing previous and ongoing interactions with meteoric water in sedimentary basins.

8) Foraminifers contain organic matter, biogenic minerals, detrital minerals, and secondary minerals of diagenetic origin. They are involved in a great number of diagenetic processes as their burial progresses through sedimentary basin evolution. Foraminifers are routinely and extensively sampled from subsurface and outcrop sections and examined by optical and SEM techniques. Therefore, foraminifers are readily available and potentially informative in regard to the thermal and diagenetic history of sedimentary basins.

## Acknowledgements

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## Streszczenie

### WPŁYW PROCESÓW DIAGENETYCZNYCH NA ZACHOWANIE SKORUPEK OTWORNIC W OSADACH BASENU BEAUFORTA–MACKENZIE I PRZYLEGŁYCH BASENACH OBSZARÓW KRATONICZNYCH

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Wiele czynników chemicznych i fizycznych oddziaływuje na mikrofaunę, w tym również na otwornice, w czasie, gdy zostaje ona pogrzebana w osadzie, dając początek prosesom fosylizacyjnym. Późnej, w trakcie jej dalszego pogrzebania, na stan zachowania skorupek wpływa szereg procesów diagenetycznych, związanych z ewolucją basenów sedymentacyjnych, a w niektórych przypadkach również z powstawaniem węglowodorów. Prosesy te można zaklasyfikować do 4 podstawowych środowisk związanych z: (i) wczesną diagenezą, (ii) pogrzebaniem osadów, (iii) nadciśnieniem płynów porowych i (iv) działalnością wód meteorycznych.

Basen Morza Beauforta i Zatoki Mackenzie w arktycznej części Kanady (Fig. 1) daje przykłady oddziaływania tych czterech grup procesów diagenetycznych na skorupki otwornic (Fig. 2). Basen wypełniony jest seriami bardzo miąższych (do 15 km) osadów klastycznych podlegających aktywnej subdukcji. Obszar ten jest otoczony serią osadów, w których zachowały się ślady procesów diagenetycznych zachodzących w starszych basenach sedymentacyjnych.

Procesy wczesnej diagenezy (Fig. 3) są zdominowane przez rozkład materii organicznej spowodowany działalnością bakterii aerobowych i anaerobowych w górnej, dziesięciometrowej warstwie osadu. Wytwarzanie $CO_2$ w trakcie tych procesów może powodować rozpuszczanie węglanowych skorupek otwornic. Redukcja siarczanów i żelaza przez bakterie powoduje w większości przypadków wypełnienie skorupek przez piryt.

Procesy diagenetyczne związane z pogrzebaniem osadu występują od głębokości około 2 do 8 km, gdy temperatura wzrośnie do około 75–200°C. W takich warunkach zmianom podlegają składniki mineralne budujące skorupki otwornic, tj. materia organiczna, krzemionka, minerały ilaste i węglany. Zmiany termiczne cementu organicznego skorupek otwornic aglutynujących (glukozoaminoglikan) można zilustrować poprzez wskaźnik zmiany barwy (Foraminiferal Colouration Index; Tabl. 1, 2; Fig. 6), który pozwala na oszacowanie dojrzałości termicznej osadu (McNeil *et al.*, 1996). Sylifikacja skorupek otwornic aglutynujących związana z wtórną mineralizacją kwarcową (Fig. 4) rozpoczyna się w temperaturze wyższej niż 80°C, co odpowiada głębokości większej niż 2,5 km. Początek tych procesów jest ważnym horyzontem diagenetycznym w basenach sedymentacyjnych (Fig. 7), wskazującym moment gdy zmienia się porowatość osadów, a nadciśnienie płynów porowych jest już znaczne. Skorupki otwornic mogą zawierać wtórne minerały ilaste, takie jak smektyt, illit i chloryt (Fig. 8–10), które są wskaźnikami procesów diagenetycznych, związanych ze znacznymi zmianami temperatury. Postępujące zmiany składu mineralnego i tekstury w skorupkach otwornic dostarczyły kryteriów do określenia poziomów pogrzebania osadu (McNeil *et al.*, 1996).

Procesy diagenetyczne związane z nadciśnieniem płynów porowych oddziaływują w strefie gdzie ciśnienie płynów przekracza normalne ciśnienie hydrostatyczne w kolumnie osadu (Fig. 10). Wstępne wyniki badań wskazują, że przemiany skorupek otwornic pod wpływem temperatury, mogą zostać spowolnione przez oddziaływanie nadciśnienia płynów porowych.

Mogą istnieć wzajemne zależności pomiędzy procesami, które powodują nadciśnienie płynów a sylifikacją skorupek otwornic aglutynujących. Początek sylifikacji może być wskaźnikiem początku oddziaływania tej grupy czynników diagenetycznych (Fig. 10). Jednocześnie sylifikacja prawdobodobnie sygnalizuje znaczny wzrost ciśnienia geostatycznego.

Szereg procesów diagenetycznych związanych z wodami meteorycznymi występuje na proksymalnach obszarach basenów sedymentacyjnych i może oddziaływać do głębokości 2 km (Fig. 11). Wody meteoryczne są natlenione, o właściwościach korodujących, przesycone względem kwarcu i słabo zasolone. Mogą one rozpuszczać węglanowe skorupki otwornic (Fig. 12) i mogą powodować wtórną mineralizację wewnątrz skorupek, w postaci niektórych minerałów ilastych, jak "robaczkowe" kryształy kaolinitu, czy euhedralne kryształy kwarcu. Wtórna mineralizacja skorupek otwornic mogłaby zatem być użyteczna w rozpoznanawaniu kolejnych reakcji z wodami meteorycznymi migrującymi w osadach basenów sedymentacyjnych.