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# Linking diagenesis to sequence stratigraphy: an integrated tool for understanding and predicting reservoir quality distribution

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**ABSTRACT**

Sequence stratigraphy is a useful tool for the prediction of primary (depositional) porosity and permeability. However, these primary characteristics are modified to variable extents by diverse diagenetic processes. This paper demonstrates that integration of sequence stratigraphy and diagenesis is possible because the parameters controlling the sequence stratigraphic framework may have a profound impact on early diagenetic processes. The latter processes play a decisive role in the burial diagenetic and related reservoir-quality evolution pathways. Therefore, the integration of sequence stratigraphy and diagenesis allows a proper understanding and prediction of the spatial and temporal distribution of diagenetic alterations and, consequently, of reservoir quality in sedimentary successions.

## INTRODUCTION

The diagenesis of sedimentary rocks, which may enhance, preserve or destroy porosity and permeability, is controlled by a complex array of interrelated parameters (Stonecipher *et al.*, 1984). These parameters range from tectonic setting (controls burial-thermal history of the basin and detrital composition of clastic sediments) to depositional facies and palaeo-climatic conditions (Morad, 2000; Worden & Morad, 2003). Despite the large number of studies (e.g. Schmidt & McDonalds, 1979; Stonecipher *et al.*, 1984; Jeans, 1986; Curtis, 1987; Walderhaug & Bjorkum, 1998; Ketzer *et al.*, 2003; Shaw & Conybeare, 2003) on the diagenetic alteration of sedimentary rocks, the parameters controlling their spatial and temporal distribution patterns in paralic and shallow-marine and particularly in continental and deep water sedimentary deposits are still not fully understood (Surdam *et al.*, 1989; Morad, 1998; Worden & Morad, 2000, 2003).

Diagenetic studies have been used independently from sequence stratigraphy as a tool to understand and predict the distribution of reservoir quality in clastic and carbonate successions (e.g. Ehrenberg, 1990; Byrnes, 1994; Wilson, 1994; Bloch & Helmold, 1995; Kupecz *et al.*, 1997; Anjos *et al.*, 2000; Spötl *et al.*, 2000; Bourque *et al.*, 2001; Bloch *et al.*, 2002; Esteban & Taberner, 2003; Heydari, 2003; Prochnow *et al.*, 2006; Ehrenberg *et al.*, 2006a).

The sequence stratigraphic approach, nevertheless, allows the prediction of facies distributions (Posamentier & Vail, 1988; Van Wagoner *et al.*, 1990; Emery & Myers, 1996; Posamentier & Allen, 1999), providing information on the depositional distribution of primary porosity and permeability (Van Wagoner *et al.*, 1990; Posamentier & Allen, 1999). Depositional reservoir quality is mainly controlled by the geometry, sorting and grain size of sediments. Sequence stratigraphy enables prediction of the distribution of mudstones and other fine-grained deposits that may act as seals, baffles and barriers for fluid flow within reservoir successions and as petroleum source rocks (Van Wagoner *et al.*, 1990; Emery & Myers, 1996; Posamentier & Allen, 1999).

Although sequence stratigraphic models can predict facies and depositional porosity and permeability distribution in sedimentary successions, particularly in deltaic, coastal and shallow-marine deposits (Emery & Myers, 1996), they cannot provide direct information about the diagenetic evolution of reservoir quality. As most of the controls on early diagenetic processes are also sensitive to relative sea-level changes (e.g. pore water compositions and flow, duration of subaerial exposure), diagenesis can be linked to sequence stratigraphy (Tucker, 1993; South & Talbot, 2000; Morad *et al.*, 2000, 2010; Ketzer *et al.*, 2002, 2003). Hence, it is logical to assume that the integration of diagenesis and sequence stratigraphy will constitute a powerful tool for the prediction of the spatial and temporal distribution and evolution of quality in clastic reservoirs, as it has already been developed for carbonate successions (Goldhammar *et al.*, 1990; Read & Horbury, 1993 and references therein; Tucker, 1993; Moss & Tucker, 1995; South & Talbot, 2000; Bourque *et al.*, 2001; Eberli *et al.*, 2001; Tucker & Booler, 2002; Glumac & Walker, 2002; Moore, 2004; Caron *et al.*, 2005). This approach can also provide useful information on the formation of diagenetic seals, barriers and baffles for fluid flow, which may promote diagenetic compartmentalization of the reservoirs. A limited number of studies has been undertaken that illustrate how the spatial distribution of diagenetic features in various types of sedimentary successions can be better understood when linked to a sequence stratigraphic framework (Read & Horbury, 1993 and references therein; Tucker, 1993; Moss & Tucker, 1995; Morad *et al.*, 2000; Ketzer *et al.*, 2002, 2003a, 2003b, 2005; Al-Ramadan *et al.*, 2005; El-Ghali *et al.*, 2006, 2009).

Carbonate sediments are more reactive than siliciclastic deposits to changes in pore-water chemistry caused by changes in relative sea-level versus rates of sediment supply (i.e. regression and transgression) (Morad *et al.*, 2000). Therefore, the distribution of diagenetic alterations can be more readily linked to the sequence stratigraphic framework of carbonate than of siliciclastic deposits (Tucker, 1993; McCarthy & Plint, 1998; Bardossy & Combes, 1999; Morad *et al.*, 2000). Cool-water limestones are commonly composed of low-Mg calcite and thus are less reactive than tropical limestones, which are composed of the metastable aragonite and high-Mg calcite. In tropical carbonate rocks, particularly, the distribution of diagenetic alterations can be recognized within third (1–10 Ma) or fourth (10s ky to 100 ky) order cycles of relative sea-level change (Tucker, 1993), whereas in siliciclastic deposits only alterations relative to third order cycles can be recognized (Morad *et al.*, 2000). Less commonly, however, diagenetic alterations can be correlated to smaller cycles (parasequences; Van Wagoner *et al.*, 1990) within third order sequences (Taylor *et al.*, 1995; Loomis & Crossey, 1996; Klein *et al.*, 1999; Ketzer *et al.*, 2002). The low rates of subsidence in marine epicontinental environments (Sloss, 1996) render linking diagenesis to sequence stratigraphy difficult.

In the following discussion, definitions of the diagenetic stages *eodiagenesis*, *mesodiagenesis* and *telodiagenesis sensu* Morad *et al.* (2000) will be applied to clastic successions, whereas the original definitions of these stages (Choquette & Pray, 1970) are applied to carbonate successions. According to Morad *et al.* (2000), eodiagenesis includes processes developed under the influence of surface or modified surface waters such as marine, mixed marine-meteoric, or meteoric waters, at depths $<2$ km ($T < 70\,^{\circ}C$), whereas mesodiagenesis includes processes encountered at depths $>2$ km ($T > 70\,^{\circ}C$) and reactions involving chemically evolved formation waters. Shallow mesodiagenesis corresponds to depths between 2 and 3 km and to temperatures between 70 and 100 °C. Deep mesodiagenesis extends from depths of $\sim$3 km and temperatures $\sim$100 °C to the limit of metamorphism, corresponding to temperatures $>$200 °C to 250 °C and to highly-variable depths, according to the thermal gradient of the area. Telodiagenesis refers to those processes related to the uplift and exposure of sandstones to near-surface meteoric conditions, after burial and mesodiagenesis. In the original definitions of Choquette & Pray (1970) there is no depth or temperature limit between eodiagenesis and mesodiagenesis, but only a vague effective burial limit, defined as the case-specific depth below which the surface fluids cannot reach and influence the sediments and there is no distinction between shallow and deep mesodiagenesis.

The goals of this paper are to: (i) demonstrate that the distribution of diagenetic alterations in sedimentary successions can, in many cases, be systematically linked to sequence stratigraphy, (ii) highlight the most common diagenetic alterations related to specific systems tracts and to key sequence-stratigraphic surfaces and (iii) apply these concepts to prediction of the spatial and temporal distribution of reservoir quality in carbonate and clastic successions.

## SEQUENCE STRATIGRAPHY: AN OVERVIEW OF THE KEY CONCEPTS

In order to emphasize the impact of rates of changes in relative sea-level versus rates of sedimentation on the distribution of diagenetic alterations in siliciclastic and carbonate sediments, it is worthwhile to provide a brief overview of the concepts and basic definitions of sequence stratigraphy. Sequence stratigraphy is the analysis of genetically-related strata within a chronostratigraphic framework. The stacking patterns of these strata are controlled by the rates of changes in relative sea-level (i.e. accommodation creation or destruction caused by subsidence/uplift and/or changes in the eustatic sea-level) compared to rates of sediment supply.

There are genetic differences in the sequence stratigraphic models developed for shallow-marine and paralic siliciclastic and carbonate successions related to: (i) origin of sediments. Siliciclastic sediments are derived mostly from outside the depositional basin and are thus influenced by lithology, tectonic setting and climatic conditions in the hinterlands (Dickinson *et al.*, 1983; Dutta & Suttner, 1986; Suttner & Dutta, 1986). Conversely, marine carbonate sediments are produced by organic and inorganic intrabasinal processes (Hanford & Loucks, 1993). (ii) Carbonate sediments are commonly produced at higher rates than siliciclastic sediments and respond differently to changes in the relative sea-level compared to siliciclastic deposits (Hanford & Loucks, 1993). (iii) Transgression coincides with higher rates of carbonate sedimentation, whereas the opposite is true regarding siliciclastic sediments. Therefore, the sequence stratigraphic framework of carbonate successions differs from that of siliciclastic successions (Hanford & Loucks, 1993; Boggs, 2006). (iv) Subaerially exposed carbonate sediments are subjected to dissolution by meteoric waters, i.e. little sediment is produced. Conversely, exposed siliciclastic deposits can be subjected to valley incision and deposition of the reworked sediments at and beyond the shelf break. Moreover, the incised valleys can act as sites for the deposition of fluvial and estuarine deposits.

The sequence stratigraphic terminology of carbonate and siliciclastic deposits presented here is largely based on the concepts introduced by Vail (1987), Posamentier *et al.* (1988) & Van Wagoner *et al.* (1990), but taking into account revisions and critical evaluations of these concepts (Sarg, 1988; Loucks & Sarg, 1993; Emery & Myers, 1996; Miall, 1997; Posamentier & Allen 1999; Catuneanu, 2006). The examples of links between diagenesis and sequence stratigraphy presented in this paper fall within the framework of so-called high-resolution sequence stratigraphy (1 to 3 Ma; Emery & Myers, 1996).

The basic principle of sequence stratigraphy is that the deposition of sediments and their spatial and temporal distribution in a basin are controlled by the interplay between the rates of: (i) sediment supply, (ii) basin-floor subsidence and uplift and (iii) changes in the eustatic sea-level (e.g. Vail, 1987; Posamentier *et al.*, 1988; Van Wagoner *et al.*, 1990). These parameters control the space within a basin that is available for sediment deposition and preservation, i.e. accommodation (Jervey, 1988). Accommodation in shallow marine environments can be created by a rise in the eustatic sea-level and/or to basin-floor subsidence. This is referred to as relative sea-level rise. Fall in the relative sea-level is caused by fall in the eustatic sea-level and/or tectonic uplift.

The stacking pattern of sedimentary packages depends on rates of accommodation creation versus rates of sediment supply (Fig. 1; Posamentier *et al.*, 1988; Van Wagoner *et al.*, 1990). If the rate of sediment supply exceeds the rate of accommodation creation, the sediment stacking will be progradational , which is referred to as normal regression (Fig. 1A). Regression may also occur either due to fall in the relative sea-level (owing to a fall in eustatic sea-level and/or tectonic uplift of basin floor), also referred to as forced regression, being characterized by a 'downstepping' geometry of the facies (Fig. 1B). Conversely, retrogradational stacking patterns are developed by lower rates of sediment supply lower than rate of accommodation creation (i.e. transgression). The shoreline will migrate landward and the vertical facies succession display an upward deepening trend (i.e., backstepping; Fig. 1C). Aggradation of depositional facies occurs if the rate of sediment supply is equivalent to the rate of accommodation creation (Fig. 1D). In this case, deposits will keep fixed position upwards in the stratigraphic section. Rates of sediment supply across a carbonate platform depend on the productivity of the carbonate factory, which depends on sea water temperature, salinity, water depth, rate of siliciclastic sediment input and nutrient supply (Hallock & Schlager, 1986). The rates of siliciclastic sediment supply depend largely on climatic conditions (i.e. rates of chemical weathering) and tectonic setting (e.g. rates of uplift, lithology of source rocks).

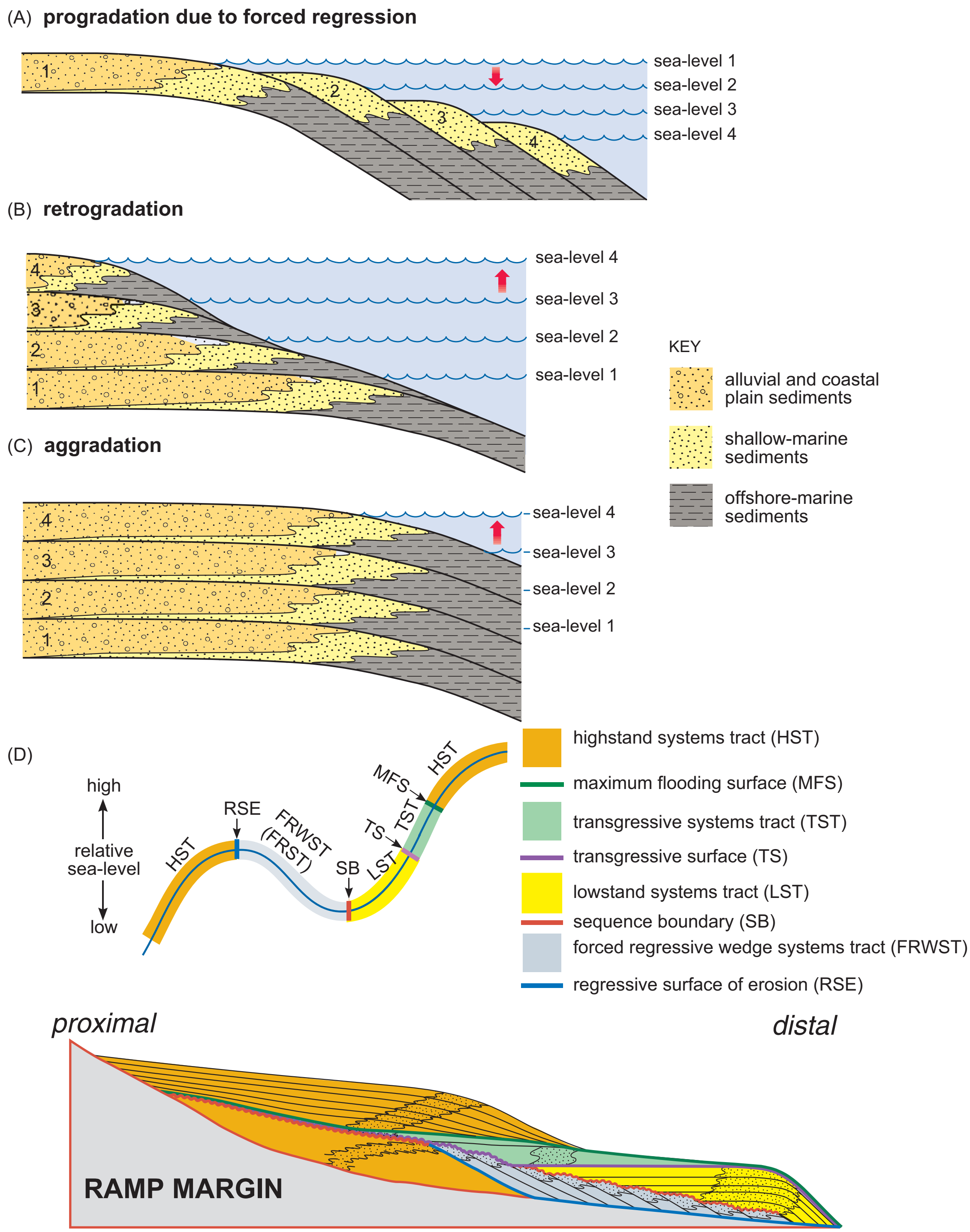


**Fig. 1.** Diagram showing the major stacking patterns of parasequences (A)–(D) and sequence stratigraphic features. (A) Progradational parasequence sets resulting from forced regression caused by substantial sediment supply derived from subaerial erosion and fluvial incision into the previously deposited sediments during sea-level fall. (B) Retrogradational parasequence sets formed when the increase in the rate of accommodation creation is larger than the rate of sediment supply. (C) Aggradational parasequence sets resulting from similar rates of sediment supply and accommodation creation. (D) Progradational parasequences showing a shallowing upward pattern bounded by marine flooding surfaces. The schematic representation of the four systems tracts, including lowstand (LST), transgressive (TST), highstand (HST) and forced regressive wedge (FRWST), also referred to simply as forced regressive (FRST). Modified after Coe (2003).

Sequence stratigraphic analysis aims to divide the sedimentary record into depositional sequences, in which the sequence boundaries are subaerial erosion surfaces (unconformities) or their correlative conformities. Sequence boundaries are formed by a rapid fall in relative sea-level (Van Wagoner *et al.*, 1990). Thus, sequences are deposited between two episodes of relative sea-level fall, which coincide, for instance, with falling inflection points on a hypothetical relative sea-level curve (Fig. 1D). If relative sea-level eventually falls below the shelf edge, valley incision, pronounced erosion of, particularly siliciclastic, shelves and deep-water turbidite deposition will occur (Posamentier & Allen, 1999). Changes in relative sea-level in carbonate depositional systems may result in exposure of the platform, stopping the carbonate factory and leading to karstification, particularly under humid climatic conditions.

Sequences are composed of systems tracts, which are, in turn, composed of parasequences (Fig. 1D). Parasequences are relatively conformable successions of genetically related beds or bedsets bounded by 'minor' marine flooding surfaces and their correlative surfaces (Van Wagoner *et al.*, 1990). A parasequence set is a succession of genetically related parasequences, which display progradational, aggradational or retrogradational stacking pattern. Hence, parasequence sets reflect the interplay between rates of deposition and accommodation creation (Van Wagoner *et al.*, 1990). If the deposition rate is higher than the accommodation creation rate, the parasequence set will be progradational, whereas if the deposition rate is equal to or lower than the accommodation creation rate, then the parasequence set is aggradational or retrogradational, respectively (Figs. 1A–D).

Parasequence sets can be associated to a specific segment of a relative sea-level curve and comprise system tracts. Systems tracts are defined as the contemporaneous depositional systems linked to a specific segment on the curve of changes in the relative sea-level (Fig. 1D). Each systems tract is defined by stratal geometries at bounding surfaces, position within the sequence and internal parasequence stacking patterns. Four main systems tracts have been described in the literature (Vail *et al.*, 1977; Van Wagoner *et al.*, 1990; Hunt & Tucker, 1992): lowstand systems tract (LST), transgressive systems tract (TST), highstand systems tract (HST) and forced regressive wedge systems tract (FRWST; Fig. 1D).

LST deposits are formed during a fall in relative sea-level (i.e. retreat of the shoreline), which results in subaerial exposure of the shelf. Sediment supply on siliciclastic shelf margin/slope is often maintained and delivered via incised valleys and redistributed via fluvial-deltaic processes (Vail *et al.*, 1977; Van Wagoner *et al.*, 1990; Handford & Louks, 1993). Carbonate sediment production by the carbonate factory is terminated or restricted to shelf margins and upper slopes. LST deposits have thus progradational parasequence sets, particularly in siliciclastic successions, as carbonate factory stops during exposure of the shelf (Posamentier *et al.*, 1992). Major Fall in the relative sea-level also causes deep submarine channel incisions on the slope of siliciclastic shelves and carbonate platforms/banks (Anselmmeti *et al.*, 2000). LST deposits include fluvial-deltaic siliciclastic deposits and shallow-marine siliciclastic and carbonate deposits include shelf margin, slope and basin-floor turbidite and debris-flow deposits.

LST deposits are bounded below by a sequence boundary (SB) and above by a transgressive surface (TS), which marks the beginning of rapid rise in relative sea-level. The SB and TS are amalgamated in shelf sites where there was little deposition and/or erosion. The TS is often marked by the occurrence of conglomeratic lag deposits, which are formed by reworking of shelf sediments by marine currents. The TST sediments are deposited due to higher rates of relative sea-level rise than rates of sediment supply, which is accompanied by landward migration of the shoreline (i.e. transgression) and of loci of siliciclastic sediment deposition. A rapid rise in the relative sea-level and deepening of water to depths greater than the photic zone may drown and shut down the carbonate factory. Conversely, slow transgression may allow the platform to remain within the photic zone and the carbonate factory to maintain carbonate sediment production. The impact of transgression on sediment production by the carbonate factory is important when occurring immediately after establishment of highstand conditions, i.e. after establishment of active carbonate factory (Catuneanu, 2005). Termination of the carbonate factory owing to drowning of the platform below the photic zone is followed by deposition of siliciclastic mud (Catuneanu, 2005). The TST deposits are bounded below by the TS and above by the maximum flooding surface (MFS), which corresponds to maximum landward advance of the shoreline (Fig. 1D).

The TS is frequently marked by the presence of coarse-grained lag deposits composed of algal bored and encrusted intrabasinal fragments derived by marine erosion of sediments (i.e. ravinement), which include palaeosol, calcrete, bioclasts and/or highstand mudstones exposed along SB on shelves/platforms (Sarg, 1988; Hunt & Tucker, 1992; Hanford & Louks, 1993). Ravinement is expected to be more pronounced in open-marine shelves, whereas insignificant in rimmed shelves, which remain subaerially exposed during rapid rise in the relative sea-level (cf. Handford & Louks, 1993). The formation of TS is commonly followed by re-establishment of the carbonate factory across the shelf, including shoreward accretion of subtidal carbonate sediments over shallow-water sediments (Handford & Louks, 1993). However, carbonate sediment production usually lags behind rise in relative sea-level, which gives way, on some mixed siliciclastic-carbonate shelves, to deposition of siliciclastic sediments followed by carbonate sediments (Handford & Louks, 1993).

The MFS represents a condensed section (hiatus surface) formed by faster rates of relative sea-level rise than rates of sedimentation, particularly in the middle and outer shelf. The TST is comprised of retrogradational (backstepping) parasequence set of shelf sediments, including shallow-marine sandstones and mudstones. Peat (coal) layers are developed during transgression of coastal plain under humid climatic conditions in both carbonate and siliciclastic successions (de Wet *et al.*, 1997).

The HST is deposited during late stages of rise, stillstand and early stages of falling relative sea-level. The HST package is bounded below by MFS and above by the upper SB. The HST is comprised of initially aggradational and later, as the rates of accommodation creation by rise in the relative sea-level diminishes, of progradational parasequence sets. Sediment production by carbonate factory is greatest during highstand, because of the slow rates of drowning of the platform (Handford & Louks, 1993). Growth of carbonate rims in shelves may result in the development of lagoons with restricted connection with the open sea encouraging deposition of evaporites under arid climatic conditions. The HST record is only partly preserved owing to erosion during the next cycle of fall in relative sea-level and formation of upper sequence boundary. The FRWST, also known as falling-stage systems tract (FSST) was proposed (Hunt & Tucker, 1992) to include deposits formed during relative sea-level fall, between the highstand and the point of maximum rate of sea-level fall (i.e., formation of the succeeding sequence boundary). The most typical sediments of FSST are sharp-based sandstones deposited in shoreface environments above erosional surfaces formed during regression (Plint, 1988). The sequence boundary is usually drawn above the FSST (the subaerial unconformity and its seaward extension), because this surface is formed when the relative sea-level reaches its lowest point and it coincides with the surface of subaerial exposure.

## PARAMETERS CONTROLLING SEDIMENT DIAGENESIS

The diagenesis of siliciclastic and carbonate sediments is controlled by a complex array of interrelated parameters, many of which are not related directly to the interplay between rates of changes in the relative sea-level versus rates of sediment supply and thus cannot be constrained only within a sequence stratigraphic context. These parameters include the tectonic setting, which controls: (i) basin type and the burial, temperature and pressure histories, (ii) relief and lithology of source rocks, which exert direct control on detrital composition of sandstones (Siever, 1979; Dickinson, 1985; Ingersoll, 1988; Zuffa, 1987; Horbury & Robinson, 1993) and (iii) depositional setting (Fig. 2A). The depositional setting controls both the primary composition and textures of carbonate sediments and hence most diagenetic processes (Fig. 2B). Tectonic setting exerts a less direct influence on the diagenetic processes of carbonate successions, as their primary composition is a product of intrabasinal processes.

The tectonic setting of the basin controls the rates of sediments supply and depth of meteoric water incursion in the basin (Fig. 2A). Under high sediment supply rates typical of tectonically active settings, such as in rift or forearc basins, there is smaller opportunity for eogenetic reactions to occur and therefore for sequence stratigraphic control on diagenesis. Detrital sand composition strongly influences the types, distribution and patterns of clastic diagenetic processes (Fig. 2A; Surdam *et al.*, 1989; De Ros, 1996; Primmer *et al.*, 1997).

Other important parameters that influence the diagenesis include palaeoclimatic conditions. The role of palaeoclimatic conditions is most

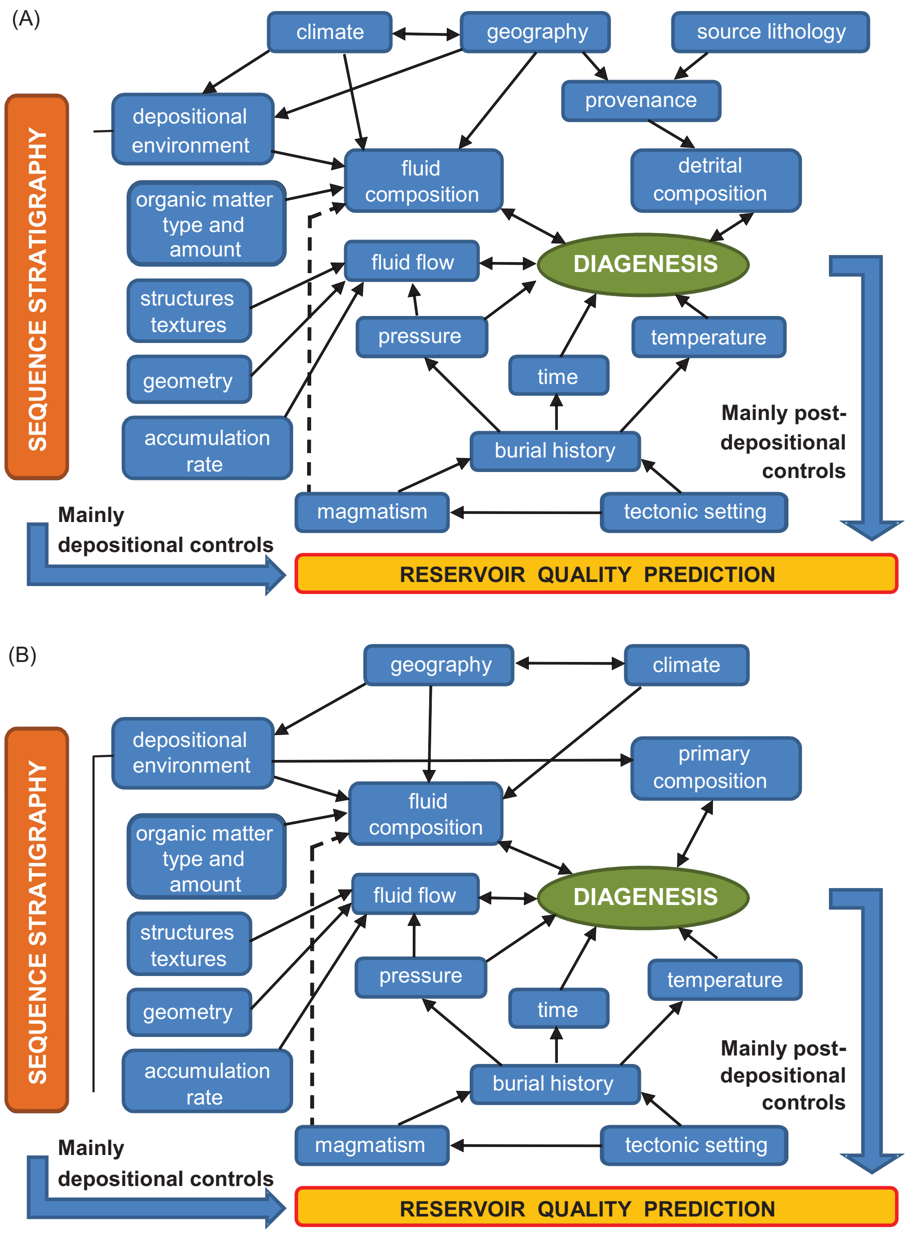


**Fig. 2.** Diagram showing the complex array of factors controlling the diagenesis of clastic (A) and carbonate (B) sediments. Sequence stratigraphy can provide useful information on depositional environment, structures, texture and composition, which directly control the diagenetic processes and patterns.

prevalent during relative sea-level fall and partial to complete exposure of the shelf, which results in meteoric water incursion into the paralic and shallow-marine deposits (Hutcheon *et al.*, 1985; Searl, 1994; Thyne & Gwinn, 1994; Worden *et al.*, 2000). The impact of meteoric water incursion into these sediments is more important under warm, humid climatic conditions than under arid to semi-arid conditions.

## BASIS FOR LINKING DIAGENESIS AND SEQUENCE STRATIGRAPHY

Linking diagenesis to sequence stratigraphy is possible because parameters controlling the sequence stratigraphic framework of sedimentary deposits, including primarily the rates of changes in the relative sea-level (interplay between tectonic subsidence/uplift and changes in the eustatic sea-level) versus rates of deposition (Van Wagoner *et al.*, 1990; Posamentier & Allen 1999), also exert profound impact on parameters that control the near-surface diagenetic alterations in these deposits, including:

(i) Changes in pore-water chemistry. Pore-water chemistry varies during near-surface eodiagenesis among marine, brackish and meteoric compositions (Hart *et al.*, 1992; Tucker, 1993; Morad, 1998; Morad *et al.*, 2000, 2010). Pore-water chemistry is the master control on a wide range of diagenetic reactions, including cementation, dissolution and neomorphism of carbonate and dissolution and kaolinization of framework silicates (Curtis, 1987; Morad *et al.*, 2000).

(ii) Residence time. The residence time of sediments under specific geochemical conditions is established as a consequence of regression and transgression. Prolonged subaerial exposure of the sediments during regression results in extensive meteoric water incursion, particularly under humid climatic conditions (Loomis & Crossey, 1996; Ketzer *et al.*, 2003). Typical diagenetic reactions encountered are dissolution of marine carbonate cements and kaolinization of chemically unstable silicates (e.g. micas and feldspars). Conversely, low sedimentation rates on the shelf results in prolonged residence time of sediments at and immediately below the seafloor and hence extended marine pore water diagenesis, which is probably mediated by diffusive mass exchange between pore waters and the overlying sea water (Kantorowicz *et al.*, 1987; Wilkinson, 1991; Morad *et al.*, 1992; Amorosi, 1995; Taylor *et al.*, 1995; Morad *et al.*, 2000). Thus, variations in residence time control the extent of diagenetic alterations under the prevailed geochemical conditions.

(iii) Variation in the framework grain composition. Transgression and regression may cause changes the proportion of extra-basinal and intra-basinal grains (Dolan, 1989; Fontana *et al.*, 1989; Garzanti, 1991; Amorosi, 1995; Zuffa *et al.*, 1995; Morad *et al.*, 2000, 2010). Framework grain composition controls the mechanical and chemical properties and hence the burial diagenetic alterations and related reservoir-quality evolution pathways of arenites (Fig. 3). Intrabasinal carbonate (bioclasts, peloids, ooids and intraclasts) and non-carbonate (e.g., glaucony peloids, berthierine ooids, mud intraclasts and phosphate; Zuffa, 1985, 1987) grains increase relatively in abundance upon marine transgression (Fig. 3). Transgressions promote the flooding of shelf areas, dramatically increasing the sites available for the generation of carbonate grains and starve extrabasinal sediment supply to the shelf edge, thereby favouring the formation of glaucony and phosphate. In contrast, regressions decrease or even shut-off the production of these grains, favouring increased erosion and redistribution of extrabasinal siliciclastic sediments (Dolan, 1989).

(iv) Organic matter content in sediments. Transgression and regression have also profound impact on the amounts and types of organic matter (Cross, 1988; Whalen *et al.*, 2000), which control, in turn, the redox potential of pore waters and consequently the oxidation-reduction reactions in the host sediments (Coleman *et al.*, 1979, Curtis, 1987; Hesse, 1990; Morad, 1998). Planktonic productivity and hence the amount of reactive marine organic matter in marine sediments, increases in abundance during transgression (Pedersen & Calvert, 1990; Bessereau & Guillocheau, 1994; Whalen *et al.*, 2000; Sutton *et al.*, 2004). Highly reactive organic-matter content in paralic and marine sediments causes rapid, progressive depletion of pore waters in dissolved oxygen below the sediment-water interface, i.e. progressively more reducing geochemical conditions (Froelich *et al.*, 1979; Berner, 1981). These conditions have profound impact on the formation of Fe-rich and Mn-rich minerals, such as pyrite, siderite, Fe-dolomite and Fe-silicates (Curtis, 1987; Morad, 1998).

Extracting valuable information about these parameters from sequence stratigraphic analyses should, hence, allow constraining diagenesis and related reservoir-quality evolution of sandstones below sequence and parasequence boundaries and marine flooding, transgressive and maximum flooding surfaces and within systems tracts (Morad et al., 2010).

In the following sections, the types, distribution patterns and impacts of diagenetic processes and products will be discussed for carbonate (Table 1) and siliciclastic (Table 2) deposits in relation to the main sequence stratigraphic surfaces and systems tracts.

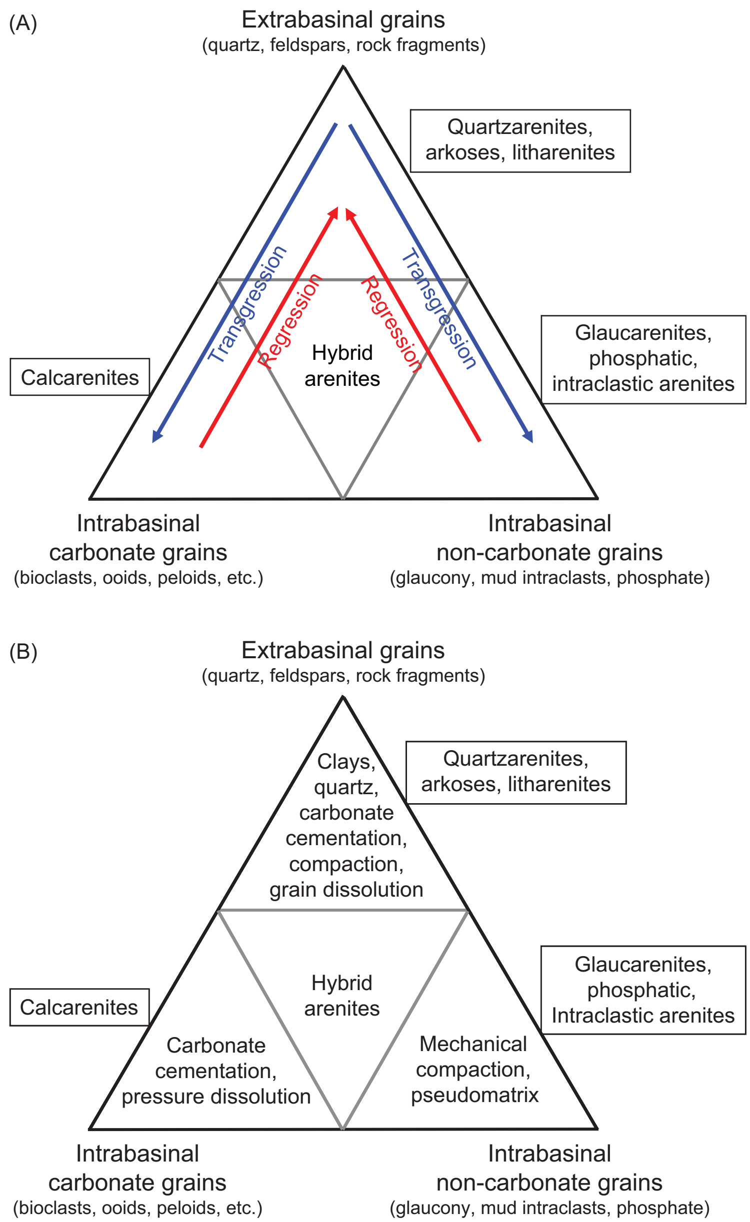


**Fig. 3.** Variations in relative proportions of extrabasinal and intrabasinal grains corresponding to transgression and regression (A) and major diagenetic processes observed in siliciclastic sandstones and intrabasinal arenites (B), shown on Zuffa (1980) diagram. Hybrid arenites usually display mixed diagenetic processes corresponding to their compositional constituents.

## DISTRIBUTION OF DIAGENETIC ALTERATIONS ALONG SEQUENCE STRATIGRAPHIC SURFACES

The distribution of diagenetic alterations along the key sequence stratigraphic surfaces (i.e. SB, PB, TS and MFS) occurs owing to more significant increase in the rates of relative sea-level rise than rates of sedimentation. Hence, considerable shifts in the parameters controlling diagenesis are encountered along these surfaces, resulting in fairly marked diagenetic alterations (Tucker, 1993; Morad *et al.*, 2000, 2010). For identification and interpretation of diagenetic patterns linked to sequence stratigraphy, it should be kept in mind that the original near-surface, eogenetic alterations

**Table 1.** Summary of major diagenetic processes and products related to sequence stratigraphic controls in carbonate deposits and main impacts on reservoir quality.

| Processes & products | Setting | Reservoir quality impact |
|---|---|---|
| **Sequence boundaries** | | |
| Dissolution and karstification | Subaerial humid | Porosity & permeability enhancement |
| Phreatic meteoric cementation | Subaerial | Major porosity & permeability reduction |
| Dolomite calcitization | Subaerial | No |
| Pedogenesis & calcrete formation | Subaerial | Porosity & permeability reduction; flow barriers |
| Formation of kaolinite & bauxite | Subaerial humid | Minor porosity & permeability reduction |
| Dolomitization (evaporation) | Coastal | Moldic & intercrystalline porosity generation |
| Dolomitization (mixing) | Coastal | Permeability reduction; some porosity generation |
| **Parasequence boundaries, transgressive surfaces, maximum flooding surfaces** | | |
| Dolomitization | Marine | Porosity generation; variable permeability |
| Hardgrounds & firmgrounds | Marine | Porosity & permeability reduction; fluid flow barriers |
| Fe and Mn oxyhydroxide nodules | Marine | No |
| Isopachous Mg-calcite and aragonite cements | Marine | Slight permeability reduction; preservation of intergranular porosity |
| Alternated dolomite and calcite cementation (mixing) | Mixed marine – meteoric | Porosity & permeability reduction |
| Dissolution related to coals on TS and in early TST | Mixed marine – meteoric | Porosity & permeability enhancement |
| **Highstand systems tracts** | | |
| Mg-calcite and aragonite cementation | Shallow marine | Permeability & porosity reduction; partial cementation may help preserve porosity |
| Alternated dolomite and calcite cementation (mixing) | Mixed marine – meteoric | Porosity & permeability reduction |
| Pressure dissolution of carbonate grains | Marine | Permeability and porosity reduction |
| Mg-calcite and aragonite cementation in carbonate grains-rich turbidites | Deep marine | Permeability & porosity reduction; layers rich in carbonate grains constitute flow barriers |
| Meteoric dissolution below SB | Meteoric or mixed | Porosity & permeability enhancement |
| **Transgressive systems tracts** | | |
| Mg-calcite and aragonite cementation | Marine | Decrease of permeability & porosity towards the MFS |
| Dolomitization (seawater) | Marine | Increase in porosity towards the MFS; permeability variable |

in siliciclastic and, particularly, in the highly reactive carbonate sediments are usually subjected to chemical (elemental and isotopic), textural and/or mineralogical modifications during subsequent eodiagenesis, mesodiagenesis and/or telodiagenesis. Such changes include: (i) recrystallization of carbonate cements, which results in decrease of $\delta^{18}O$ and of $\delta^{13}C$ signatures of marine calcite cements, (ii) calcitization of dolomite and dolomitization of calcite and (iii) transformation of clay minerals, such as illitization of kaolinite and chloritization of berthierine and smectite (Morad *et al.*, 2000; Worden & Morad, 2003).

## Sequence boundaries (SB)

Subaerial sediment exposure due to major fall in the relative sea-level (i.e. formation of SB), is accompanied by basinward migration of the meteoric pore water zone (Fig. 4; Morad *et al.*, 2000), which is accompanied by characteristic diagenetic alterations in carbonate and siliciclastic sediments (outlined below). However, the extent and depth of meteoric water flux into siliciclastic and carbonate successions depend on the hydraulic head, tilting of the permeable bed(s), climatic conditions, duration of subaerial exposure, reactivity of the sediments and intensity and connectivity of fracture systems (Galloway, 1984; Worthington, 2001; Burley & MacQuaker, 1992; Longstaffe, 1993; Mátyás & Matter, 1997). Hence, meteoric-water flux below SB is more extensive in unconfined than in confined aquifers (Coffey, 2005).

The extent of shelf exposure as consequence of a fall in the relative sea-level increases with decrease in tilting of the shelf. Fall in the relative sea-level by

**Table 2.** Summary of major diagenetic processes and products related to sequence stratigraphic controls in siliciclastic deposits and main impacts on reservoir quality.

| Processes & products | Setting | Reservoir quality impact |
|---|---|---|
| **Sequence Boundaries** | | |
| Clay infiltration | Continental dry | Permeability reduction; variable porosity reduction; fluid flow barriers |
| Illitization of infiltrated clays | Continental dry | Permeability reduction; pressure dissolution; quartz overgrowth inhibition |
| Chloritization of infiltrated clays | Continental dry | Preservation of intergranular porosity |
| Calcretes and dolocretes | Continental dry | Permeability & porosity reduction; fluid flow barriers |
| Grain dissolution and kaolinization | Continental humid | Porosity & permeability enhancement |
| **Parasequence boundaries, transgressive surfaces, maximum flooding surfaces** | | |
| Stratabound continuous or concretionary calcite, dolomite or siderite cementation | Marine | Porosity & permeability reduction; fluid flow barriers |
| Carbonate cementation of bioclastic or intraclastic lags | Marine | Porosity & permeability reduction; fluid flow barriers |
| Compaction of intraclastic lags to pseudomatrix | Marine | Porosity & permeability reduction; fluid flow barriers |
| Calcite and pyrite cementation along coal layers | Marine | Porosity & permeability reduction; fluid flow barriers |
| Dissolution and kaolinite cementation below coal layers | Marine | Porosity & permeability enhancement |
| Autochthonous glaucony | Marine | Porosity & permeability reduction |
| Odinite coatings | Paralic mixed marine – meteoric | Permeability reduction; chlorite (chamosite) from odinite transformation may preserve porosity |
| Berthierine oolites | Paralic – mixed | Porosity & permeability reduction; may constitute flow barriers |
| **Highstand systems tracts** | | |
| Mg-calcite & aragonite cementation | Shallow marine | Permeability & porosity reduction; partial cementation may help preserving porosity |
| Mg-calcite & aragonite cementation in carbonate grains-rich turbidites | Deep marine | Permeability & porosity reduction; layers rich in carbonate grains constitute flow barriers |
| Meteoric dissolution below SB | Marine-mixing | Porosity & permeability enhancement |
| **Lowstand systems tracts** | | |
| Grain dissolution & kaolinization | Continental Humid | Porosity & permeability enhancement |
| Pore-lining & grain-replacive authigenic smectite | Continental dry | Permeability reduction; porosity reduction or limited preservation |
| Clay infiltration | Continental dry | Permeability reduction; variable porosity reduction; flow barriers |
| Illite from authigenic or infiltrated smectite transformation | Continental dry | Permeability reduction; pressure dissolution; quartz overgrowth inhibition |
| Chlorite from authigenic or infiltrated smectite transformation | Continental dry | Permeability reduction; porosity preservation |
| Compaction of mud intraclasts eroded from HST into pseudomatrix | Marine | Porosity & permeability reduction; flow barriers |
| **Transgressive and early highstand systems tracts** | | |
| Continuous or concretionary stratabound calcite cementation | Marine | Porosity & permeability reduction; continuously-cemented layers constitute fluid flow barriers |
| Pyrite from bacterial sulfate reductions | Marine | No |
| Phosphate cementation, replacement, nodules | Marine | Porosity & limited permeability reduction; commonly restricted to mudstones |
| Autochthonous glaucony | Marine | Porosity & permeability reduction |
| Silica (opal, opal-CT, chalcedony, microquartz) coatings | Marine | Permeability reduction; may help preserve porosity through formation of grain-coating micro-quartz |
| Silica (opal, opal-CT, chalcedony, microquartz) coatings | Marine | Permeability reduction; may help preserve porosity through formation of grain-coating micro-quartz |

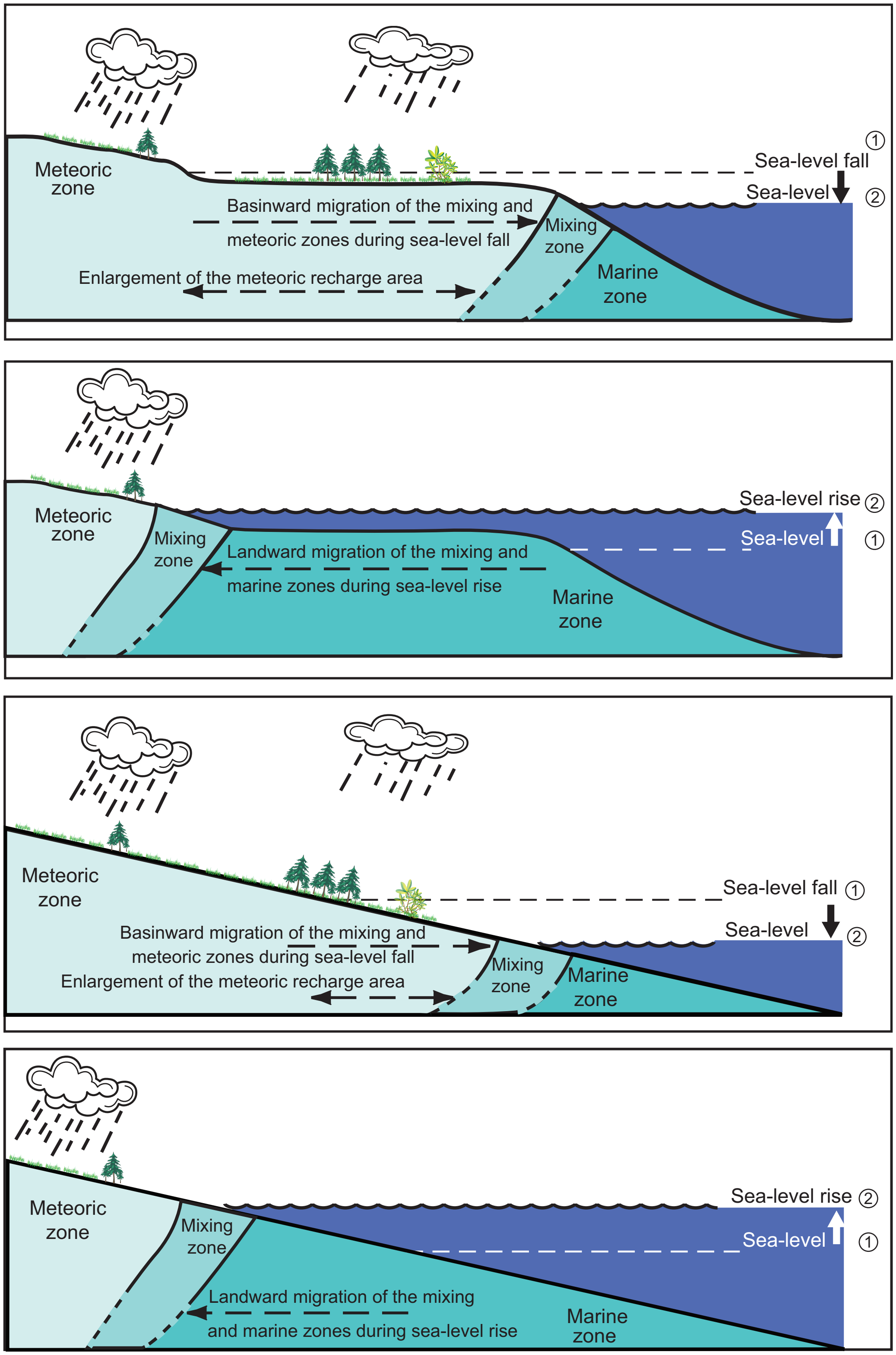


**Fig. 4.** Shift observed in the distribution of meteoric, mixing-zone and marine zones in platforms and ramps during sea-level fall and rise. A larger area is affected in platforms than in homoclinal ramps.

tens of metres would expose most shallow water shelves with break (for siliciclastic deposits) as well as platforms and rimmed shelves (for carbonate deposits) (Wilkinson, 1982; Read, 1985; Hanford & Loucks, 1993). Conversely, a similar fall in the relative sea-level would expose a much smaller area of homoclinal shelves (Fig. 4; Harris, 1986; Calvet *et al.*, 1990). A fall in the relative sea-level subsequent to transgression and early sea-level highstand is expected to be associated with a progressive change in pore water chemistry across the shelf from fully marine to mixed-marine-meteoric and, finally, fully meteoric composition.

### *Carbonate deposits*

Meteoric-water flux influences the exposed upper parts of ramp and, particularly, platform sediments, whereas the deeper parts may undergo marine pore water diagenesis. This depth-related variation in pore-water composition can be attributed to the 'floating' of meteoric waters over the denser marine pore waters (Hitchon & Friedman, 1969). Typical diagenetic alterations below the SB include (Table 1):

1. Karstification due to dissolution of TST and HST carbonate sediments by meteoric and brackish waters (Smart *et al.*, 1988; Moss & Tucker, 1992; Evans *et al.*, 1994; Jones & Hunter, 1994), which are undersaturated with respect to most marine carbonate sediments, particularly to high-Mg calcite and aragonite. Dissolution of aragonitic bioclasts and ooliths may lead to the formation of moldic and vuggy porosity and hence in improvement of reservoir quality (Tucker & Wright, 1992; Benito *et al.*, 2001; Fig. 5A ). Therefore, the original mineralogy of the carbonate sediments controls the intensity of creation of fabric-selective, secondary porosity. The low-Mg calcitic Jurassic-Cretaceous and mid-Palaeozoic oolites, as well as the Palaeozoic bioclasts and cool water limestones are expected to display smaller extent of meteoric water diagenesis (dissolution-cementation) than the aragonitic Mesozoic-Cenozoic bioclasts as well as the Permian-Triassic and Cenozoic oolites (Tucker, 1993).

   The dissolution of carbonate grains may lead to saturation of the meteoric fluids relative to low-Mg calcite, typically promoting precipitation of meteoric equant spar (Bourque *et al.*, 2001), which occludes primary intergranular and intragranular porosity (Figs. 5B and C). These molds and vugs may also be filled by coarse-crystalline, mesogenetic blocky calcite, dolomite and/or anhydrite (Choquette & James, 1987; Emery *et al.*, 1988; Moore, 2004) or by eogenetic, marine radiaxial and fascicular calcite or botryoidal aragonite cements during the following marine transgression (Kendall, 1977, 1985; Mazzulo & Cys, 1979; Csoma *et al.*, 2001). Karstification is intense under humid climatic conditions, which is due to the high rates of meteoric water recharge and extensive vegetation (Longman, 1980; James & Choquette, 1988, 1990; Wright, 1988). Vegetation acts as source of $CO_2$ and organic acids, which accelerate the dissolution of carbonates owing to acidification of meteoric waters. Meteoric-water diagenesis below SB results also in neomorphism of marine aragonite and high-Mg calcite cements and grains to low-Mg calcite (Fig. 5D; Longman, 1980; James & Choquette, 1990). Cementation of limestones below SB by phreatic blocky, equant, drusiform, syntaxial overgrowth and isopachous low-Mg calcite spar (Figs. 5B and C) (Carney *et al.*, 2001). Meteoric calcite cement contains very low but variable Mn and Fe owing to the overall oxic to weakly sub-oxic pore waters (Froelich *et al.*, 1979; Berner, 1981). Thus, meteoric-water calcite cement is non-luminescent or displays zones of dull and light brown/orange luminescence (Moss & Tucker, 1995), which are attributed to fluctuation in the redox potential in the pore waters (Edmunds & Walton, 1983).

   Despite the fact that transgression is accompanied by largely marine pore-water diagenesis, the concomitant rapid, yet local, expansion of ooid sands and barrier island formation is associated to meteoric diagenesis (Grammer *et al.*, 2001). Diagenesis of these carbonate sands commonly result in dissolution of metastable carbonate grains (aragonite and high-Mg calcite) and hence in eogenetic near-surface enhancement of reservoir quality. Local precipitation of poikilotopic, low-Mg calcite cement may occur, however, causing deterioration of reservoir quality (Moore, 1985; Scholle & Halley, 1985; Emery *et al.*, 1988; Moore, 2004).
2. Calcitization of dolomite (dedolomitization). Changes in pore water chemistry from marine and mixed marine/meteoric to meteoric composition are associated with shifting of the mineral stability field from dolomite to calcite that is commonly encountered below SB

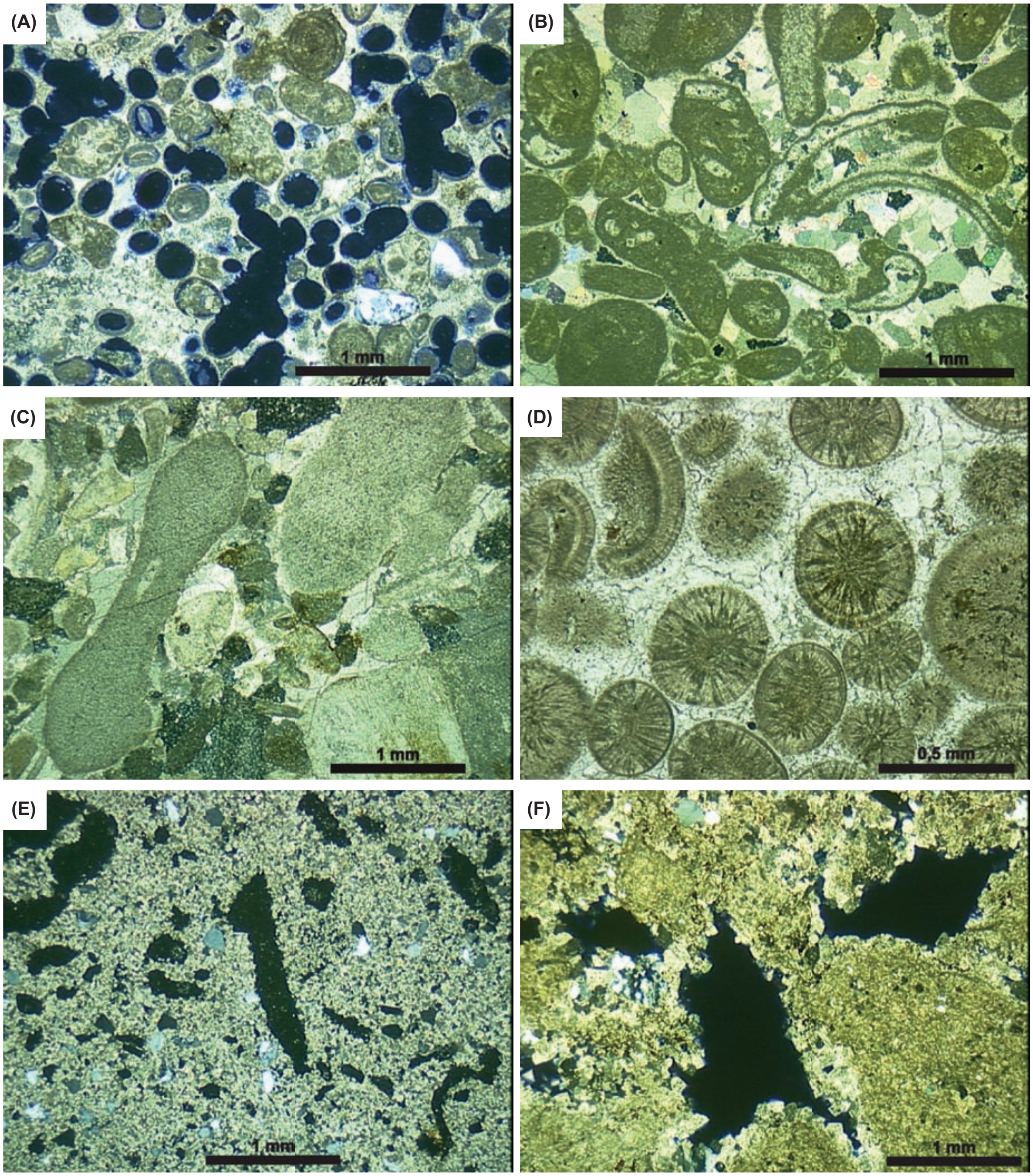


**Fig. 5.** Diagenetic processes related to depositional and stratigraphic setting in carbonate rocks. (A) Photomicrograph showing the development of vuggy pores by the coalescence of moldic pores from the dissolution of carbonate ooids by meteoric water, related to exposure. Albian, Sergipe-Alagoas Basin, NE Brazil. Crossed polarized light (XPL). (B) Intraclastic-bioclastic grainstone pervasively cemented by meteoric low-Mg calcite mosaic after fibrous rims. Albian, Potiguar Basin, NE Brazil. XP. (C) Pervasive syntaxial calcite overgrowths on crinoid bioclasts. Cambrian, South Australia. XPL. (D) Radial ooids (some of which have ostracodes nuclei) cemented by fibrous rims extensively recrystallized and microcrystalline mosaic. Permian, Paraná Basin, southern Brazil. Plane-polarized polarizers (PPL). (E) Moldic pores formed by dissolution of bioclasts in microcrystalline dolostone with sand and silt grains. Upper Cretaceous, Sergipe-Alagoas Basin, NE Brazil. XPL. (F) Dolomite crystals lining vuggy pores in partially dolomitized intraclastic rudstone. Albian, Jequitinhonha Basin, E Brazil. XPL.

(e.g. Fretwell *et al.*, 1997). Calcitization of dolomite may be associated by dissolution of Ca-sulphate cements and dolomite and thus result in improvement of reservoir quality (Sellwood *et al.*, 1987).

3. Pedogenesis under semi-arid climatic conditions, which may be accompanied by the formation calcrete (caliche) horizons with typical meniscus and pendular cement textures, laminated crusts and root casts/rhizocretions in the upper vadose zone (Harrison, 1978; Adams, 1980; Esteban & Klappa, 1983; Wilson, 1983; Wright, 1988, 1996; Tucker & Wright 1990; James & Choquette 1990; Charcosset *et al.*, 2000). Exposure surfaces may constitute impervious horizons forming fluid-flow barriers in carbonate reservoirs. In some cases, evidence of pedogenesis includes subtle changes in stable isotopes and trace element compositions of limestones, e.g. decrease in $\delta^{13}C$, $\delta^{18}O$ and Sr concentrations and increase in $^{87}Sr/^{86}Sr$ ratio (Cerling, 1984; Railsback *et al.*, 2003).
4. The formation of kaolinite and bauxite. Humid climatic conditions and extensive vegetation cover lead, in rare cases, to the formation of patches of kaolinite and, in rare cases, bauxite layers in clay-mineral rich carbonate successions (Bardossy & Combes, 1999; Csoma *et al.*, 2004). The low mobility of $Al^{3+}$ probably precludes its transportation in dissolved form with the percolating meteoric waters (Maliva *et al.*, 1999; Morad *et al.*, 2000).
5. Dolomitization. Dolomitization may occur due to fall in the relative sea-level, presumably through: (a) evaporation of marine pore water, particularly in near-shore environments (Zenger, 1972; M'Rabet, 1981; Machel & Mountjoy, 1986) and (b) in the mixed meteoric/marine (brackish) pore water zone that lies between the phreatic marine and phreatic meteoric pore water zone (Badiozamani, 1973; Humphrey, 1988). Dolomitization under these circumstances is commonly associated with the development of moldic or vuggy pores by selective or non-selective dissolution of aragonite or Mg-calcite constituents (Figs. 5E and F). According to the evaporative models, dolomitization is caused by an increase in the $Mg^{2+}/Ca^{2+}$ ratio, which is attributed to precipitation of gypsum and anhydrite (Adams & Rhodes, 1960; Hardie, 1987; Machel & Mountjoy, 1986; Morrow, 1990).

   The mixed marine/meteoric pore-water zone is shifted landwards during relative sea-level fall, which may account for an upwards increase in dolomitization in regressive carbonate successions (Taghavi *et al.*, 2006). These extensively dolomitized, extremely tight zones, which display high density log responses, may baffle vertical hydrocarbon flow (Taghavi *et al.*, 2006). However, the absence of considerable, if any, amounts of dolomite in modern mixed marine/meteoric zones castes doubts on the viability of mixing zone dolomitization model (Machel, 1986; Machel & Burton, 1994; Melim *et al.*, 2004). Instead, it is generally agreed that mixing zone diagenesis results in the dissolution of aragonite and high-Mg calcite and precipitation of bladed and overgrowth low-Mg calcite (e.g. Csoma *et al.*, 2004).

   Therefore, upward increase in extent of dolomitization in regressive sequences can probably be attributed to more restricted connection of shelf waters with open marine water, leading to evaporative precipitation of Ca-sulphates and concomitant increase in $Mg^{2+}/Ca^{2+}$ ratio in pore waters. A major fall in the relative sea-level and consequent subaerial exposure of tidal limestone deposits may thus induce dolomitization of HST and TST limestones below SB according to the supratidal-evaporative seepage reflux model, which requires warm, arid climatic conditions (Tucker, 1993).
6. A less common yet distinctive feature of exposed limestone includes darkened limestones and limestone intraclasts known as black pebbles, which occur in shallow subtidal, intertidal and supratidal environments (Strasser, 1984; Leinfelder, 1987; Shinn & Lidz, 1988). The blackening is attributed to the presence of organic matter (decayed cyanobacteria) (Strasser, 1984). Blackened limestones, which can be used to recognize SB, are commonly associated with gamma ray peaks (Evans & Hine, 1991).

### *Siliciclastic deposits*

Diagenetic processes affecting siliciclastic sediments below the SB on the continental shelf (typically the HST sediments), which are conducted by dominantly meteoric waters, include (Table 2):

1. Mechanical clay infiltration. Grain-coating clay minerals may be introduced into sandy deposits by the infiltration of muddy rivers waters into sandy deposits (Fig. 6; Ketzer *et al.*, 2003b). Clay infiltration (Fig. 7A) is more pervasive under

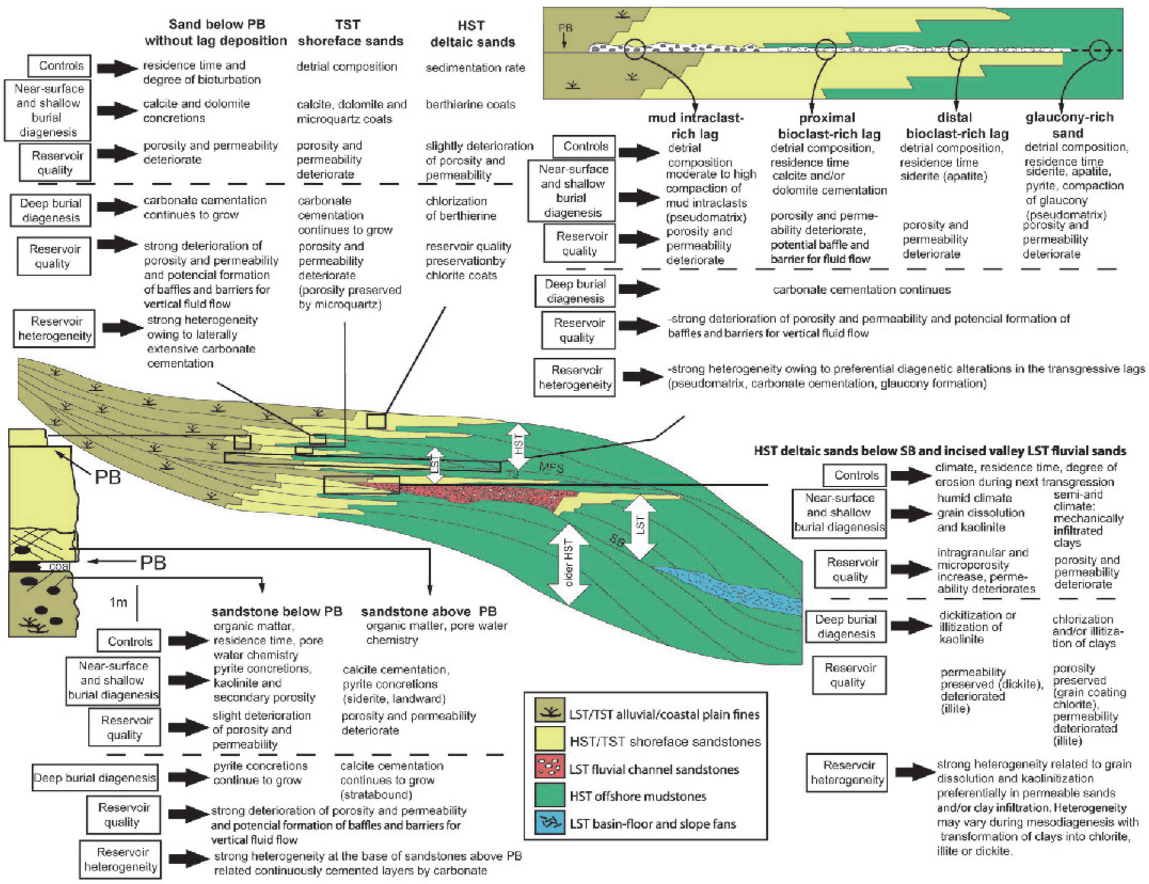


**Fig. 6.** Summary of diagenetic processes and patterns observed in fluvial, deltaic, coastal and shallow marine sandstones of key sequence stratigraphic surfaces and systems tracts (modified after Morad *et al.*, 2000; Ketzer *et al.*, 2003b).

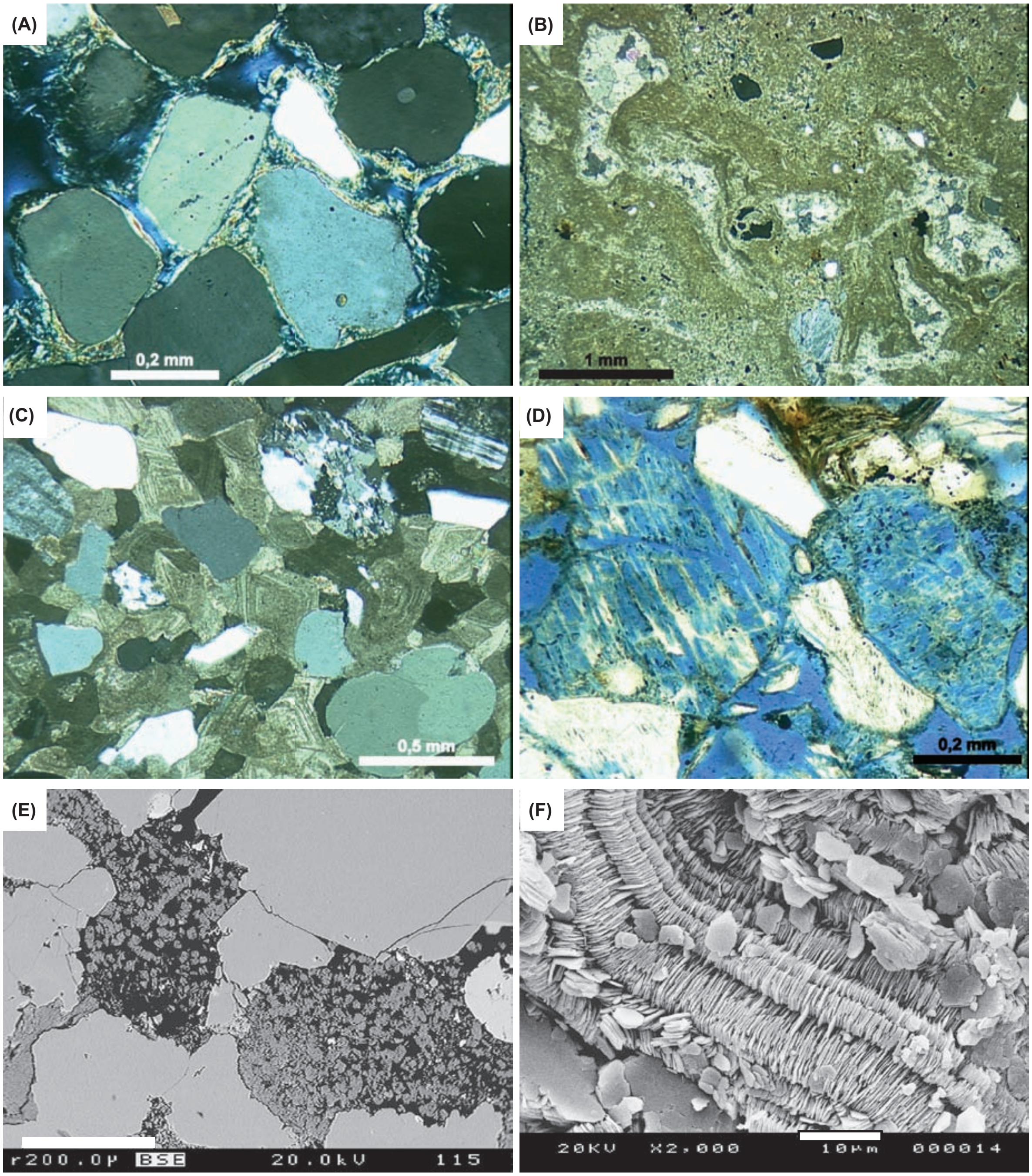


**Fig. 7.** (A) Irregular, anisopachous, discontinuous coatings of mechanically-infiltrated clays in Early Cretaceous fluvial sandstone, Recôncavo Basin, NE Brazil. Crossed polarizers (XPL). (B) Calcrete formed by multiple, displacive crusts of microcrystalline low-Mg calcite. Displaced, 'floating' sand grains. Albian, Espírito Santo Basin. E Brazil. XPL. (C) Phreatic dolocrete constituted by coarsely crystalline, displacive dolomite with strong zoning defined by fluid inclusions and 'floating' sand grains. Jurassic, Recôncavo Basin, NE Brazil. XPL. (D) Strongly dissolved feldspar grains. Late Cretaceous, Espírito Santo Basin, E Brazil. XPL. (E) Feldspar grains replaced by vermicular kaolinite. Backscattered electrons (BSE) image. Cretaceous. Sirte Basin, Libya. (F) Vermicular kaolinite aggregate made of stacked platelets with aligned defective edges, characteristic of low-temperature precipitation. Secondary scanning electron microscope (SEM) image. Late Cretaceous, Utah, USA.

semi-arid climate, owing to the deeper position of the phreatic level that allows muddy waters to infiltrate through a thick vadose zone (Moraes & De Ros, 1990). The preservation potential of sandstones containing mechanically infiltrated clays below SB is relatively low because of marine erosion of such sandstones during the next transgression event and formation of the transgressive surface (Molenaar, 1986; Ketzer *et al.*, 2003b; Fig. 6).

The formation of grain-coating, infiltrated clays may have a profound impact on the mesogenetic and related reservoir-quality evolution pathways (Moraes & De Ros, 1990; Jiao & Surdam, 1994; De Ros & Scherer, this volume). As product of dry climate weathering, infiltrated clays are originally smectitic in composition (De Ros *et al.*, 1994; Worden & Morad, 2003), being transformed into illite or chlorite during burial. Grain-coating illite in sandstones may cause either: (i) deterioration of reservoir permeability due to the fibrous and filamentous crystal habits of illite crystals and their distribution as rims blocking pore throats (Glassman *et al.*, 1989; Burley & MacQuaker, 1992; Ehrenberg & Boassen, 1993), (ii) deterioration of reservoir quality through enhancement of pressure dissolution (i.e. chemical compaction; Tada & Siever, 1989; Thomson & Stancliffe 1990), or (ii) enhancement of reservoir quality through the retardation or inhibition of precipitation of syntaxial quartz overgrowths (Morad *et al.*, 2000; Worden & Morad, 2003; Al-Ramadan *et al.*, this volume; De Ros & Scherer, this volume).

Whether illite or chlorite from eogenetic smectites is conditioned by: (i) the original composition of the smectite; illite is preferably derived from dioctahedral smectite, whereas chlorite is derived from trioctahedral smectite (Chang *et al.*, 1986). (ii) Derivation of $K^+$ from the dissolution and albitization of detrital K-feldspars, which encourages the formation of illite (Fig. 6; Morad, 1986; Aagaard *et al.*, 1990). (iii) Derivation of $Fe^{2+}$ and $Mg^{2+}$ from the dissolution or replacement of abundant ferromagnesian grains (e.g. biotite) and volcanic rock fragments favours the formation of chlorite (Morad, 1990). (iv) Derivation of fluids from associated mudrocks and evaporites may form illite or chlorite (Boles, 1981; Gaup *et al.*, 1993; Gluyas & Leonard, 1995). In cases where the presence of grain-coating chlorite in LST incised valley sandstones cannot be related to mechanical clay infiltration, formation by chemical precipitation from pore waters is probable (Salem *et al.*, 2005; Luo *et al.*, 2009).

2. Formation of calcretes and dolocretes. Subaerial cementation of siliciclastic sediments by calcite (calcrete) and dolomite (dolocrete) may occur in the vadose and phreatic zones below SB (Figs. 7B and C). Dolocretes are most common under arid climatic conditions, whereas calcretes occur under semi-arid climatic conditions (Watts, 1980; Khalaf, 1990; Spötl & Wright, 1992; Burns & Matter, 1995; Colson & Cojan, 1996; Williams & Krause, 1998; Morad *et al.*, 1998). Calcretes and dolocretes developed in the vadose zone commonly display rhizocretions and crusts formed around and plant roots (Fig. 7B; Semeniuk & Meagher, 1981; Purvis & Wright, 1991; Morad *et al.*, 1998). Calcretes and dolocretes occur as scattered concretions or as aerially extensive cement, which may act as fluid flow baffles (Khalaf, 1990; Beckner & Mozley, 1998; Morad, 1998; Morad *et al.*, 1998; Williams & Krause, 1998; Worden & Matray, 1998; Schmid *et al.*, 2004).
3. Grain dissolution and kaolinization. Meteoric waters are undersaturated with respect to most framework silicate grains. Therefore, percolation of these waters below the SB typically results in the dissolution (i.e. formation of intragranular and moldic porosity) and kaolinization of unstable framework silicates (e.g. micas and feldspars) (Figs. 6 and 7D–F), most extensively under humid climatic conditions (Worden & Morad, 2003; Ketzer *et al.*, 2003a). Dissolution and kaolinization of mica is commonly accompanied by the formation of siderite (Fig. 8A; Morad, 1990). Siderite, which induces expansion to the mica flakes, forms under sub-oxic to anoxic pore-water conditions by fermentation of organic matter and may also occur as concretions and scattered cement patches with microcrystalline and spherulitic habits (Hutcheon *et al.*, 1985; Mozley & Hoernle, 1990; Baker *et al.*, 1995; Morad *et al.*, 1998; Huggett *et al.*, 2000). The formation of siderite within expanded mica causes local occlusion of pore throats and, hence, reduction in reservoir permeability.
4. Reworking of autochthonous glaucony. A fall in the relative sea-level below shelf break and consequent valley incision may also result in

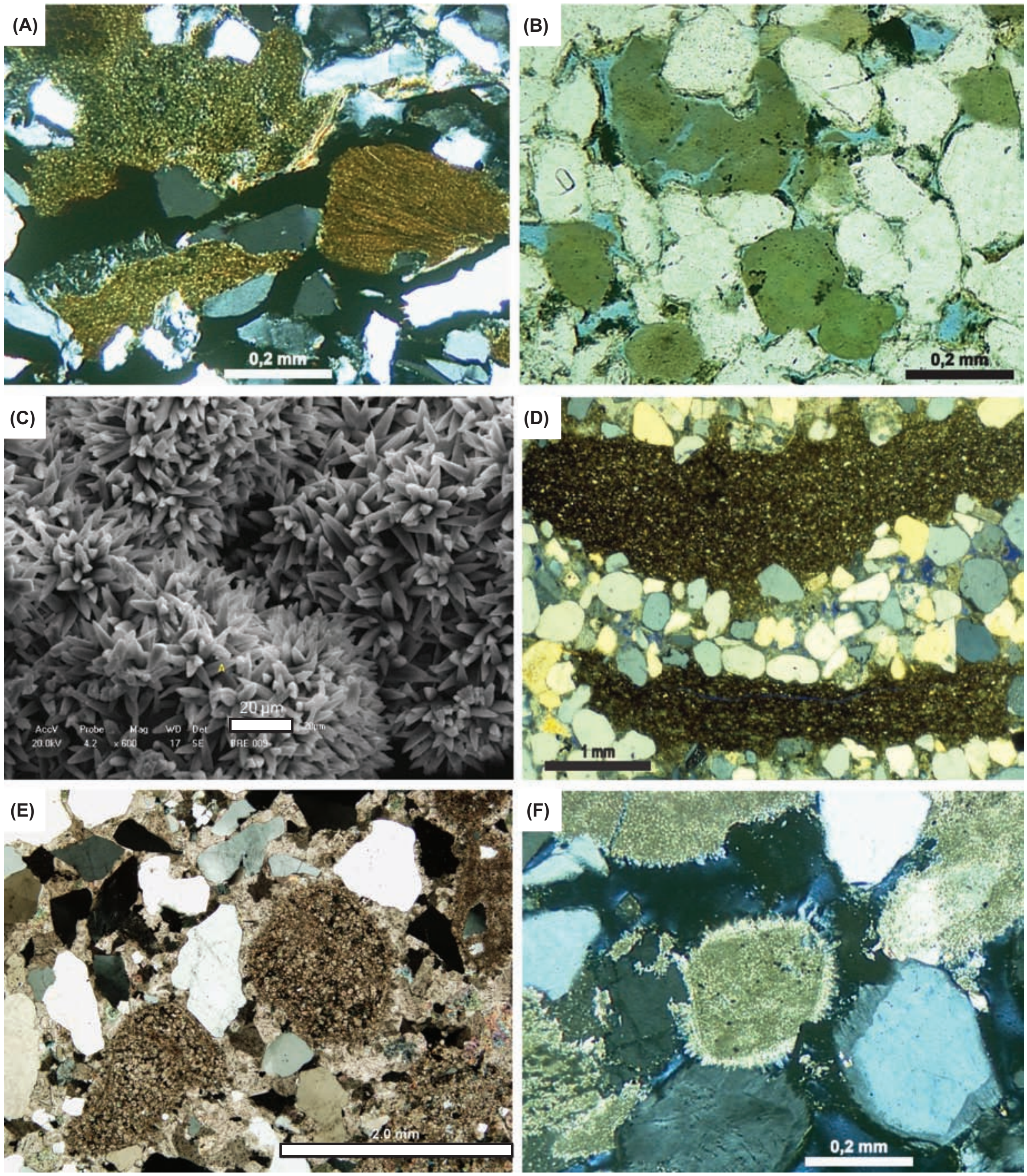


**Fig. 8.** (A) Biotite flakes widely expanded and replaced by microcrystalline siderite (brown). Carbonaceous fragments (black). Late Cretaceous, Espírito Santo Basin, E Brazil. XPL. (B) Parautochtonous glauconite in shallow-water Cretaceous sandstone from Oriente Basin, Equador. PPL. (C) Divergent aggregates of scalenohedral, 'dogtooth' high Mg-calcite crystals rimming the grains in Holocene beachrock, NE Brazil. (D) Mud intraclasts partially compacted to pseudomatrix. Jurassic, Recôncavo Basin, NE Brazil. XPL. (E) Dolomitized carbonate intraclasts in sandstone cemented by blocky dolomite. XPL. Cretaceous, Sirte Basin, Libya. (F) Hybrid arenite with carbonate intraclasts and bioclasts rimmed by originally high-Mg calcite. Potassic feldspar grains with distinct epitaxial overgrowths. Cenomanian, Potiguar Basin, NE Brazil.

the erosion of autochthonous glaucony-rich, TST and early HST sediments (Baum & Vail, 1988; Glenn & Arthur, 1990; Ketzer *et al.*, 2003). Parautochthonous glaucony may be re-deposited in paralic and shallow-marine settings (Fig. 8B), as well as the slope fan and deep-sea fan sand deposits (Amorosi, 1995). Thus, abundant locally reworked glaucony in paralic and deep marine sand deposits can be used as a criterion for the recognition of the SB. This is particularly important in marine turbidites, in which the recognition of systems tracts and key sequence stratigraphic surfaces is problematic (Amorosi, 1995, 1997).

### *Parasequence boundaries (PB), transgressive surfaces (TS), maximum flooding surfaces (MFS)*

These key sequence stratigraphic surfaces, which are the product of faster rates of rise in the relative sea-level than the rates of sediment supply (i.e. transgression or retrogradation), lead to domination of marine pore waters.

### *Carbonate deposits*

The impact of changes in the relative sea-level and shelf physiography on the distribution of near-surface, eogenetic alterations in carbonate deposits is depicted in Fig. 9 and summarized in Table 1.

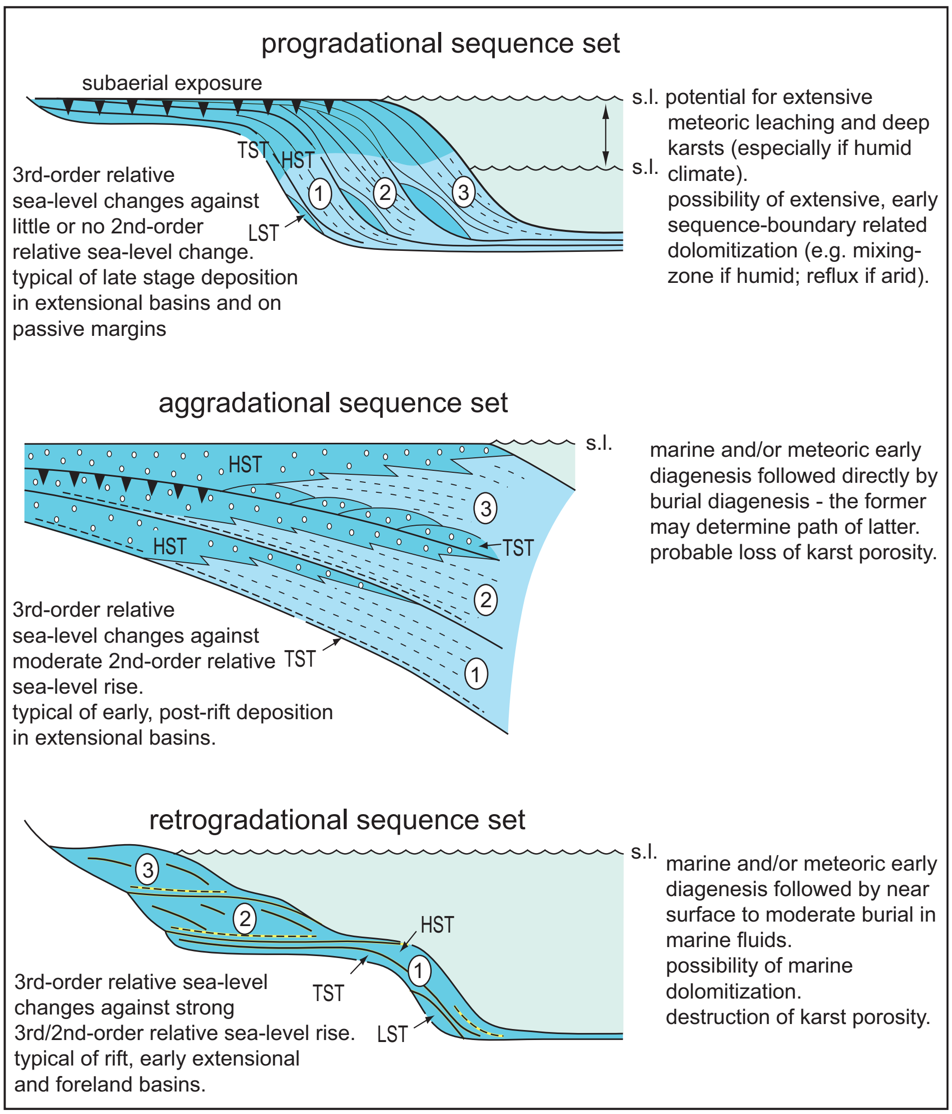


**Fig. 9.** Diagenetic models of carbonate shelves and ramps in response to 3rd order sequence stacking patterns in a background of 2nd order sequences. Progradational and retrogradational patterns are more typical of carbonate shelves, while aggradational sets are shown for a carbonate ramp. Modified from Tucker (1993).

Transgressive surfaces in marine carbonate successions can be recognized by distinct pattern of diagenetic alterations, increase in gamma ray responses (owing to increase in clay minerals) and/or increase in extent of bioturbation (Tucker & Chalcraft, 1991). Prior to the establishment of fully marine pore-water composition as a consequence of rise in relative sea-level, migration of the marine/meteoric mixing and meteoric zones landward may result in cementation by marine calcite or by alternating marine calcite and mixing zone dolomite (Folk & Siedlecka, 1974; Hardie, 1987; Humphrey, 1988; Morad *et al.*, 1992; Frank & Lohmann, 1995). However, establishment of fully marine pore waters may preclude the occurrence of these latter diagenetic alterations across a shelf. Thus, alterations mediated by marine pore waters include cementation by high-Mg-calcite or aragonite and dolomitization of limestone (Tucker, 1993).

Sediment diagenesis in the subtidal zone and basinward is presumably mediated dominantly by diffusive rather than advective flux of $Ca^{2+}$, $Mg^{2+}$ and $HCO_3^-$ from the overlying sea water. Increasing number of field, stable O-isotopic, C-isotopic and Sr-isotopic data and thermodynamic equilibrium studies (Machel & Mountjoy, 1986; Machel & Burton, 1994; Whitaker *et al.*, 1994; Budd, 1997; Swart & Melim, 2000; Ehrenberg *et al.*, 2006b) suggests that dolomitization occurs by normal or slightly modified sea water. Apart from tidal pumping, there is little evidence to suggest that circulation of sea water occurs in sediments buried at shallow depths below the seafloor. Dolomitization by advective sea water flux requires long lasting circulation of large volumes through the sediments (Machel & Mountjoy, 1986; Hardie, 1987; Budd, 1997). Circulation of sea water in the subsurface of carbonate platforms is suggested to be driven by a combination of salinity and thermal gradients (Whitaker *et al.*, 1994; Kaufman, 1994; Ehrenberg *et al.*, 2006b).

Diffusive ionic flux from sea water into pore waters may result in the development of hardground and firmground by extensive cementation of carbonate sediments below TS, PB and MFS by calcite and/or dolomite (± phosphate, glaucony, Fe-oxide). Cementation commonly extends for few decimetres below the seafloor (Folk & Lynch, 2001; Mutti & Bernoulli, 2003). The development of hardgrounds and firmgrounds may baffle fluid flow and hence causes reservoir compartmentalization in carbonate successions (Mancini *et al.*, 2004).

Coal layers may be deposited on carbonate shelves primarily along the transgressive surfaces and early stages of TST deposition in humid climatic conditions (de Wet *et al.*, 1997; Longyi *et al.*, 2003; Shao *et al.*, 2003). Organic acids generated by coals may promote extensive dissolution of carbonate horizons below transgressive surfaces. The formation of Mn-oxyhydroxide and Fe-oxyhydroxide nodules in the abyssal plains of modern oceans, which is favoured by low sedimentation rates (i.e. similar conditions to condensed sections), suggests that the occurrence of such oxyhydroxides in the stratigraphic record of shelf deposits may be used as analogs to recognize MFS (cf. McConachie & Dunster, 1996).

Although reddish colouration of carbonate sediments is typically attributed to oxidation of iron during subaerial exposure, it has been argued by several authors (Jenkyns, 1986; Van Der Kooij *et al.*, 2007) that staining in sediment along the MFS in platform top, slope and the basin floor implies fully marine conditions. Staining by marine pore waters was further evidenced by elevated $\delta^{18}O$ values (+2 to +3‰) of the carbonate cement (van der Kooij *et al.*, 2007). Reddening has been attributed by these authors to iron oxidation during early diagenesis by iron bacteria, which occurred upon upwelling of cold, nutrient-rich water masses.

### *Siliciclastic deposits*

Diagenetic alterations related to PB, TS, MFS and TST in siliciclastic successions include (Table 2): (i) formation of concretionary or continuous marine calcite, dolomite and siderite cementation of sandstone and mudstone beds; (ii) carbonate cementation or formation of pseudomatrix in transgressive lag deposits, (iii) calcite, pyrite and kaolinite cementation in sandstones below and above coal-bearing PB and (iv) formation of autochthonous glaucony (Whalen *et al.*, 2000; Amorosi, this volume). The formation of carbonate cements along PB, TS and MFS (De Ros *et al.*, 1997; Ketzer *et al.*, 2002; Coffey, 2005) is probably related to the increase in extent of bioturbation (Hendry *et al.*, 2000); and in amounts of marine organic matter content, which helps increasing the carbonate alkalinity and decrease Eh of the pore waters (Curtis, 1987; Morad, 1998; Al-Ramadan *et al.*, 2005).

Coalesced concretionary or continuous carbonate cementation (± phosphate, Fe-oxides, Fe-silicates) are favoured below the MFS in dominantly sandstones or mudstone successions

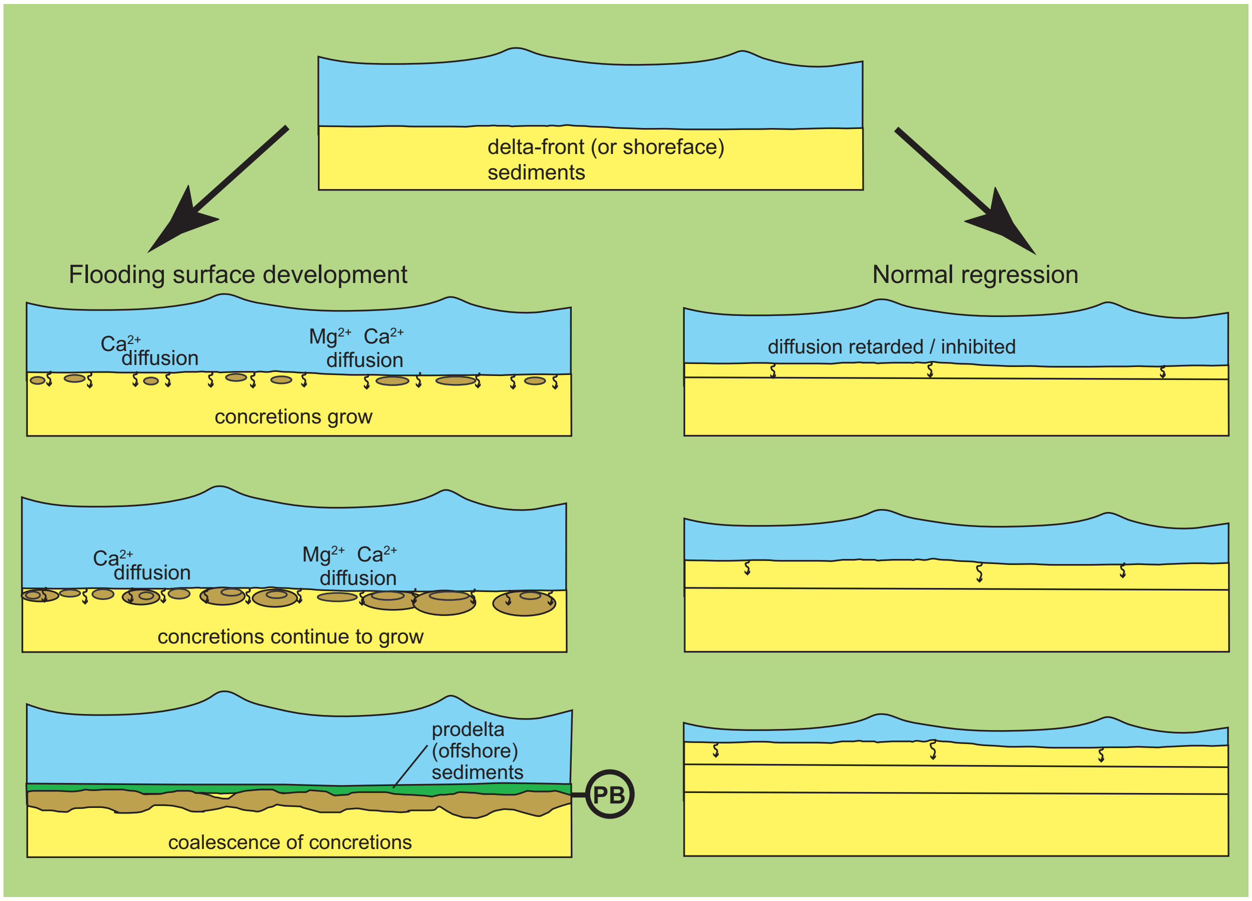


**Fig. 10.** Schematic representation of the development of carbonate cementation in clastic sediments below flooding surfaces, as opposed to absence of cementation under normal regressive conditions. Extensive cementation below marine transgressive surfaces act as baffles for fluid flow and may thus result in reservoir compartmentalization.

(Morad *et al.*, 2000; Wetzel & Allia, 2000; Al-Ramadan *et al.*, 2005). Cementation (most commonly calcite) is suggested to occur at very shallow depth below the seafloor being facilitated by reduced sedimentation rates (long residence time below the seafloor), which allows prolonged diffusion of $Ca^{2+}$ and $HCO_3^-$ into pore waters from overlying sea water (Figs. 10 and 11) (Kantorowicz *et al.*, 1987; Raiswell, 1988; Savrda & Bottjer, 1988; Morad & Eshete, 1990; Wilkinson, 1991). Once nucleation of calcite cement occurs within the sediment (e.g. around bioclasts and/or in locally concentrated marine organic matter), chemical gradients of $Ca^{2+}$ and $HCO_3^-$ are established between the sites of carbonate precipitation from pore water (concentration is nil; Berner, 1982) with overlying sea water (contains high amounts of dissolved calcium and carbon) column (Fig. 11; Morad & De Ros, 1994). Petrographic and oxygen isotopic signature suggest that concretion growth may commence below the sediment-water interface but continues during burial diagenesis (Klein *et al.*, 1999; Raiswell & Fisher, 2000; Al-Ramadan *et al.*, this volume).

The presence of concretionary or continuous stratabound cementation within mudstone sections (referred to as hiatus limestones by Wetzel & Allia, 2000) is important in two respects: (i) it aids the recognition of major transgressive surfaces within thick, monotonous siliciclastic mudstone successions; (ii) Act as baffle fluid flow, including primary migration of hydrocarbon within source rocks. Calcite and, less commonly, dolomite cements in diagenetic concretions and beds within mudstones have micritic and radial habits and occur between the clay mineral flakes

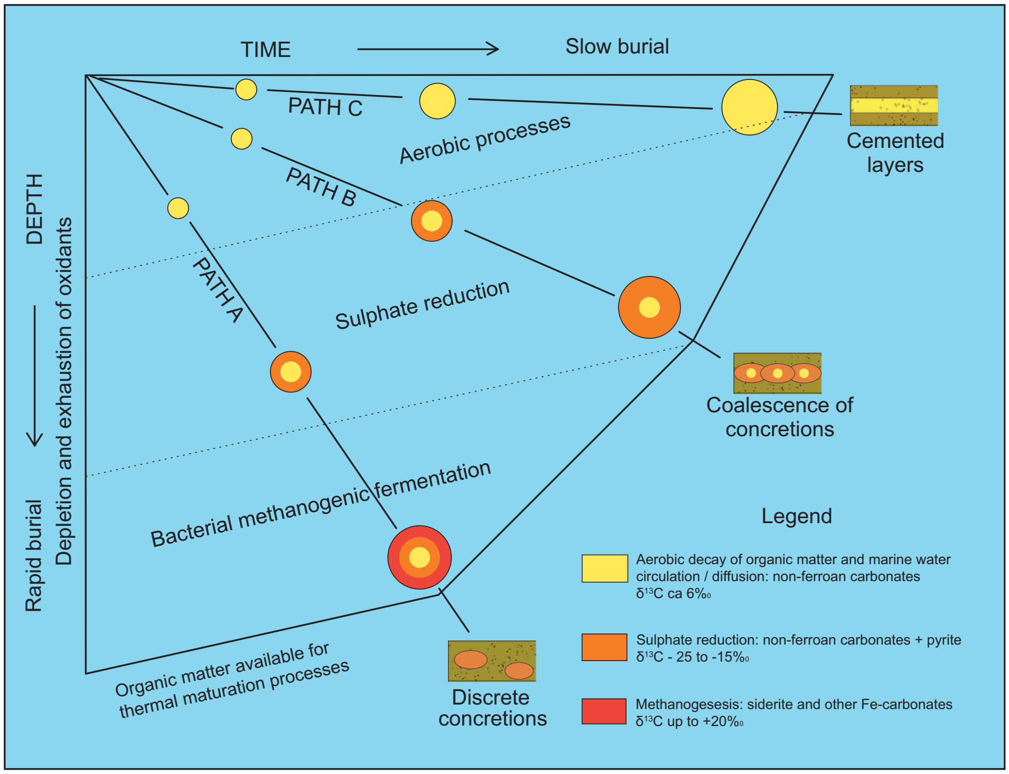


**Fig. 11.** Schematic representation of the impact of sedimentation rate on the styles of carbonate cementation in marine sandstones. Sediments which experience long residence time at shallow depth below sea bottom remain within the aerobic zone and may be cemented by isotopically homogeneous, laterally continuous, stratabound calcite. Under larger sediment supply rates, the carbon and oxygen compositions of carbonate cements tends to be concentrically arranged, reflecting the diverse zones of bacterial organic matter degradation. Modified after Kantorowicz *et al.* (1987).

and, in some cases, silt-sized quartz and feldspar (Morad & Eshete, 1990; Wetzell & Allia, 2000; Al-Ramadan *et al.*, 2005). These calcite cements have variable and overall low $\delta^{13}C_{V\text{-}PDB}$ (−40‰ to −2‰) and $\delta^{18}O_{V\text{-}PDB}$ (−12‰ to −4‰) compositions. The carbon isotopic signatures indicate derivation of carbon from various sources, ranging from sea water to microbial alteration of organic matter (e.g. methanogenesis and sulphate reduction; Fig. 10; Kantorowicz *et al.*, 1987; Morad & Eshete, 1990; Coleman & Raiswell, 1993; Wetzel & Allia, 2000). The lower oxygen isotopic signatures of the calcite than expected for inferred precipitation from marine pore waters was attributed to recrystallization and/or additional cementation during burial diagenesis (Morad & Eshete, 1990; Mozley & Burns, 1993; Raiswell & Fisher, 2000).

Like the presence of extensive carbonate cements within mudstone successions, the occurrence of considerable amounts of diagenetic phosphates and Fe-minerals (siderite, glauconite and berthierine ± pyrite) can also be used to recognize MFS and TS in mudstone (MacQuaker & Taylor, 1996).

The TS in siliciclastic successions is commonly marked by the presence of heavily carbonate-cemented lag deposits formed by carbonate bioclasts as well as carbonate and/or mud intraclasts reworked by waves from earlier fine-grained sediments (Posamentier & Allen, 1999). In rare cases, such lag deposits are rich in mud intraclasts, which are derived from marine erosion of shelf, lagoonal, deltaic or even fluvial deposits (Fig. 8D); the same lag layer may be rich in marine bioclasts in basinward direction (Fig. 6). The composition of such lags, which is thus controlled by the type of reworked sediments and their degree of lithification, has a substantial impact on the eogenetic and related reservoir-quality evolution pathways. The mechanical compaction of mud intraclasts results in the formation of abundant pseudomatrix and hence deterioration of reservoir quality (Fig. 6). Lags rich in carbonate bioclasts or intraclasts are pervasively cemented by calcite, dolomite and siderite (Fig. 6) because carbonate clasts act as nuclei or also as source for these cements (Figs. 8E and F; Ketzer *et al.*, 2002; De Ros & Scherer, this volume). Siderite, in particular, is formed in more distal sediments compared to calcite and dolomite cemented lags (Fig. 6), possibly because of the prolonged suboxic diagenetic conditions (Ketzer *et al.*, 2003a). Therefore, the formation of baffles for fluid flow and reservoir compartmentalization may occur if amalgamated sandstone bodies are separated by pseudomatrix-rich or heavily carbonate-cemented transgressive lags (Ketzer *et al.*, 2002, 2005).

The PB, TS and MFS in clastic successions are eventually marked by the presence of marine deposits that may occur on top of coal layers (Van Wagoner *et al.*, 1990). Eodiagenesis and mesodiagenesis of organic matter in these coal deposits may result in the formation of pyrite concretions, extensive calcite cementation and kaolinization of framework silicates in adjacent sandstone beds (Ketzer *et al.*, 2003a; Fig. 6). Pyrite concretions form in sandstones above and below the peat/coal deposits, presumably owing to the bacterial reduction of sulphate-charged sea water supplied during transgression, which is promoted by abundant organic matter in the peat/coal layers (Curtis, 1986; Petersen *et al.*, 1998; Ketzer *et al.*, 2003a).

Heavily calcite-cemented sandstones occur on top of coal deposits and disappear in both the landward and basinward terminations of the coal deposits (Fig. 6; Ketzer *et al.*, 2003a). The formation of this carbonate cement, which is attributed to bacterial alteration of organic matter and consequent increase in the carbonate alkalinity of pore waters (Curtis, 1987), can also act as baffles for fluid flow and reservoir compartmentalization (Ketzer *et al.*, 2003a).

The formation of kaolinite in sandstone beds underlying the coal (Ketzer *et al.*, 2003a) has been attributed to percolation of acidic waters originating from generation of $CO_2$ and organic acids produced during microbial decay of organic matter in the coal/peat layer. These acidic meteoric waters cause the dissolution of silicate grains (e.g. feldspars and micas) and the formation of kaolinite in sandstones beneath the coal layers (Fig. 6; Taylor *et al.*, 2000). In addition to the formation of kaolinite and pyrite, diagenetic Fe-silicates (berthierine/chlorite) and ferroan carbonates are also closely associated with coal layers (Iijima & Matsumoto, 1982; Dai & Chou, 2007). The formation of these Fe-silicates is presumably facilitated by the overall reducing conditions, which result in the availability of $Fe^{2+}$ in the pore waters (Curtis, 1987).

In contrast to kaolinite and berthierine, glaucony is typically encountered along the outer shelf extension of PB, TS and MFS. Glaucony is concentrated along these surfaces by wave or tidal reworking (parautochthonous glaucony; cf. Amorosi, 1995; Ketzer *et al.*, 2003b) or be formed *in situ* (autochthonous glaucony). The formation of autochthonous glaucony is favoured by: (i) low sedimentation rates owing to low siliciclastic input to the distal shelf, i.e. long residence time of the sediments at very shallow depths below the seafloor and (ii) moderate amounts of organic matter causing the establishment of mildly reducing conditions, in which $Fe^{+2}$ and $Fe^{+3}$ can coexist (nitrate- and manganese-reducing, suboxic conditions; Berner, 1981; Curtis, 1987) for a prolonged time (Amorosi, 1995, 1997). The occurrence of glaucony at TS and MFS makes these surfaces fairly reliable stratigraphic markers, such as in the Cretaceous to Oligocene glaucony-rich successions of northern-central Europe (Robaszynski *et al.*, 1998; Vandenberghe *et al.*, 1998).

## DISTRIBUTION OF DIAGENETIC ALTERATIONS WITHIN SYSTEMS TRACTS

The distribution of diagenetic alterations within systems tracts is in principle similar to those encountered below the major sequence stratigraphic surfaces (Tables 1 and 2). Some of the diagenetic alterations may display trends of increase or decrease towards these surfaces (Morad *et al.*, 2000). However, as pointed out earlier, there are some genetic differences between the development of systems tracts in siliciclastic and carbonate depositional systems. Diagenetic alterations within the LST and upper parts of the HST of siliciclastic successions are similar to those encountered below SB. LST deposits are poorly developed or lacking in carbonate systems, whereas the HST deposits are extensive and undergo extensive marine pore-water diagenesis. Transgression subsequent to sea-level fall/lowstand brings about changes in pore water chemistry across shelves/platforms from meteoric to mixed and, ultimately, fully marine. Hence, the TST in both siliciclastic and carbonate systems undergo marine pore-water diagenesis under progressive increase in the residence time at and immediately below the seafloor towards the MFS.

*Carbonate systems*

The early diagenetic processes and products in carbonate successions show different patterns characteristic of the various system tracts (Table 1). The HST is marked by a substantial expansion of the shelf areas with active generation of carbonate sediments, which tend to be cemented by marine aragonite and/or Mg-calcite rims and pore-filling cements. The increase in the production of shallow-water carbonate sediments is reflected also in an increased contribution of intrabasinal carbonate grains to deep-water fan deposits (Fontana *et al.*, 1989). This corresponds to extensive calcite cementation of such resedimented carbonate (allodapic; Dolan, 1989) or hybrid turbiditic deposits during burial, owing mostly to release of $Ca^{2+}$ and $HCO_3^-$ from the pressure dissolution of the carbonate bioclasts and other allochems (Mansurberg *et al.*, 2009). Subaerially exposed HST deposits on a shelf display progressive marine grain and cement dissolution and development of karstic features towards the SB owing to meteoric water percolation (Evans *et al.*, 1994; Jones & Hunter, 1994). TST carbonate deposits may display upward increase in the amounts of marine carbonate cements (such as aragonite/high-Mg calcite rims and syntaxial overgrowths) and dolomitization along the TS and towards the MFS. Dolomitization, in particular, is expected to occur in TST and older HST sediments aided by the basinward movement of the marine pore water zone and active circulation of sea water and mixing of meteoric and sea waters in sediments (Tucker, 1993).

*Clastic systems*

Diagenetic alterations in siliciclastic successions may display systematic distribution within the various systems tracts (Morad *et al.*, 2000; Table 2). Late LST deposits, particularly fluvial, incised valleys sandstones display an increase in dissolution and kaolinization of framework silicate grains owing to meteoric water circulation towards the SB (Fig. 6; Ketzer *et al.*, 2003b). Therefore, LST sandstones are expected to be characterized by enhanced reservoir quality (Morad *et al.*, 2000). Conversely, under semi-arid climate, percolation of meteoric water in LST fluvial sandstones is limited and kaolinite is scarce or absent (Ketzer *et al.*, 2003b). The clay mineral is, instead, grain-rimming and grain-replacive smectite (Fig. 12A ), which eventually evolve to chlorite and/or illite during burial diagenesis (Moraes & De Ros, 1990; Humphreys *et al.*, 1994; Ketzer *et al.*, 2003b). The impact of grain-coating smectite on the diagenetic and related reservoir-quality evolution of sandstones has been discussed earlier in this paper.

Mechanically infiltrated clays are commonly abundant in braided fluvial systems of semi-arid settings owing to frequent avulsion of the channels, which allows muddy fluvial waters to infiltrate through the vadose zone in areas with lowered water table (Fig. 7A; Moraes & De Ros, 1990; De Ros & Scherer, this volume).

Horizons of infiltrated clays concentration are formed in braided fluvial sandstones (late LST), along the positions of the phreatic level at the infiltration events (Walker *et al.*, 1978; Moraes & De Ros, 1990; De Ros & Scherer, this volume). Recurrent clay infiltration may result in complete occlusion of the intergranular pores, resulting in early diagenetic destruction of reservoir quality of braided fluvial sandstones (Moraes & De Ros, 1990) and formation of flow barriers in fluvial reservoirs (De Ros & Scherer, this volume).

Other sites for the concentration of mechanically infiltrated clays include the proximal alluvial conglomerates, below recurrently flooded ephemeral channels or above impermeable barriers such as palaeosols, shallow basement (Walker *et al.*, 1978; Moraes & De Ros, 1990). Mud intraclasts, which can cause deterioration of reservoir quality upon mechanical compaction and formation of pseudomatrix, are a common product of erosion of HST deposits and incorporation in LST meandering and braided fluvial (Fig. 8D), deltaic, shallow marine and deep marine facies. In turbiditic sequences, mud intraclasts eroded from slope deposits are concentrated in coarse, channel complex deposits (Carvalho *et al.*, 1995: Bruhn & Walker, 1997; Mansurbeg *et al.*, 2009) and/or during periods of intense tectonic activity, when rejuvenation of source terrains topography and accentuation of margin angle causes the turbidity currents to cut new canyons and channels in the slope (Fetter *et al.*, 2009). Eogenetic and detrital smectites are transformed during burial into illite or chlorite (Fig. 12D), through mixed-layer illite-smectite and chlorite-smectite, respectively (Nadeau *et al.*, 1985; Chang *et al.*, 1986; Humphreys *et al.*, 1994; Niu *et al.*, 2000; Anjos *et al.*, 2003).

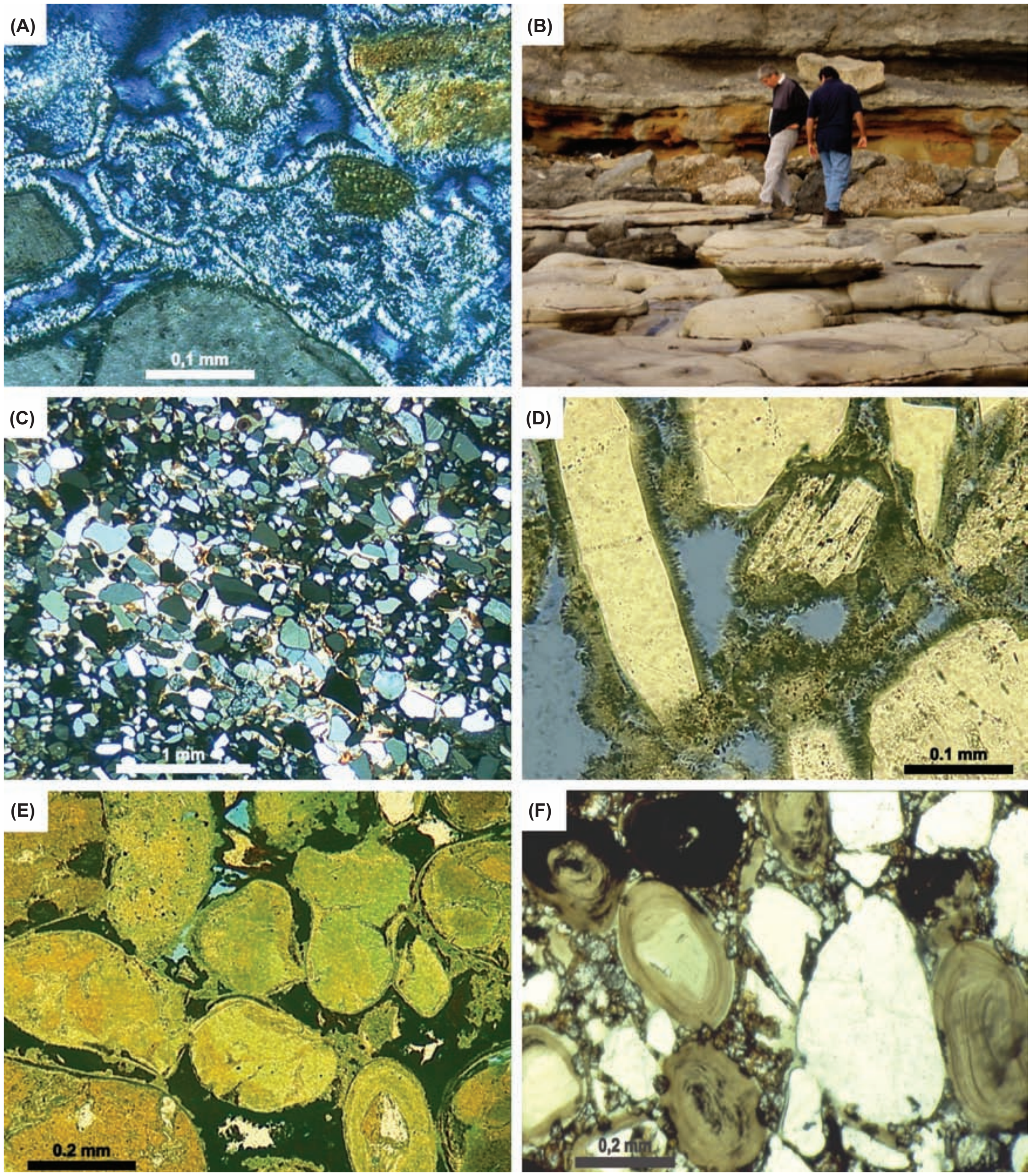


**Fig. 12.** (A) Smectite rims surrounding and replacing grains. Aptian, Espírito Santo Basin. XPL. (B) Disk-shaped concretions coalescing along sequence boundary. Jurassic, France. (C) Poikilotopic calcite selectively cementing the coarser-grained lamina in sandstone. Jurassic, Recôncavo Basin. XPL. (D) Chlorite rims preserving intergranular porosity in deeply buried sandstone. Upper Cretaceous, Santos Basin, E Brazil. Uncrossed polarizers (PPL). (E) Autochtonous glauconite peloids and ooids. Cretaceous, New Jersey, USA. PPL. (F) Chamosite ooids after berthierine in hybrid sandstone. Intergranular and grain-replacive siderite. Devonian, Paraná Basin, Brazil. PPL.

The TST and early HST paralic and shallow-marine sandstones have higher potential to be cemented by carbonates (notably calcite) and small amounts of pyrite than late HST and LST deposits (Fig. 6; South & Talbot, 2000; Morad *et al.*, 2000; Ketzer *et al.*, 2002). This is because marine transgression causes trapping of coarse-grained sediments in estuaries, reducing the sediment flux to the shelf (Emery & Myers, 1996), which implies prolonged residence time on the seafloor and enhanced diffusion of dissolved $Ca^{2+}$ and $HCO_3^-$ from sea water (Kantorowicz *et al.*, 1987; Wilkinson, 1991; Morad, 1998. Limited clastic input promotes the incorporation of intrabasinal carbonate bioclasts into the sand deposits, which act as potential sources and nuclei for carbonate cementation (Ketzer *et al.*, 2002). The extent of concretionary and continuous carbonate cementation is large in TST and early HST sandstones (Figs. 11 and 12B). Upward increase of carbonate cements in shoreface TST sand deposits is probably enhanced by upward increase in bioturbation, which acts as sites for local increase in carbonate alkalinity by decay of organic matter (Curtis, 1987; Wilkinson, 1991; Morad *et al.*, 2000; Al-Ramadan *et al.*, 2005; Ketzer *et al.*, 2002).

The close association of substantial amounts of pyrite with calcite and dolomite cements is common in organic-matter rich, TST and early HST deltaic, paralic and shelf sandstones. Pyrite formation occurs by bacterial reduction of dissolved sulphate into sulphide ions, which react with dissolved $Fe^{2+}$ derived from the reduction of Fe-oxides and oxyhydroxides (Berner, 1982). Diagenetic apatite, which is rare, and minor cement in siliciclastic sediments, displays trend of upward increase within the TST towards the MFS, particularly along shelf edge and upper slope (Parrish & Curtis, 1982; Edman & Surdam, 1984). The precipitation of apatite is favoured by the presence of abundant organic matter, which is related to upwelling of deep oceanic waters (Burnett, 1977; Glenn *et al.*, 2000).

Marine transgression is also accompanied by a systematic upward increase in the amounts and maturity (i.e., increase in K content) of glaucony within the TST and early HST (Amorosi, 1995), reaching a maximum below the MFS (Fig. 12E). The distribution of glaucony is related to its type and origin; autochthonous glaucony refers to grains formed in situ within the sediment framework, while allochthonous glaucony refers to grains reworked and re-deposited within the same sedimentary sequence. Detrital or extraformational glaucony includes grains derived by erosion of older sequences (Amorosi, 1995: Amorosi this volume). However, autochthonous glaucony deposited along shelf edges may be reworked by waves, tides or storms at parasequence boundaries. In the LST, the reworking of glaucony by storms to shelf and estuarine environments and by turbidity currents to deep water fans results in the deposition of parautochthonous glaucony (Amorosi, 1995).

The TST and early HST are preferential sites for the occurrence of coastal coal deposits (Ryer, 1981; Cross, 1988; Shanley & McCabe 1993). Thus, diagenetic alterations related to coal at parasequence boundaries, such as the formation of pyrite, extensive calcite cement and kaolinite, will potentially be more common or extensive within TST and early HST (Love *et al.*, 1983; Ketzer *et al.*, 2003a).

Estuarine-deltaic sandstones (TST) are commonly rich in grain coating berthierine or odinite as ooids or coatings on sand grains (Fig. 12F; Odin, 1990; Ehrenberg, 1993; Hornibrook & Longstaffe, 1996; Kronen & Glenn, 2000; Fig. 6). High flux rates of organic matter and detrital Fe-oxides and oxyhydroxides by rivers promote a rapid establishment of post-oxic, Fe-reducing geochemical conditions, which favour the formation of these Fe-silicates (Odin, 1988, 1990; Aller, 1998) The formation of these clay minerals is presumably enhanced by the low sulphate concentration in pore-waters (i.e., less $Fe^{2+}$ is incorporated in pyrite and other Fe-sulphides) caused by mixing of marine and meteoric waters during shoreline progradation. Berthierine and odinite are precursors for the formation of ferroan chlorite (chamosite) during burial diagenesis in late HST and lowstand wedge sandstones (Fig. 6). Continuous pore-lining chlorite, which is commonly derived from grain-coating Fe-clay (e.g. odinite) precursor, has been reported to effectively preserve anomalously high porosity in deeply buried reservoir sandstones (Ehrenberg, 1993; Ryan & Reynolds, 1996; Bloch *et al.*, 2002). Chlorite rims (Fig. 12D) may also evolve from pore-lining smectite, particularly in sandstones rich in detrital Fe-silicates and/or volcanic rock fragments (Humpreys *et al.*, 1994; Anjos *et al.*, 2003; Salem *et al.*, 2005).

Another diagenetic feature characteristic of TST shallow and deep marine sandstones is silica authigenesis, particularly as opal, opal-CT, chalcedony and microquartz coatings, rims and pore-filling aggregates, as well as replacing mud intraclasts and derived pseudomatrix (Sears, 1984; van Bennekon *et al.*, 1989; Hendry & Trewin, 1995; Aase *et al.*, 1996; Lima & De Ros, 2002). These occurrences are normally related to the availability of silica by the dissolution of biogenic opal from radiolarians, diatoms and sponge spicules concentrations favoured by the transgressive setting. Microcrystalline and cryptocrystalline silica coatings and rims may help to preserve the porosity in deep sandstone reservoirs (Hendry & Trewin, 1995; Aase *et al.*, 1996; Lima & De Ros, 2002) but may promote resistivity anomalies problematic to wireline log evaluation of oil saturation.

During continuous sedimentation, shallow marine sediments deposited during late HST display upward shallowing, coarsening and thickening of sandstone bodies, accompanied by a trend of decrease in degree of bioturbation. Upon fall in the relative sea-level and formation of a regressive surface of marine erosion, deposition of falling stage systems tract occur by aggradation of shoreface deposits (Hunt & Tucker, 1992; Miall, 2000). A pause in fall of the relative sea-level results in re-establishment of shoreface conditions and deposition of shoreface sand (called sharp-based sand bodies) on the regressive erosion surface. These sand bodies are cemented by poikilotopic calcite, which may form large (e.g. >1 m diameter) strata-bound concretions (Al-Ramadan *et al.*, 2005). A major fall in relative sea-level and exposure of the shoreface sand is accompanied by their erosion by prograding fluvial systems, which is accompanied by dissolution of calcite cement and bioclasts as well as framework silicates dissolution and kaolinite.

## CONCLUDING REMARKS

- The integration of diagenesis into the sequence stratigraphic framework (i.e. the interplay between the rates of changes in the relative sea-level and rates of sedimentation) of siliciclastic and carbonate successions allows the development of predictive conceptual models for the reservoir-quality evolution pathways. These models constrain preferential sites for cementation (i.e. porosity and permeability destruction) or dissolution (i.e. porosity and permeability enhancement).
- Precipitation of diagenetic minerals such as calcite, dolomite, siderite, pyrite, kaolinite, glaucony and berthierine/odinite and formation of pseudomatrix, mechanical clay infiltration and intragranular porosity show a systematic distribution in sandstones lying in the vicinity of sequence boundaries (SB) and parasequence boundaries (PB), transgressive surfaces (TS) and maximum flooding surfaces (MFS) and in sandstones of the lowstand (LST), transgressive (TST) and highstand (HST) systems tracts.
- The main sequence stratigraphic controls on the distribution and type of diagenetic alterations in siliciclastic successions include: (i) detrital composition (mainly the proportion and type of intra- and extrabasinal grains), (ii) pore water chemistry, (iii) presence and quantity of organic matter and (iv) residence time of the sediments under specific geochemical conditions. The last three parameters control also the sequence stratigraphic distribution of diagenetic alterations in carbonate successions.
- Climatic conditions prevailing during subaerial exposure of the sediments due to a major fall in the relative sea-level (i.e. formation of a sequence boundary) have a profound impact on the types and extent of diagenetic alterations. Under humid climatic conditions, reservoir-quality of sandstones is enhanced by meteoric-water percolation beneath SB owing to the dissolution and kaolinization of feldspars, rock fragments and micas. Reservoir-quality enhancement of carbonate successions below SB occurs by karstification. Semi-arid climatic conditions may result in deterioration of porosity and permeability of sandstone successions can be deteriorated immediately below the SB by mechanical clay infiltration and development of calcrete/dolocrete, which act as baffles or barriers for fluid flow.
- The PB, TS and MFS are common sites of porosity destruction in sandstones and carbonate successions due to extensive carbonate cementation (i.e. development of hardgrounds and firmgrounds). Cementation is attributed to long residence time of the sediments at shallow depths below the seafloor and hence extensive diffusive flux of dissolved calcium and carbon from the overlying sea water into the pore waters. Therefore, these surfaces can, thus, form potential baffles and barriers for fluid flow, which create reservoir compartments even between parasequences.
- Lower oxygen isotopic values than expected for marine carbonate cement in sandstones along PB, TS and MFS may indicate that: (i) cementation may commence immediately below the seafloor and continue during burial diagenesis, or (ii) eogenetic carbonate cement is subjected to recrystallization by meteoric waters or at elevated temperatures.

- The presence of peat/coal layers, which occur along marine transgressive surfaces (e.g. PB), favours the growth of concretionary pyrite and continuous calcite cementation in the underlying and overlying sandstones. The degradation of plant remains in these layers induces anoxic pore water conditions and concomitant increase in carbonate alkalinity and thus results in the precipitation of pyrite and carbonate cement, respectively.
- The TST and early HST paralic sandstones are more prone to porosity deterioration owing to carbonate cementation than LST and late HST deposits. This difference is encountered because TST and early HST deposits are more likely to incorporate intrabasinal carbonate grains into the sand deposits, which act as nuclei and source of ions for carbonate cementation.
- The TST estuarine and deltaic deposits are prone to the formation of grain-coating Fe-silicates, which eventually evolve to chlorite rims during burial diagenesis. Such chlorite inhibits or retard extensive cementation by syntaxial quartz overgrowths and thus helps preserving anomalously high porosity in these sandstones.
- Extensive percolation of meteoric waters into the fluvial, incised valley filling sandstones (late LST) causes greater extent of porosity enhancement by framework silicate grains dissolution than TST and HST sandstones.
- The transformation of kaolinite and grain-coating smectitic clays into illite in the fluvial, incised valley sandstones (late LST) during burial diagenesis is favoured by contemporary dissolution and albitization of detrital K-feldspar. Illitization may result in considerable deterioration to permeability of these sandstones.
- Diagenesis of carbonate sediments is characterized by the formation of marine calcite cement in the TST, which increases in abundance towards the MFS. Conversely, the HST carbonates are characterized by sparse amounts of mixing zone dolomite and equant and drusiform calcite and considerable amounts of moldic and vuggy porosity.
- It is suggested that the presented patterns of linkages between diagenesis and sequence stratigraphic framework should be tested in a larger variety of settings and depositional environments. A greater challenge is to apply these concepts to marine turbidite reservoirs, which represent the ultimate frontier for hydrocarbon exploration.

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