UNIVERSIDADE FEDERAL DO RIO GRANDE DO SUL

INSTITUTO DE GEOCIÊNCIAS

PROGRAMA DE PÓS-GRADUAÇÃO EM GEOCIÊNCIAS

CONTROLES DEPOSICIONAIS E DIAGENÉTICOS DAS PROPRIEDADES PETROFÍSICAS DOS RESERVATÓRIOS APTIANOS/BARREMIANOS DO GRUPO LAGOA FEIA NO NORTE DA BACIA DE CAMPOS

RONALDO HERLINGER JUNIOR

Orientador – Prof. Dr. Luiz Fernando de Ros

Porto Alegre – 2016

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Porto Alegre, 2016

Herlinger Junior, Ronaldo

Controles deposicionais e diagenéticos das propriedades petrofísicas dos reservatórios Aptianos/Barremianos do Grupo Lagoa Feia no norte da Bacia de Campos . / Ronaldo Herlinger Junior. - Porto Alegre: IGEO/UFRGS, 2016.

[124 f.] il.

Dissertação (Mestrado).- Universidade Federal do Rio Grande do Sul. Programa de Pós-Graduação em Geociências. Instituto de Geociências. Porto Alegre, RS - BR, 2016.

Orientador(es):Luiz Fernando De Ros

1. Carbonatos lacustres 2. Diagênese 3. Porosidade 4. Permeabilidade I. Título.

CDU 55

Catalogação na Publicação
Biblioteca Instituto de Geociências - UFRGS
Veleida Ana Blank CRB 10/571

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Para Alice e Ivan.

AGRADECIMENTOS

À minha esposa Alice, pelo apoio e paciência durante o tempo que tive que me ausentar e dedicar-me ao trabalho.

Ao meu filho Ivan, que me fez entender o significado da palavra superação.

Ao meu orientador e amigo De Ros, por ceder parte de seu tempo escasso e passar um pouco de seu vasto conhecimento.

À coorientadora informal Eveline, sempre prestativa e pronta a auxiliar.

Ao Garibaldi, pelas dicas petrográficas e musicais.

Ao meu gerente imediato Carlos Pedroso Jr. pelo incentivo e liberação parcial para realização deste trabalho.

Ao gerente de geologia de reservatórios Bruhn, pelo auxílio com a liberação dos dados.

À gerente do laboratório Rute, pela facilitação ao acesso das instalações do laboratório de sedimentologia da UO-ES.

À empresa Petrobras, sem a qual não seria possível a realização deste trabalho.

“As pessoas que resolviam as coisas em geral tinham muita persistência e um pouco de sorte. Se a gente persistisse o bastante, a sorte em geral chegava. Mas a maioria das pessoas não podia esperar a sorte, por isso desistia.”

Charles Bukowski

RESUMO

Os reservatórios lacustres do Grupo Lagoa Feia, seção rift da Bacia de Campos, margem Leste brasileira, tem mantido há décadas uma expressiva produção a partir de campos localizados em águas rasas. A descoberta de grandes acumulações na seção *rift* e *sag* (pré-sal) da Bacia de Santos reativou a exploração por reservatórios análogos na Bacia de Campos e em outras bacias marginais. Um estudo petrográfico e petrofísico sistemático foi executado sobre os reservatórios *rift* da Formação Coqueiros e *sag* da Formação Macabu do Norte da Bacia de Campos, com objetivo de caracterizar os principais controles sobre a gênese e evolução daqueles reservatórios não-convencionais e seus sistemas porosos. As principais petrofácies de reservatório reconhecidas foram *grainstones* e *rudstones* bioclásticos, arenitos ooidais argilosos e dolomitos na Formação Coqueiros, e crostas coalescentes e não-coalescentes de calcita fascicular, rudstones e grainstones intraclásticos e dolomitos na Formação Macabu. A evolução dos reservatórios bioclásticos foi controlada pelo balanço entre dissolução ou neomorfismo dos bioclastos aragoníticos de bivalves, favorecendo a geração de porosidade móldica pouco conectada, ou a preservação da porosidade interpartícula bem conectada, controlando a permeabilidade dos reservatórios. Os arenitos de oóides de argilas magnesianas sofreram dissolução e substituição por dolomita e sílica, o que gerou sistemas porosos altamente heterogêneos, compostos por poros móldicos, intercristalinos, vugulares e microcristalinos. O crescimento de agregados cristalinos arborescentes nas crostas coalescentes de calcita gerou porosidade de crescimento do arcabouço primária, que foi reduzida principalmente por cimentação dolomítica, ou alargada por dissolução, o que ampliou sua permeabilidade. Crostas não-coalescentes de calcita mostram forte interação com argilominerais magnesianos, que preenchem interstícios, e/ou estão intercalados com as crostas. Sua porosidade está relacionada com a dissolução das argilas, o que produziu baixa permeabilidade. Rochas intraclásticas comumente mostram matriz argilosa, ou estão compactadas e cimentadas. Onde exibem porosidade interpartícula primária ou dissolução da matriz, podem ter boas porosidades e permeabilidades. A dolomitização heterogênea de ambas as Formações em alguns casos destruiu a porosidade primária ou eodiagenética, ou em outros gerou altos valores de porosidade e permeabilidade nos dolomitos. Relações de substituição e de compactação indicam que muitos dos processos diagenéticos ocorreram durante a eodiagênese, controlados principalmente pela instabilidade da aragonita nos reservatórios bioclásticos da Formação Coqueiros, e dos argilominerais magnesianos na Formação Macabu. Este estudo representa a primeira caracterização petrográfica publicada dos reservatórios carbonáticos não-convencionais do *sag*, e salienta a importância crucial dos estudos petrográficos sistemáticos para a compreensão e previsão da qualidade de reservatórios complexos.

Palavras-chave: Grupo Lagoa Feia, carbonatos lacustres, diagênese, porosidade, permeabilidade.

ABSTRACT

Lacustrine carbonate reservoirs from the Lagoa Feia Group, rift section of Campos Basin, offshore eastern Brazil, have sustained for decades a significant production from shallow water oil fields. The discovery of giant accumulations in the rift and sag (pre-salt) section of the adjacent Santos Basin has reactivated the exploration for equivalent reservoirs in the Campos Basin and in other marginal basins. A systematic petrographic and petrophysical study was performed on the rift Coqueiros Formation and the sag Macabu Formation carbonates from the Lagoa Feia Group in northern Campos Basin, in order to characterize the main controls on the origin and evolution of those unconventional reservoirs and their pore systems. The main types of reservoir petrofacies recognized were grainstones and bioclastic rudstones, magnesian clay ooidal arenites and dolostones from the Coqueiros Formation; coalescent and non-coalescent crusts of fascicular calcite, intraclastic rudstones and grainstones, and dolostones from the Macabu Formation. The evolution of bioclastic reservoirs was controlled by the balance between dissolution and neomorphism of the aragonitic bivalve bioclasts, favoring the generation of poorly-connected moldic porosity or the preservation of well-connected interparticle porosity, which controlled the permeability of the reservoirs. The magnesian clay (stevensite) ooidal arenites suffered dissolution and replacement by dolomite and silica, what generated highly heterogeneous pore systems, composed by moldic, intercrystalline, vugular and microcrystalline pores. The growth of crystal shrubs in coalescent calcite crusts generated growth-framework primary porosity, which was reduced mostly by dolomite cementation, or enlarged by dissolution, what enhanced their permeability. Non-coalescent calcite crusts usually show strong interaction with syngenetic magnesian clay minerals, which fill interstices and/or are interbedded with the crusts. Their porosity is related to dissolution of the clays, what generated poor permeability. Intraclastic rocks usually display clay matrix, or are compacted and cemented. Where they show interparticle primary porosity or matrix dissolution, they may have good porosities and permeabilities. The heterogeneous dolomitization of both formations, either destroyed the primary or early diagenetic porosity, or generated high porosity and permeability values in the dolostones. Relationships of replacement and compaction indicate that most of the diagenetic processes occurred during eodiagenesis, controlled mostly by the instability of the aragonite in the bioclastic Coqueiros reservoirs, and of the magnesian clay minerals in the Macabu Formation. This study represents the first published petrographic characterization of the unconventional sag carbonate reservoirs, and stresses the crucial importance of systematic petrographic studies for the understanding and prediction of the quality of complex reservoirs.

Key words: Lagoa Feia Group, lacustrine carbonates, diagenesis, porosity, permeability.

Sumário

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ESTRUTURA DA DISSERTAÇÃO

Esta dissertação está estruturada em torno de artigo científico submetido para publicação em periódico indexado, que compreende o estudo petrográfico e petrofísico dos reservatórios Aptianos/Barremianos das Formações Coqueiros e Macabu do Grupo Lagoa Feia no Norte da Bacia de Campos, e tem como objetivo o entendimento da geração e evolução da porosidade e permeabilidade dos reservatórios. Sua organização compreende as seguintes seções:

a) Sumário Integrado da dissertação, incluindo descrição do objeto da pesquisa de Mestrado, contendo o estado da arte sobre o tema e metodologia desenvolvida, seguido de uma síntese integradora com os principais resultados do trabalho.

b) Artigo científico intitulado *Depositional and Diagenetic Controls on the Quality of Lacustrine Pre-Salt Carbonate Reservoirs from Northern Campos Basin, Offshore Brazil* submetido à revista *Marine and Petroleum Geology*, em 01 de fevereiro de 2016.

c) CD-ROM contendo seguintes anexos:

   a. Versão digital do Sumário Integrado da dissertação;

   b. Versão digital do artigo científico;

   c. Descrições petrográficas;

   d. Documentação fotomicrográfica;

   e. Análises de microscopia eletrônica de varredura;

   f. Análises de difração de raios-x;

   g. Análises de petrofísica de laboratório.

1 Introdução

Os reservatórios carbonáticos da Formação Coqueiros do Grupo Lagoa Feia, depositada durante a fase rifte da Bacia de Campos, tem sido explorados há décadas, com significativa produção em campos localizados em lâmina d'água rasa. Esses reservatórios, representados por *rudstones e grainstones* bioclásticos, foram intensamente estudados, principalmente durante as décadas de 80 e 90 (BAUMGARTEN, 1985; BERTANI; CAROZZI, 1985a, 1985b; BAUMGARTEN *et al.*, 1988; DIAS; OLIVEIRA; VIEIRA, 1988; ABRAHÃO; WARME, 1990; HORSCHUTZ; SCUTA, 1992; CARVALHO *et al.*, 2000; CASTRO, 2006), apresentando melhores porosidades e permeabilidades do que as rochas da Formação Macabu, depositadas durante o estágio *sag* da bacia. Após a descoberta de expressivas acumulações de hidrocarbonetos leves com alto valor comercial em reservatórios da fase *sag* da Bacia de Santos, localizados em águas profundas na província petrolífera que ficou conhecida como Pré-Sal, os reservatórios análogos das Bacias de Campos e de outras bacias voltaram a ser estudados (CASTRO, 2011; ALTENHOFEN, 2013; MUNIZ, 2014; THOMPSON; STILWELL; HALL, 2015). A Bacia de Campos contribui atualmente com a maior parte da produção de petróleo do Brasil, principalmente de reservatórios turbidíticos e subordinadamente de carbonatos Albianos e Aptianos/Barremianos.

Este trabalho tem como foco o estudo petrográfico e petrofísico dos reservatórios do Grupo Lagoa Feia no Norte da Bacia de Campos, incluindo tanto os depósitos bioclásticos da Formação Coqueiros, quanto os depósitos da Formação Macabu, cuja gênese ainda é altamente controversa. A partir de estudos petrográficos com microscopia ótica e microscopia eletrônica de varredura e análises petrofísicas pretende-se caracterizar o sistema poroso dos reservatórios estudados, de forma a entender sua gênese e evolução, em busca de uma melhor compreensão e previsibilidade da distribuição da porosidade e permeabilidade nas rochas-reservatório do Grupo Lagoa Feia.

2 Localização da Área de Estudo

A Bacia de Campos está situada na costa Norte do Estado do Rio de Janeiro (Figura 1), estendendo-se até o Sul do Estado do Espírito Santo, entre os paralelos 21º e 22º Sul. É limitada ao norte pelo Alto da Vitória, que a separa da Bacia do Espírito Santo, e ao sul pelo Alto de Cabo Frio, que a separa da Bacia de Santos. Compreende uma área total de aproximadamente 120.000 km^2, com exposição *onshore* de apenas cerca de 5800 km^2. A área de estudo localiza-se em águas profundas ao Norte da bacia, compreendendo um total de 5 poços.

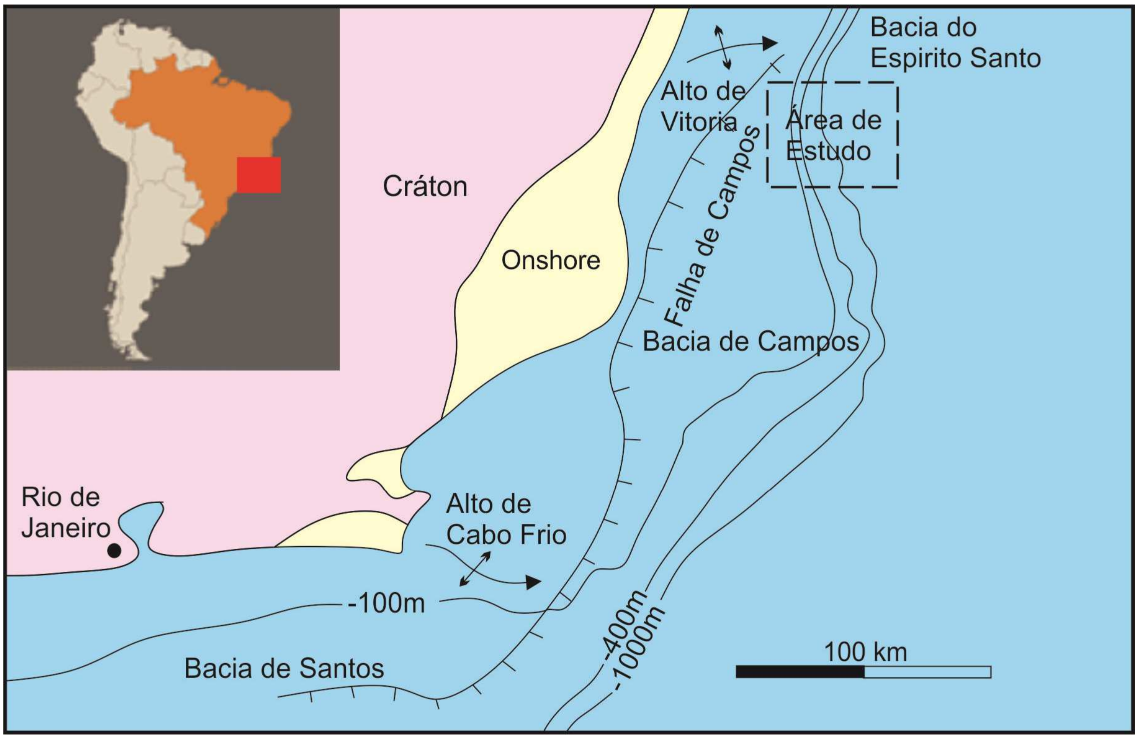

Figura 1. Mapa de localização da área de estudo no Norte da Bacia de Campos.

3 Geologia Regional

3.1 Contexto Tectônico Regional

A Bacia de Campos foi formada em um regime tectônico extensivo durante o rompimento do Continente Gondwana no Neocomiano (GUARDADO *et al.*, 2000), antecedendo a separação da América do Sul e África, que deu origem ao Oceano Atlântico, podendo ser classificada como uma bacia de margem passiva, segundo a classificação de KLEMME (1980). Está localizada no lado leste do Atlântico Sul e tem a bacia Kwanza como correspondente no lado africano.

A formação do Oceano Atlântico se deu a partir de um afinamento crustal e ruptura dos Crátons São Francisco-Congo-Rio de la Plata-Kalahari e o cinturão de dobramento Proterozóico, com posterior formação do assoalho oceânico durante o Cretáceo Superior – Jurássico Inferior. O rompimento do Atlântico Sul inicia-se com um rifteamento ao sul, na região da Argentina, durante o Jurássico e através da margem equatorial (MOHRIAK; NEMCOK; ENCISO, 2008). A porção central teve um início de rompimento tardio no Hauveteriviano (CLEMSON; CARTWRIGHT; BOOTH, 1997), controlado por um núcleo cratônico resistente (Cráton São Francisco - Congo), resultando no desenvolvimento de uma bacia de rifte estreita nesta região. Porções controladas pelo cinturão de dobramento Proterozóico desenvolveram bacias mais largas. Mohiak, Nemcok e Enciso (2008) definiram 5 estágios de evolução das bacias marginais brasileiras (Figura 2):

O primeiro estágio de formação das bacias sedimentares marginais brasileiras é denominado Pré-Rifte, e corresponde à extensão litosférica que depois levou à separação da África e América do Sul no Cretáceo Superior – Jurássico Inferior. O modelo para esta fase corresponde à um pequeno grau de soerguimento litosférico e afinamento crustal e do manto superior, com falhamento incipiente da crosta superior, que controlam depocentros locais (CAINELLI; MOHRIAK, 1999).

O estágio rifte compreende a segunda fase de evolução (CHANG; KOWSMANN; FIGUEIREDO, 1988; CHANG *et al.*, 1992), sendo caracterizado por um aumento na extensão litosférica e soerguimento astenosférico (MCKENZIE, 1978; WHITE; MCKENZIE, 1988). Este representa o principal episódio de rifteamento intracontinental, associado com extenso vulcanismo toleiítico intracratônico e ao longo da margem continental (MIZUSAKI; THOMAZ FILHO; CESERO, 1998; MEISLING; COBBOLD; MOUNT, 2001). O evento magmático foi seguido de um falhamento normal que afetou toda a crosta continental, gerando meios-grábens localizados à leste do Cinturão Ribeira (MOHRIAK; NEMCOK; ENCISO, 2008). Os estresses extensionais da segunda fase concentraram-se ao longo de riftes intracontinentais, desenvolvidos como uma série de lagos alongados e profundos, paralelos à costa e aos alinhamentos do embasamento adjacente (DIAS *et al.*, 1990), preenchidos com rochas vulcânicas e sedimentares Neocomianas a

Barremianas. A presença de zonas de transferência teve importante papel na compartimentação da margem rifteada em sub-bacias (MEISLING; COBBOLD; MOUNT, 2001).

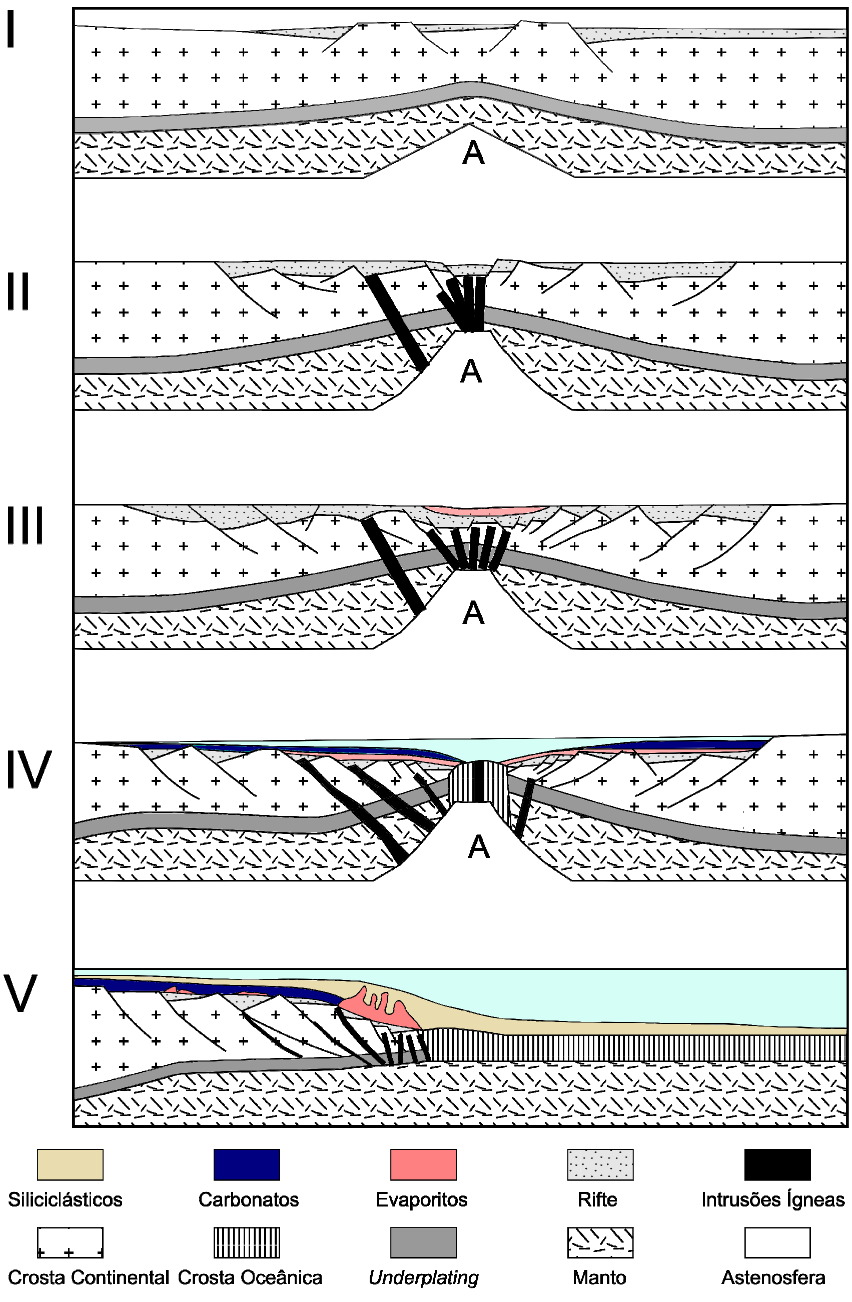

Figura 2. Representação esquemática das cinco fases de evolução tectônica da Bacia de Campos (MOHRIAK; NEMCOK; ENCISO, 2008).

A fase Pós-Rifte, Barremiana a Aptiana, representa o fim de extensão sin-rifte, sendo caracterizada pela reativação de grandes falhas e erosão da fase rifte, resultando em uma discordância regional. Esta discordância separa os sedimentos lacustres continentais da fase rifte de sedimentos lacustres da fase *sag*, Estes sedimentos são sobrepostos por evaporitos formados em clima árido após incursões marinhas (ABRAHÃO; WARME, 1990).

A quarta fase é caracterizada pelo desenvolvimento da crosta oceânica (CHANG *et al.*, 1992; MOHRIAK; NEMCOK; ENCISO, 2008). Após a ruptura, a maior parte da atividade tectônica esteve relacionada com a evolução do Oceano Atlântico, com subsidência térmica típica de margens passivas. As sucessões sedimentares pós-Aptiano correspondem a carbonatos do Albiano, sugerindo um ambiente de águas rasas que progressivamente tornaram-se mais profundas.

O último estágio estende-se do Albiano ao recente, representando um aumento progressivo da batimetria. Após o Albiano, a tectônica é dominada por falhas relacionadas a halocinese, de geometria lístrica, com anticlinais e calhas associadas, domos e diápiros de sal e estruturas geneticamente relacionadas (MOHRIAK; NEMCOK; ENCISO, 2008).

3.2 Estratigrafia da Bacia de Campos

A evolução tectono-estratigráfica da Bacia de Campos está relacionada ao rompimento do continente Gondwana e evolução do oceano Atlântico Sul. Dentro deste contexto, Winter; Jahnert e França (2007) dividem a estratigrafia de acordo com os estágios de evolução da Bacia, que compreendem as Supersequências Rifte, Pós-Rifte e Drifte (Figura 3).

A Supersequência Rifte foi formada durante o estágio rifte da bacia e corresponde à Formação Cabiúnas, de idade Halteriviano e à porção Basal do Grupo Lagoa Feia, composto pelas Formações Atafona, Itabapoana e Coqueiros, depositadas do Barremiano ao Aptiano Inferior.

A Supersequência Pós-Rifte corresponde às rochas depositadas sobre a discordância Pré-Alagoas. É composta pelas Formações Itabapoana, Macabu e Gargau, depositadas no Aptiano Médio/Superior, Andar Alagoas, e Formação Retiro (Andar Albiano Inferior/Alagoas Superior).

A Superseqüência Drifte é formada por sedimentos marinhos depositados durante um regime de subsidência térmica associada com tectonismo adiastrófico. É composta pelo Grupo Macaé, de Idade Albiana/Cenomaniana e o Grupo Campos, de idade Turoniano ao Recente.

3.2.1 Formação Cabiúnas

A Formação Cabiúnas é composta de derrames ígneos subalcalinos, subaéros e subaquosos, representados por basaltos, diabásios e rochas vulcanoclásticas (MIZUSAKI; THOMAZ FILHO; CESERO, 1998). Localmente são encontrados arenitos e siltitos como intertrapes entres as sucessões de derrames. Datação através de isótopos K-Ar indica idades de 111 a 134 Ma (MIZUSAKI *et al.*, 1992).

3.2.2 Grupo Lagoa Feia

A porção basal do Grupo Lagoa Feia, depositada na fase rifte, tem como limites uma discordância no topo da Formação Cabiúnas, na porção basal, e a discordância Pré-Alagoas no topo. É constituída pelas Formações Itabapoana, Atafona e Coqueiros. A Formação Itabapoana é constituída de sedimentos proximais da borda da bacia, compostos de conglomerados, arenitos, siltitos e folhelhos avermelhados. A Formação Atafona é composta de arenitos e siltitos depositados em lagos alcalinos e constituídos essencialmente por argilominerais magnesianos singenéticos (BERTANI; CAROZZI, 1985a, 1985b; REHIM et al., 1986; DIAS; OLIVEIRA; VIEIRA, 1988; ABRAHÃO; WARME, 1990). A Formação Coqueiros é formada pela intercalação de folhelhos ricos em matéria orgânica e carbonatos continentais compostos dominantemente de bivalves (BAUMGARTEN *et al.*, 1988; DIAS; OLIVEIRA; VIEIRA, 1988; CARVALHO *et al.*, 2000; CASTRO, 2006).

A porção superior do Grupo Lagoa Feia, depositado no estágio *sag*, é constituída pela Formação Itabapoana, descrita anteriormente, e pelas Formações Gargaú, Macabu e Retiro. A Formação Gargaú é formada por arenitos, margas e carbonatos depositadas em ambiente lacustre (WINTER; JAHNERT; FRANÇA, 2007), gradando para os sedimentos clásticos da Formação Itabapoana. A Formação Macabu foi acumulada em lagos alcalinos sob regime climático árido, onde foram precipitados argilominerais magnesianos e calcita fascicular-ótica e fibrorradial na forma de crostas. A Formação Retiro corresponde a uma espessa acumulação de evaporitos (Alagoas Superior), composta essencialmente de anidrita, halita, silvita e carnalita (WINTER; JAHNERT; FRANÇA, 2007), formada durante incursões marinhas, sob condições de clima árido (LEYDEN *et al.*, 1976). Os evaporitos da Formação Retiro possuem grande importância econômica, visto que constituem o selo dos volumes gigantes de hidrocarbonetos da província petrolífera conhecida como Pré-Sal na Bacia de Campos, permitindo a acumulação de hidrocarbonetos nos reservatórios das Formações Coqueiros e Macabu, que são o foco deste trabalho.

3.2.3 Grupo Macaé

O Grupo Macaé corresponde a sedimentos carbonáticos depositados em ambiente de plataforma rasa (Formação Quissamã), gradando a marinho profundo (Formação Outeiro e Imbetiba), sedimentos clásticos depositados em regiões proximais (Formação Goitacás), e marinho profundo representados pela Formação Namorado (WINTER; JAHNERT; FRANÇA, 2007). A Formação Goitacás é representada por conglomerados, arenitos, margas e calcilutitos depositados em ambientes aluviais que gradam a marinho raso (RANGEL *et al.*, 1994). A Formação Quissamã é composta de *grainstones* oolíticos e oncolíticos, depositados em ambiente de alta energia associados a *packstones* bioclásticos, oolíticos e oncolíticos depositados em barras de energia moderada (BRITO; OLIVEIRA, 2011). A Formação Outeiro é constituída de calcilutitos, margas e folhelhos, por vezes com a intercalação de camadas arenosas isoladas da Formação Namorado (RANGEL *et al.*, 1994). A Formação Namorado, por sua vez, é formada por depósitos turbidíticos, constituídos dominantemente de arenitos finos a muito finos compostos de quartzo, feldspato e litoclastos (BARROSO *et al.*, 1998). A Formação Imbetiba constitui a parte superior do Grupo Macaé, sendo dominada por pelitos e margas intercalados com esparsas ocorrências de corpos turbidíticos (WINTER; JAHNERT; FRANÇA, 2007).

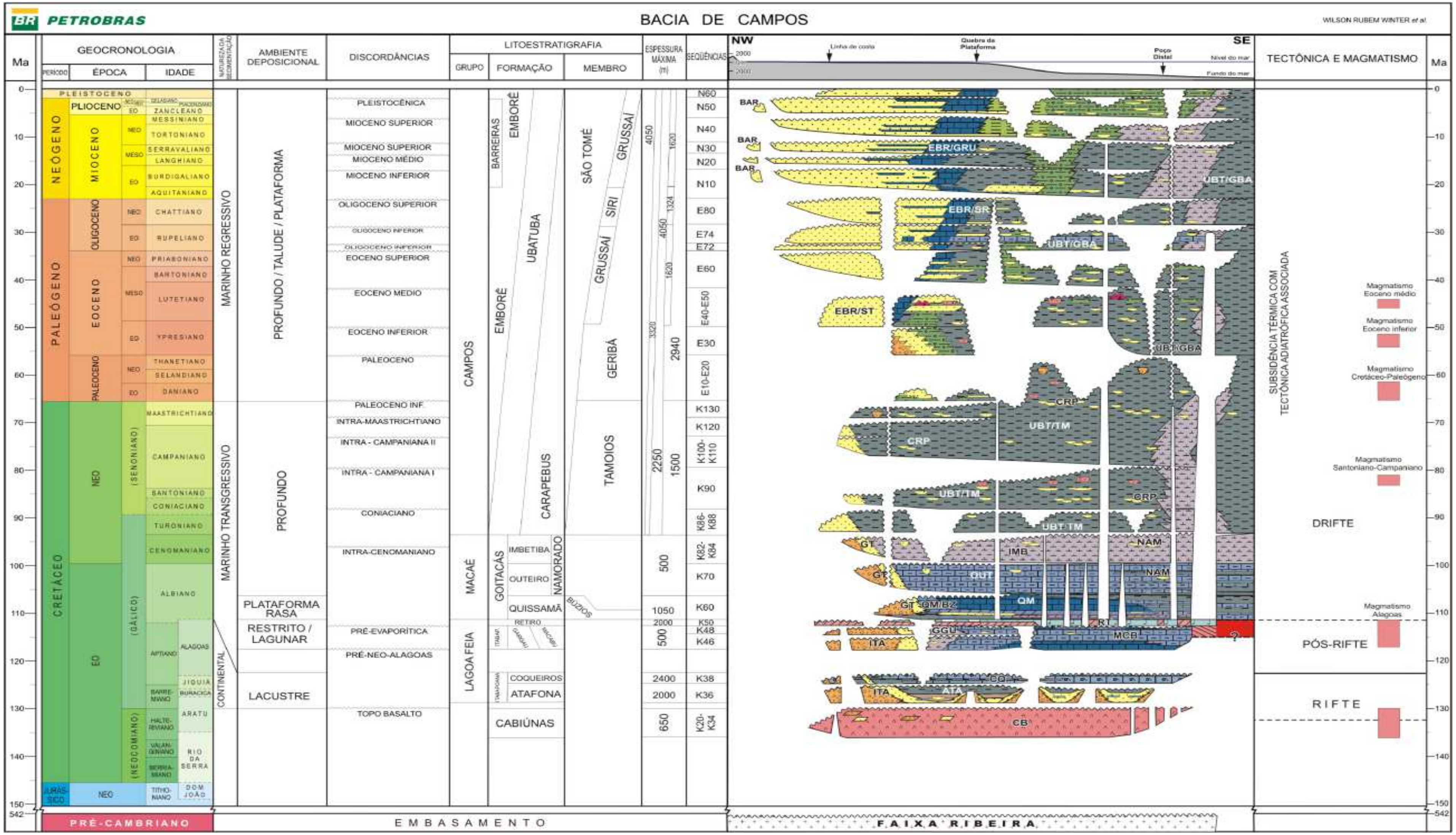

Figura 3. Coluna estratigráfica da Bacia de Campos (WINTER; JAHNERT; FRANÇA, 2007).

4 Revisão Conceitual

4.1 Classificação de Rochas Carbonáticas

As classificações das rochas carbonáticas são fortemente influenciadas pelo contexto geológico das rochas estudas pelos autores que as propuseram. Desta forma, as classificações existentes não são abrangentes o suficiente para caracterizar a ampla gama de variação faciológica encontrada nas rochas carbonáticas. Duas das classificações mais utilizadas foram propostas por Folk (1959) e Dunhan (1962). Ambas as classificações subdividem os carbonatos com base no conteúdo de matriz versus arcabouço das rochas. A classificação de Folk distingue as litologias através da proporção de partículas, lama carbonática e cimento espático, juntamente com o tipo de aloquímico predominante (Figura 4), tendo uma importante relação com a energia do sistema onde as rochas foram depositadas.

A classificação de Dunham, assim como a classificação de Folk, faz distinção entre os tipos litológicos com base no suporte da fábrica e proporção entre lama e partículas, definindo os termos *mudstone, wackestone, packstone* e *grainstone*, referentes a rochas onde os componentes não estavam ligados durante a deposição. Além dessas, esse autor criou a denominação *boundstone* para rochas bioconstruídas (Figuras 5).

Posteriormente, Embry e Klovan (1971) criaram uma terminologia específica para rochas bioconstruídas (Figura 6). Esses autores separaram as rochas em alóctones, onde os sedimentos não são ligados organicamente durante a deposição, e autóctones, com sedimentos ligados durante a deposição. Os sedimentos alóctones possuem distinção pela forma de suporte da fábrica e a proporção entre lama e partículas. Já os sedimentos classificados como autóctones são classificados de acordo com a estrutura da bioconstrução.

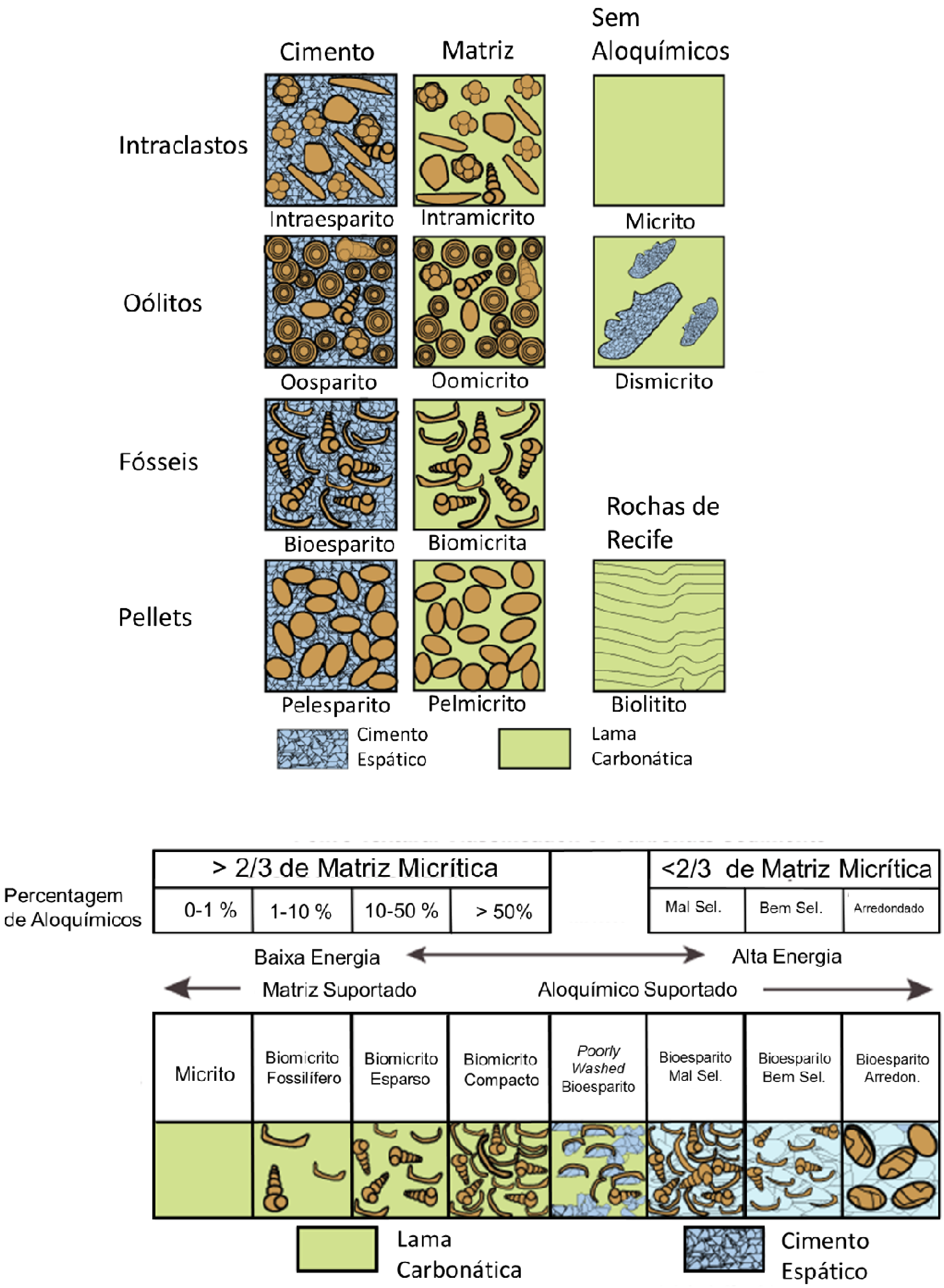

Figura 4. Classificação das rochas carbonáticas, segundo Folk (1959).

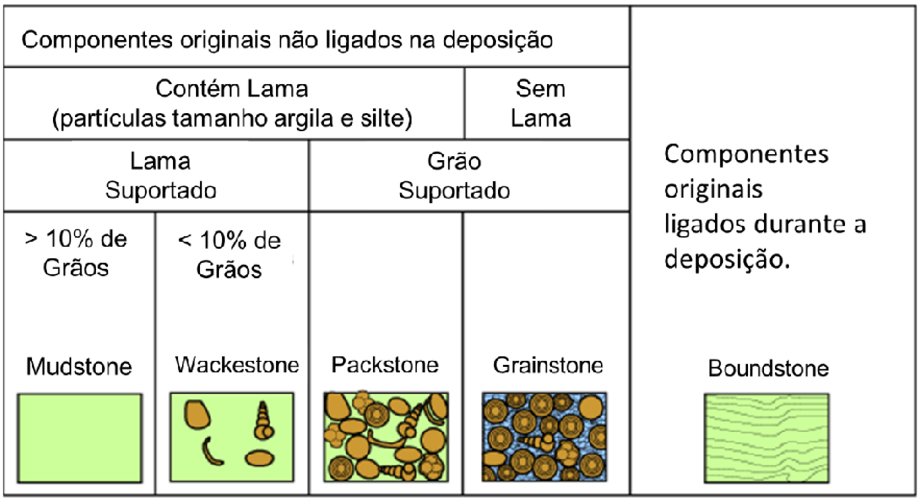

Figura 5. Classificação de Dunhan (1962).

Mais recentemente, Wright (1992) propôs uma unificação entre a classificação de Dunhan (1962) e a classificação de Embry e Klovan (1971). Este autor baseou-se no pressuposto que as rochas carbonáticas tem sua textura controlada pela combinação de fatores deposicionais, atividade biológica e diagênese (Figura 7). Dessa forma, a classificação divide as rochas carbonáticas em 3 grupos. No primeiro, chamado Deposicional, são agrupadas as rochas formadas por partículas, adaptado da classificação de Dunham. No segundo, denominado Biológico, são agrupadas as rochas bioconstruídas, assim como definidas por Embry e Klovan (1971). No terceiro tipo, são agrupadas as rochas onde os processos diagenéticos condicionaram ou mudaram fortemente sua fábrica original. A categoria diagenética é separada em dois tipos, com base na obliteração ou não obliteração do sistema poroso da rocha.

4.2 Diagênese

Diagênese compreende todos os processos geológicos que ocorrem em um campo específico de condições físicas e químicas que atua sobre todos os tipos de materiais na superfície da crosta terrestre e nos primeiros milhares de metros de profundidade, englobando inclusive o intemperismo. Os processos diagenéticos têm forte influência sobre as rochas-reservatório, influenciando diretamente na sua porosidade e permeabilidade. A diagênese pode atuar de forma negativa, destruindo a porosidade primária, ou positiva, melhorando a qualidade de armazenamento e fluxo, preservando e/ou ampliando a porosidade original da rocha.

Alóctone		Autóctone		
Componentes não ligados organicamente durante a deposição		Componentes ligados organicamente durante a deposição		
>10% de grão >2mm		Por organismos que atuam como obstáculo	Por organismos que encrustam e unem	Por organismos que constroem um arcabouço rigido
Matriz suportado	Suportado por components com >2mm			
Floatstone	Rudstone	Bafflestone	Bindstone	Framestone

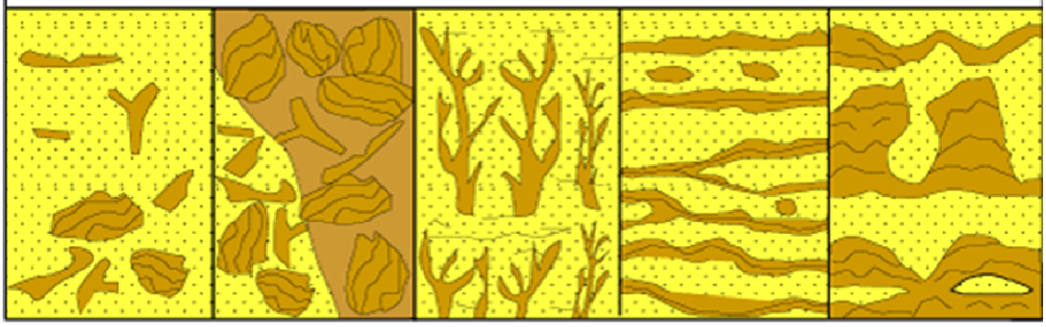

Figura 6. Classificação de carbonatos de recifes de Embry e Klovan (1971).

4.2.1 <u>Estágios</u>

De acordo com Choquette e Pray (1970), a diagênese em rochas carbonáticas pode ser dividida em três estágios distintos, os quais apresentam diferentes controles sobre a geração e destruição de porosidade:

A **eodiagênese** aplica-se ao intervalo entre o final da deposição/acumulação e a profundidade onde a diagênese não tem mais influência de processos e fluidos superficiais (denominada de soterramento efetivo). O limite superior é a superfície, sendo normalmente a interface deposicional, podendo ser uma superfície de não deposição temporária, ou mesmo uma superfície de erosão durante uma interrupção temporária da sedimentação. O limite inferior, com a mesodiagênese, tende a ser gradacional, uma vez que a intensidade do controle superficial diminui gradualmente com o soterramento. A espessura da zona eodiagenética pode ser de apenas alguns metros até poucos quilômetros de profundidade, em locais onde ocorre a circulação profunda de águas superficiais (CHOQUETTE; PRAY, 1970).

Deposicional	Suportado Pela Matriz (Silte e Argila)	<10% Grãos	*Calcimudstone*		
		>10% Grãos	*Wackestone*	Floatstone	Grãos >2mm
	Suportado pelos Grãos	com matriz	*Packstone*		
		sem matriz	*Grainstone*	Rudstone	
Biológico	Organismos *in situ*	Organismos encrustantes	*Boundstone*		
		Organismos que agem como barreias	*Bafflestone*		
		Organismos rígidos	*Framestone*		
Diagenético	Não Obliterado	Principal componente é o cimento	*Cementstone*		
		Muitos grãos micro-estilolíticos em contato	*Grainstone* condensado		
		Maior parte dos grãos microestilolíticos em contato	*Grainstone fitted*		
	Obliterado	Cristais >10 µm	Espatito	Cristais <10 µm	Microespatito

Figura 7. Classificação de carbonatos segundo Wright (1992).

A **mesodiagênese** constitui o estágio de evolução diagenética no qual inexiste influência direta da superfície. Os fluidos não são mais fluidos superficiais marinhos ou meteóricos, mas fluidos modificados pela dissolução e precipitação de minerais, com pressões e temperaturas que podem atingir até cerca de 2000 Kg/cm2 e 200ºC. O limite inferior da mesodiagênese corresponde à passagem ao metamorfismo, ou anquimetamorfismo, normalmente através de uma faixa gradacional.

A **telodiagênese** corresponde à fase onde rochas que estiveram soterradas e submetidas à mesodiagênese voltam a sofrer a influência de condições superficiais, particularmente da circulação de águas meteóricas, devido ao soerguimento, exposição e consequente erosão e geração de discordâncias, ou à circulação profunda de águas meteóricas por falhas e fraturas (CHOQUETTE; PRAY, 1970). Os processos telodiagenéticos agem, portanto, superimpostos aos constituintes mesodiagenéticos.

4.2.2 Processos

Os principais processos diagenéticos são sumarizados a seguir (*sensu* DUNOYER DE SEGONZAC, 1970; FÜCHTBAUER, 1974; WILSON; PITTMAN, 1977; LARSEN; CHILINGAR, 1979; BLATT, 1992; BOGGS, 1992; MORAES; DE ROS, 1992):

A **compactação** é gerada pelo soterramento, reduzindo o espaço poroso. Pode ser física, onde ocorre rearranjo, fraturamento ou esmagamento dos grãos, ou química, através da dissolução por pressão nos contatos intergranulares ou formando planos de dissolução conhecidos como estilolitos.

A **dissolução** é um processo que afeta tanto constituintes primários quanto diagenéticos. Pode ser congruente ou total, onde ocorre remoção completa dos materiais como íons em solução, ou incongruente, onde a dissolução incompleta pode deixar como “resíduo” uma nova fase mineral.

Reações de **hidratação/desidratação** correspondem a entrada/saída de água na estrutura de minerais.

Autigênese refere-se à formação de novos constituintes minerais após a deposição do sedimento, que podem precipitar em poros (cimentação) ou substituir constituintes pré-existentes.

Oxidação corresponde à remoção de elétrons dos elementos dos minerais sob influência da superfície, na presença de O_2 ou bactérias aeróbicas, comumente formando óxidos.

A **redução** é um processo químico governado pela adição de elétrons nos elementos, sob influência da matéria orgânica e de bactérias anaeróbicas. Como exemplo, a redução do Fe^{3+} para Fe^{2+}, levando à formação de pirita.

O termo **recristalização** é aplicado quando o novo constituinte autigênico tem a mesma composição mineral, mas forma e tamanho cristalino diferentes daqueles do constituinte anterior. Como exemplo, uma calcita microcristalina é substituída por calcita de cristalinidade grossa.

Neomorfismo (também denominada inversão ou estabilização) designa uma substituição por um constituinte de composição mineralógica e elementar similares, mas diferentes do anterior. Este termo é utilizado, por exemplo, para designar a substituição de aragonita por calcita não-magnesiana.

4.3 Porosidade em Rochas Carbonáticas

A porosidade em rochas carbonáticas apresenta grande diversidade de formas, com padrões bastante complexos (CHOQUETTE; PRAY, 1970; LUCIA, 1995; MOORE, 2001; MAZZULLO, 2004; AHR, 2008; MAZZULLO; WILHITE; WOOLSEY, 2009), resultado de processos atuantes na sua formação e evolução diagenética, controlados pela alta reatividade dos minerais que compõe essas rochas. A porosidade gerada durante a deposição/formação do sedimento é chamada primária, enquanto a porosidade criada durante a diagênese é referida com secundária. A caracterização da porosidade é de fundamental importância no estudo de reservatórios, uma vez que dá subsídio para a interpretação petrofísica e previsão da qualidade dos reservatórios através de modelos geológicos.

Diversas classificações de poros em rochas carbonáticas foram criadas, sendo a mais utilizada a proposta por Choquette e Pray (1970), que é baseada nas relações dos poros com a fábrica original da rocha. Esta classificação destaca-se por dar um significado evolutivo ao espaço poroso, permitindo o entendimento dos processos geradores da porosidade em função da fábrica deposicional e diagênese ao longo do tempo (Figura 8). Esses autores agruparam os tipos de poros em 3 classes. A primeira, seletiva quanto à fábrica, corresponde aos poros controlados diretamente pela fábrica da rocha, seja ela deposicional ou diagenética. Neste grupo são reconhecidos 7 tipos, onde destacam-se poros interpartícula, intercristalina, móldica e de crescimento. A segunda refere-se aos poros não condicionados pela fábrica, tais como *vugs*, canais e fraturas. A terceira classe agrupa tipos de poros reconhecidos por sua gênese, tais como brecha, escavação, perfuração e contração, seletivos quanto à fábrica o não.

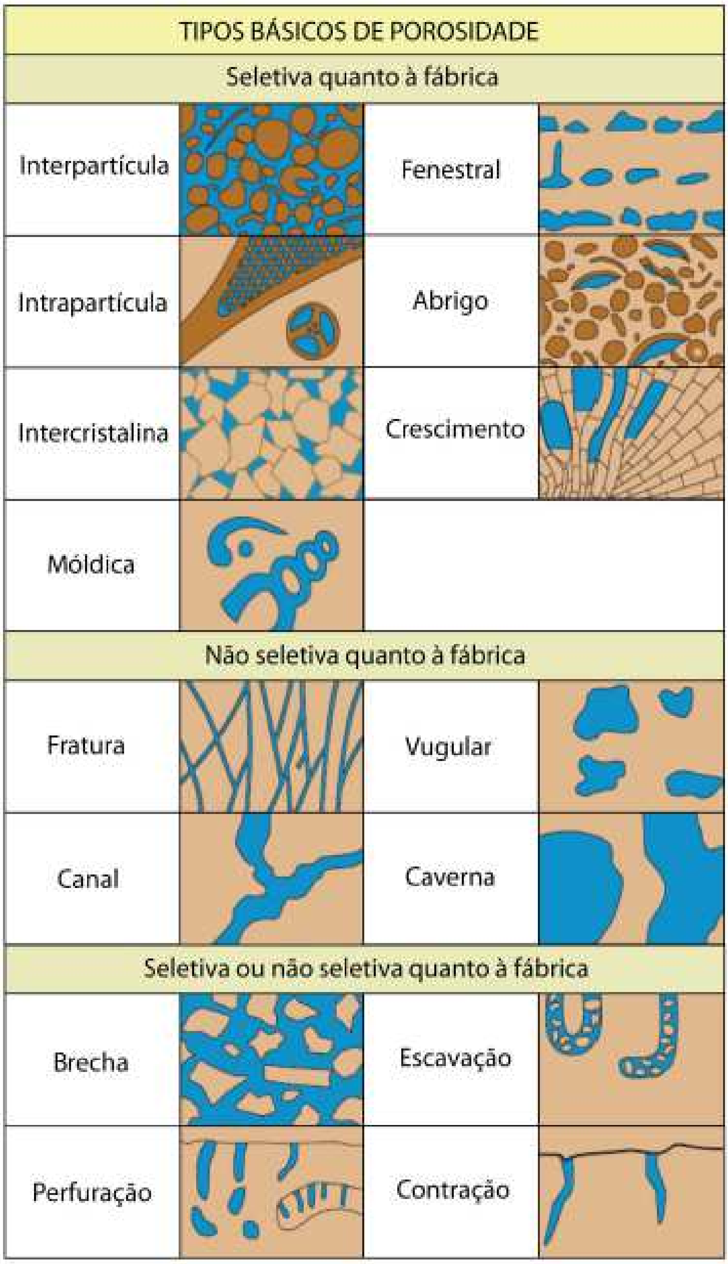

Figura 8. Classificação proposta por Choquette e Pray (1970) para a descrição da porosidade em carbonatos.

4.4 Carbonatos Lacustres

Define-se como lago uma depressão continental coberta por uma coluna de água independente da profundidade. Lagos podem ser classificados como perenes, quando a coluna d'água é permanente ou efêmeros, quando os sedimentos são expostos sazonalmente (GIERLOWSKI-KORDESCH, 2009). Lagos são formados em diferentes tipos de bacias (PLATT; WRIGHT, 1991), podendo ser classificados segundo o ambiente tectônico onde foram formados: divergência litosférica, convergência crustal e regimes transformantes (COHEN, 2003).

A deposição de carbonatos lacustres possui diversos controles, dos quais destacam-se três fatores principais: hidrológico (que inclui a entrada e saída de águas superficiais, precipitação pluviométrica e fluxo de águas subterrâneas), entrada de sedimentos e variações de temperatura (PLATT; WRIGHT, 1991; TUCKER; WRIGHT, 1991). Embora o clima e a tectônica possuam forte influência sobre esses três fatores, distingui-los em ambientes lacustres, muitas vezes não é uma tarefa fácil (GIERLOWSKI-KORDESCH, 2009).

A deposição de carbonatos em ambientes lacustres tem sido associada a quatro principais processos. O primeiro corresponde à concentração por evaporação, levando à precipitação em ambientes alcalinos com água doce e em lagos salinos, sob controle químico. Altas temperaturas contribuem para a evaporação e precipitação de carbonatos (MÜLLER; IRION; FÖRSTNER, 1972; GIVEN; WILKINSON, 1985; JONES; RENAUT, 1994; WRIGHT, 2012). Como segundo processo destaca-se a biomediação, onde carbonatos precipitam sob influência de condições criadas pela presença de atividade microbial em lagos (TALBOT, 1990; LENG; MARSHALL, 2004; VASCONCELOS *et al.*, 2006). Por fim, um terceiro processo corresponde ao transporte de partículas carbonáticas para ambientes lacustres, sejam elas formadas por processos intrabaciais ou extrabaciais, como a erosão de rochas carbonáticas preexistentes. Essas partículas podem ser reconhecidas por características mineralógicas, isotópicas e texturais (KELTS; TALBOT, 1990; JIANG *et al.*, 2007).

Sucessões lacustres comumente mostram diversas espécies de plantas e vegetais preservadas em seu registro. A seguir são destacados os principais tipos de registros de fauna e flora encontrados nesses ambientes.

Carófitas são algas verdes comumente encontrados em ambientes não marinhos desde o Paleozóico, presentes em ambientes tanto salinos quanto de água doce (SANZ *et al.*, 1988; GARCÍA, 1994; DÉTRICHÉ *et al.*, 2009). Carófitas fossilizam facilmente através da biomineralização de carapaças de órgãos reprodutivos e caules por carbonato de cálcio (ANADÓN; UTRILLA; VÁZQUEZ, 2000).

Ostracodes são microcrustáceos que vivem em conchas bivalves, vivendo tanto de forma bentônica quanto pelágica (DE DECKKER, 2002; HORNE; COHEN; MARTENS, 2002). São encontrados em uma grande variedade de ambientes. Seu alto potencial de preservação faz com que sejam excelentes indicadores paleoambientais (FROGLEY; GRIFFITHS; MARTENS, 2002). Ostracodes podem formar bioacumulações, com conchas articuladas acumuladas *in situ*, e têm grande resistência à diagênese (BERTANI; CAROZZI, 1985a, 1985b; CARVALHO et al., 2000).

Moluscos são bem representados no registro geológico durante o Fanerozóico, na forma de bivalves e gastrópodes (COHEN, 2003; PARK; GIERLOWSKI-KORDESCH, 2007). Podem viver em ambientes litorâneos, sublitorâneos ou ambientes profundos, desde que sejam bem oxigenados. Podem formar espessas acumulações, sendo encontrados *in situ* ou na forma de coquinas, quando retrabalhados (BAUMGARTEN, 1985; BAUMGARTEN *et al.*, 1988; HORSCHUTZ; SCUTA, 1992; CARVALHO *et al.*, 2000; RANGEL; CARMINATTI, 2000; ALTENHOFEN, 2013; THOMPSON; STILWELL; HALL, 2015)

4.5 Crostas Microbiais e Abióticas

O termo estromatolito foi inicialmente proposto por (KALKOWSKY *apud* RIDING, 2008) para definir depósitos carbonáticos não-esqueletais com estrutura laminada. Kalkowsky enfatizou que as estruturas laminadas deviam possuir origem orgânica, e que os organismos envolvidos deveriam ser simples e similares a plantas, de onde conclui-se que este autor considerou os estromatolitos essencialmente como depósitos microbiais laminados (RIDING, 1999). Baseado na morfologia de depósitos análogos atuais, diversos autores têm considerados os estromatolitos essencialmente como esteiras microbiais litificadas. No entanto, similaridades morfológicas entre estromatolitos e uma variedade de outros depósitos geológicos tem confundido seu reconhecimento e gerado um debate sobre como o termo estromatolito deveria ser usado (SEMIKHATOV *et al.*, 1979; GROTZINGER; KNOLL, 1995, 1999; RIDING, 1999, 2008; GROTZINGER, 2000). Embora o conceito original de estromatolito não tenha sido claramente associada à uma gênese orgânica, ele acabou sendo aplicado à qualquer produto de litificação de qualquer tipo de depósito microbial, com ou sem estrutura laminada, de acordo com a definição de Walter (1976): "estromatolitos são estruturas organossedimentares produzidas por captura, agregação e/ou precipitação de sedimentos, como resultado do crescimento e atividade metabólica de microorganismos, principalmente cianofíceas". Mais recentemente, Grotzinger e Knoll (1999) propuseram a adoção da definição não-genética recomendada por Semikhatov *et al.* (1979): "...estruturas de crescimento sedimentar laminadas, litificadas, aderidas, acrecionais a partir de um ponto ou superfície limitada de iniciação." Uma vez que muitos depósitos geológicos possuem morfologias que enquadram-se na definição original mas não possuem gênese microbial, parece prudente a utilização de uma definição que não possua um caráter genético, mas simplesmente descritivo.

Microbialitos ou crostas microbiais são depósitos organossedimentares formados pela interação entre comunidades microbiais bentônicas (CMB) e sedimentos detríticos ou químicos (BURNE; MOORE, 1987; RIDING, 2000). Os processos reconhecidos para a formação dos depósitos microbiais incluem a captura de sedimentos, que podem ser carbonáticos ou não (ANDREWS; BRASIER, 2005; SCHIEBER *et al.*, 2007; TAKASHIMA; KANO, 2008; DRUSCHKE *et al.*, 2009; NOFFKE, 2010; MEISTER *et al.*, 2011; NOFFKE; AWRAMIK, 2013) e a precipitação mineral, mais comumente de carbonato, e mais raramente de outros minerais como sílica e fosfato (SANCHES-NAVAS; MARTIN-ALGARRA; NIETO, 1998). A precipitação é provocada diretamente pela atividade metabólica das CMB através de processos como fotossíntese de cianobactérias ou quimiossíntese por oxidação de metano. Por fim, também é reconhecido como depósito microbial aquele cuja precipitação inorgânica ocorre sobre um substrato orgânico formado pelas CMB, comumente na forma de esteiras, tapetes e filmes (RIDING, 2000). Este último processo pode ocorrer pela decomposição do substrato orgânico formado por CMB sintetizantes promovida por outros micróbios, como bactérias de redução de sulfato (VASCONCELOS *et al.*, 2006; SPADAFORA *et al.*, 2010), cuja ação promove a substituição deste substrato por carbonatos e/ou outros minerais; ou por incrustação de bactérias ou filamentos (CHAFETZ; FOLK, 1984); ou ainda filmes orgânicos formados pelas CMB com minerais precipitados por reações que podem ser puramente inorgânicas, tais como concentração por evaporação ou perda de CO_2, ou mediadas pelas CMB, como por exemplo mudanças de pH por oxidação da matéria orgânica e redução de nitrato.

A precipitação de carbonatos através de processos puramente químicos na forma de crostas tem sido reconhecida em grande diversidade de ambientes geológicos, como cimentos marinhos, travertinos, lagos alcalinos, espeleotemas, fumarolas, etc (FOLK; CHAFETZ; TIEZZI, 1985; PERYT *et al.*, 1990; SAMI; JAMES, 1996; FOUKE, 2011; PERRI; TUCKER; MAWSON, 2013). O reconhecimento e a diferenciação de processos bióticos e abióticos necessita de um enfoque baseado nos processos, orientado para o reconhecimento das texturas originais e de substituição e recristalização. Só então as texturas de laminação originais podem ser usadas na dedução dos prováveis mecanismos de acreção (GROTZINGER; KNOLL, 1999). Estes autores afirmam que processos de formação totalmente diferentes, como a captura microbial de sedimentos ou a precipitação por cianobactérias, e a precipitação inorgânica (abiótica) direta podem gerar estruturas com forma similar, mas que os processos genéticos são reconhecíveis através de diferentes texturas e microfábricas.

Os critérios para a identificação da origem microbial ou abiótica dos estromatolitos e depósitos similares com base em suas texturas e fábricas foram organizados e formalizados por (RIDING, 2008), que identificou texturas radiais, laminações isópacas e fábricas microdigitadas como as mais comumente produzidas por precipitação abiótica. Depósitos estromatolíticos e similares foram caracterizados como produto da combinação de três componentes principais: precipitados abióticos, depósitos microbiais litificados e grãos "alóctones", incluindo grãos carbonáticos e não-carbonáticos. A aplicação de critérios petrográficos coerentes para identificar a origem dos constituintes dos estromatolitos e depósitos similares permitiu a identificação dos estromatolitos Arqueanos e Paleoproterozóicos como dominantemente químicos, abióticos (GROTZINGER, 1990, 2000; GROTZINGER; KNOLL, 1995, 1999; POPE; GROTZINGER; SCHREIBER, 2000).

4.6 Argilominerais Magnesianos

A precipitação e acumulação de argilominerais magnesianos em ambientes lacustres alcalinos tem sido descrita em diferentes idades, contextos geotectônicos (TETTENHORST; MOORE JR, 1978; YURETICH, 1986, 2002; DARRAGI; TARDY, 1987; POZO; CASAS, 1999; FURQUIM *et al.*, 2008; WASSON *et al.*, 2012; BRISTOW *et al.*, 2012) e planetários (MCLENNAN *et al.*, 2013). Ocorrências significativamente abundantes de argilominerais magnesianos ao longo do Grupo Lagoa Feia foram relatadas, tanto no Andar Alagoas quanto no Jiquiá (BAUMGARTEN, 1985; REHIM *et al.*, 1986; DIAS; OLIVEIRA; VIEIRA, 1988; HORSCHUTZ; SCUTA, 1992; WRIGHT, 2012, 2013; ALTENHOFEN, 2013; MADRUCCI *et al.*, 2013; TOSCA; WRIGHT, 2014; WRIGHT; BARNETT, 2014).

Argilominerais trioctaédricos magnesianos são frequentemente encontrados em lagos alcalinos e depósitos evaporíticos, como produto da precipitação singenética diretamente das águas superficiais. Comumente ocorrem na forma de sepiolita, palygorskita, estevensita, saponita, kerolita e talco (DARRAGI; TARDY, 1987). Sepiolita e palygorskita são argilominerais pobres em alumínio com hábitos fibrosos, comumente encontrados em lagos salinos e alcalinos (BRISTOW *et al.*, 2012). Estes minerais diferem de outros argilominerais por conterem folhas descontínuas de octaedros e tetraedros, organizadas em fitas 2:1, onde cada fita é ligada por uma inversão de tetraedros de sílica ao logo das ligações Si-O-Si (GALAN, 1996). Durante muito tempo acreditou-se que sepiolita fosse o argilomineral mais comum formado em ambientes salinos, principalmente pelo fato de que estevensita era confundida com outras esmectitas (DARRAGI; TARDY, 1987).

Estevensita, talco e kerolita são argilominerais trioctaédricos com estrutura e composição muito semelhantes (Figura 9), porém formados em condições geoquímicas distintas com campos de estabilidade específicos para cada espécie. Estevensitas possuem o espaço intercamadas ocupado por água e cátions, como por exemplo sódio, potássio, cálcio e magnésio. A kerolita, por sua vez, constitui uma forma hidratada de talco, onde a deficiência de cargas é suprida apenas por moléculas de H_2O e OH^-. Já o talco possui exclusivamente camadas de duas folhas tetraédricas e uma folha octaédrica, frouxamente conectadas por ligações van der Waals (TOSCA; MASTERSON, 2014). Saponita é um argilomineral trioctaédrico do grupo das esmectitas. É caracterizada pela maior presença de Al substituindo Si nos tetraedros do que ocorre na estevensita, ocorrendo muitas vezes associada à outros argilominerais magnesianos, tal como a estevensita (CASAL, 1997).

A autigênese destes minerais ocorre em condições físico-químicas distintas, controladas pelo pH, salinidade e razão Mg/Si (JONES, 1986; JONES; GALAN, 1988; STOESSELL, 1988; BIRSOY, 2002; JONES; CONKO, 2011; TOSCA; WRIGHT, 2014; TOSCA; MASTERSON, 2014). Estudos experimentais mostram que em baixas salinidades, a kerolita predomina em condições de alta razão Mg/Si e alto pH, enquanto que a sepiolita é formada sob pH baixo e baixa razão Mg/Si. Em alta salinidade e alta razão Mg/Si, a estevensita é dominante em alto pH (acima de 9,4), enquanto que a kerolita é a fase predominante em pH mais baixo (TOSCA; WRIGHT, 2014; TOSCA; MASTERSON, 2014). A alta área específica aliada ao tênue campo de estabilidade faz com que estes argilominerais sejam fortemente solúveis, sendo facilmente dissolvidos ou substituídos por fases minerais mais estáveis (TOSCA; WRIGHT, 2014; TOSCA; MASTERSON, 2014). A precipitação destes minerais ocorre comumente na forma de géis hidratados (CASAL, 1997; YURETICH, 2002; WRIGHT; BARNETT, 2014). São encontrados no registro geológico como laminações e também na forma de oóides e pelóides, frequentemente substituídos, dissolvidos e deformados.

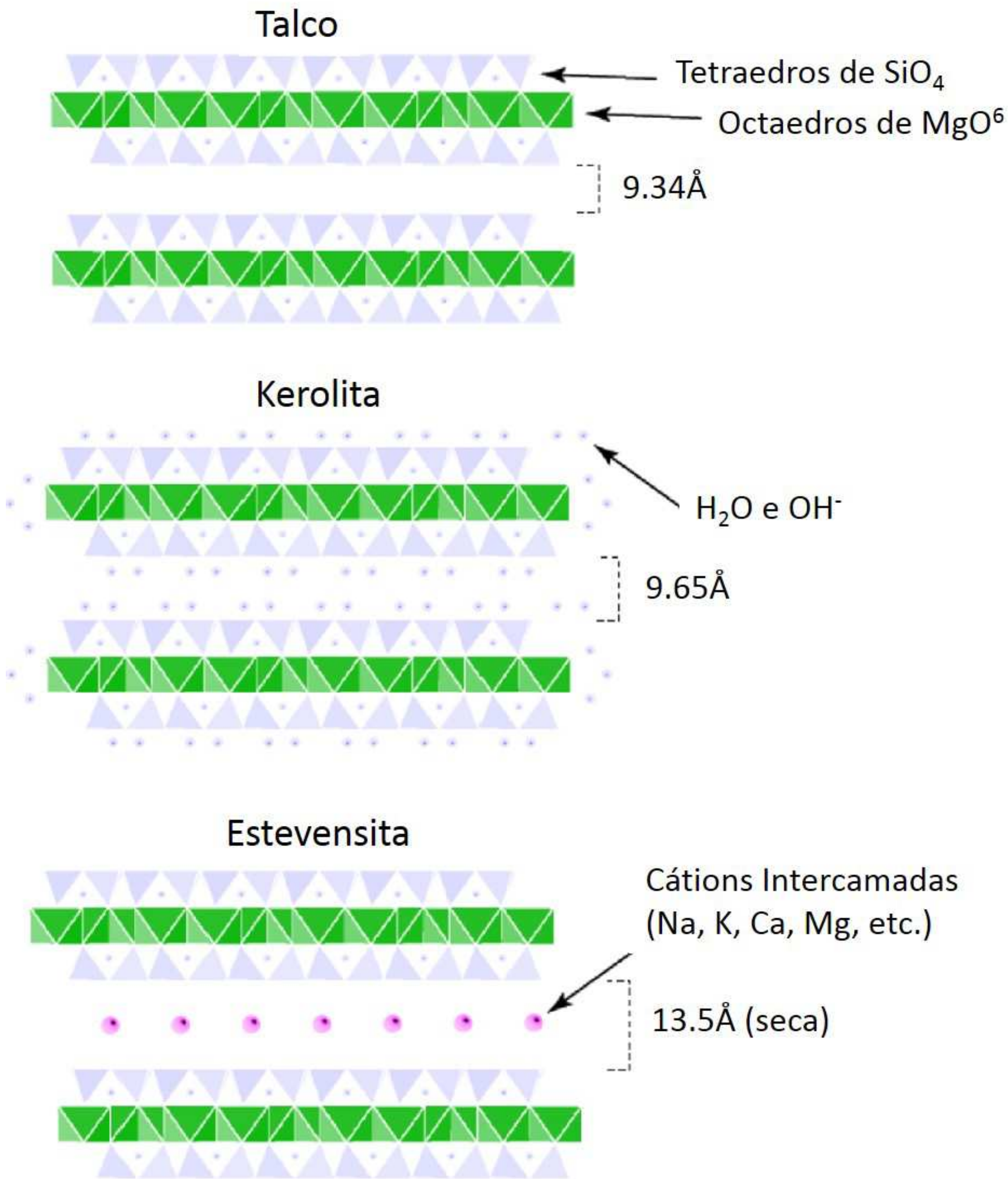

Figura 9. Estrutura dos argilominerais talco, kerolita e estevensita (modificado de TOSCA e WRIGHT (2014)).

5 Metodologia

5.1 Petrografia

Foi realizada a caracterização petrográfica em microscopia ótica de 139 lâminas delgadas, selecionadas de testemunhos e amostras laterais de 5 poços. As lâminas foram impregnadas com resina epoxy azul para o realce da porosidade e tingidas com solução ácida diluída de alizarina vermelha e ferrocianeto de potássio para a melhor caracterização dos minerais carbonáticos (DICKSON, 1965). Foi feita a identificação dos constituintes primários, bem como das fases diagenéticas, além da caracterização dos tipos de poros, suas localizações e relações paragenéticas, com uso do software *Petroledge*® (DE ROS *et al.*, 2007). Além disso, todas as amostras com petrofísica básica, quando possuíam amostra macro e/ou lâmina, foram classificadas de acordo com o tipo de petrofácies e porosidade presentes.

Foram analisadas 13 lâminas através de catodoluminescência (CL) em um equipamento *Cambridge Image Technology Ltd*. (CITL) modelo CL8200 MK5-2, operando com corrente de 300 μA e 17 kV, acoplado à um microscópio ótico, com o objetivo de caracterizar o crescimento e recristalização das fases minerais autigênicas.

Além da petrografia por luz polarizada, foi utilizada a técnica de epifluorescência de luz ultravioleta, emitida por LED (*Light Emitting Diode*). Os fótons absorvidos por algumas moléculas excitam seus elétrons à um nível de energia mais alto. Conforme os elétrons voltam a seu estado normal, a energia absorvida é emitida como fótons com comprimento de onda na faixa visível. Foram analisadas 91 lâminas por epifluorescência.

5.2 Microscopia eletrônica de varredura (MEV)

Foram selecionadas 10 lâminas e 2 fragmentos de rocha para a análise por MEV. Os fragmentos foram recobertos por uma delgada camada de carbono, aderidas a um suporte condutor de latão e analisadas em um microscópio eletrônico de varredura ZEISS EVO LS15, com detector de elétrons secundários, operando em alto vácuo a 20 kV com distância de trabalho de 12mm. As lâminas foram recobertas por uma delgada camada de carbono, aderidas a um suporte condutor de alumínio. Duas lâminas foram analisadas em um equipamento JEOL JSM 6490LV e as demais em um MEV ZEISS EVO LS15, ambos com detector de elétrons retroespalhados (*backscattered electrons*; BSE), operando em alto vácuo a 20 kV com distância de trabalho de 10mm. As imagens apresentam tons de cinza mais claros para minerais de peso atômico médio maior, e tons de cinza mais escuros para minerais de peso atômico médio menor. Simultaneamente, microanálises elementares por espectrometria de energia dispersada (EDS) foram obtidas através de equipamentos OXFORD Inca, acoplados a ambos os MEVs.

5.3 Análises de Difração de Raios-X

As espécies de argilominerais foram analisadas em 2 amostras de calha e em 3 amostras de plugues laterais. As amostras foram imersas em água destilada e desagregadas por ultra-som de ponta. Em seguida, retirou-se por centrifugação o material contido na fração granulométrica inferior a 2µm (calha e plugue) e 20µm (calha) da suspensão estável obtida. Este material foi concentrado por centrifugação e orientado por esfregaço. As lâminas foram tratadas com etilenoglicol e aquecidas a 490°C (plugue) e 540°C (calha) a fim de se identificar os argilominerais.

As análises das amostras laterais foram realizadas no Centro de Pesquisas e Desenvolvimento Leopoldo A. Miguez de Mello (CENPES) da Petrobras, em um difratômetro RIGAKU D/MAX - 2200/PC, utilizando-se radiação k-alfa do cobre nas condições de 40 kV e 40 mA de corrente de filamento e fendas de 2mm; 2mm; 0,3mm e 0,6mm. A velocidade de varredura do goniômetro foi de 6 graus por minuto.

As amostras de calha foram analisadas no Laboratório de Difratometria de Raios-X do Centro de Petrologia e Geoquímica (CPGq) da UFRGS com um difratômetro SIEMENS D5000 com goniômetro theta-theta. Para a avaliação da presença da argila, varreu-se o intervalo angular de 2° até 19° 2θ. O passo (*step size*) utilizado foi de 0,02° com fendas de 1° (divergente e de antiespalhamento) e de 2 mm no monocromador de grafite.

5.4 Ensaios de Petrofísica Convencional

Foram realizados ensaios petrofísicos em 780 plugues e amostras laterais obtidos de testemunhos e amostras laterais, amostrados nos cinco poços estudados. As amostras foram limpas com solventes polares (tolueno e clorofórmio), para a remoção de óleo, e com metanol, para a remoção de sais solúveis. Posteriormente os plugues e amostras laterais foram secos em estufa, medidos com paquímetro e pesados.

5.4.1 Densidade

O primeiro estágio para o cálculo da densidade reside na obtenção do volume da amostra de rocha através da determinação do volume da câmara vazia (VCV) e do volume da câmara cheia (VCC) no equipamento denominado porosímetro. A amostra é inserida na câmara de leitura, conectada à linha com gás nitrogênio, e após a estabilização da pressão é realizada a leitura com a câmara cheia. O VCV é obtido após a retirada da amostra da câmara de leitura, depois do tempo de estabilização da pressão dentro da câmara. O volume de sólidos é obtido através da fórmula abaixo:

$$VS = VCV - VCC$$

Onde:

VS = Volume Sólidos;

VCV = Volume da Câmara Vazia;

VCC = Volume da Câmara Cheia.

A densidade dos sólidos é calculada segundo a equação abaixo:

Densidade = massa/volume de sólidos

5.4.2 Porosidade

A medida do volume poroso da amostra é realizada sob pressão de confinamento. Para isso, deve-se acondicionar a amostra totalmente limpa, seca e fria na célula de confinamento do porosímetro. Passado o tempo mínimo de confinamento da amostra (cerca de 1h), conecta-se as linhas do porosímetro permitindo que amostra seja rapidamente preenchida por nitrogênio. Após o equilíbrio ser atingido, é lida a quantidade de gás que permeou na amostra, que é proporcional ao volume poroso da rocha. A porosidade efetiva é, então, obtida a partir da seguinte equação:

$$Phi = \frac{VP}{VP+VS} x100$$

Onde:

Phi = porosidade efetiva (%);

VS = volume de sólidos (cm3);

VP = volume poroso (cm3).

5.4.3 Permeabilidade

A amostra limpa e seca é inserida no permeâmetro, onde reside por 1 hora até que seja atingida a pressão de confinamento adequada. Então o permeâmetro é alimentado com nitrogênio, com pressão de entrada na amostra entre 30 e 40 psi. A seguir, a válvula micrométrica é aberta de forma que o gás flua através da amostra, sendo medida a vazão através de um sensor de fluxo. São medidas as pressões na linha do medidor de vazão, na face de entrada da amostra e a pressão de saída. A permeabilidade aparente então é calculada pela equação de Darcy modificada para fluxo laminar de gases (CUNNINGHAM; WILLIAMS, 1980):

$$K_{AP} = \frac{Q_B \times \mu_g \times L \times P_b \times 2 \times 10^3}{(P_1^2 - P_2^2) \times A}$$

Onde:

K_{AP} = Permeabilidade aparente ao gás;

Q_B = Vazão volumétrica do gás à pressão da linha do medidor de vazão;

P_1 = Pressão na face de entrada da amostra;

P_2 = Pressão de saída;

P_b = Pressão da linha do medidor de vazão (1atm);

μ_g = Viscosidade do gás;

L = Comprimento da amostra;

A = Área da seção reta da amostra.

A permeabilidade absoluta é calculada através da correção da permeabilidade aparente pelo efeito Klinkenberg (KLINKENBERG, 1941), usando seguinte equação:

$$K_{AP} = K_{ABS}\left(1 + \frac{b}{Pm}\right)$$

Esta equação relaciona a permeabilidade medida, também chamada de permeabilidade aparente (K_{AP}), com a permeabilidade absoluta (K_{ABS}) em função da pressão média do gás fluindo através da amostra (Pm) e do fator de escorregamento (b). O fator de escorregamento “b” é função da permeabilidade e da rocha em questão, obtido de relação empírica determinada por McMahon (1949).

5.5 Análise de Injeção de Mercúrio

Foram realizadas análises de injeção de mercúrio em 49 amostras, com o objetivo de obter-se a distribuição do raio equivalente das gargantas de poros. O ensaio de injeção de mercúrio determina a distribuição de volume de poros acessíveis a determinadas pressões, equivalentes a determinados diâmetros de gargantas de poros. O equipamento utilizado foi um *Micromeritics AutoPore* IV 9500 com quatro portas de baixa pressão (para pontos da análise abaixo da pressão atmosférica) e duas portas de alta pressão para pressões até 60000psi. A curva de pressão capilar é obtida colocando-se na abscissa de um gráfico a saturação de mercúrio e, na ordenada, a pressão de injeção. A saturação é obtida dividindo-se o volume de mercúrio invadido até determinada pressão pelo volume poroso da amostra. A curva de distribuição de diâmetros de gargantas de poros é obtida empregando-se a equação de Laplace:

$$PC = \frac{2 * \sigma * \cos\theta}{r} * 1450.377 \qquad \therefore \qquad D = 2 * \left(\frac{2 * \sigma * \cos\theta}{PC} \right) * 1450.377$$

PC = Pressão capilar [psi];

σ = Tensão superficial [dina/cm]; para o Hg, 480 dina/cm;

θ = Ângulo de contato [graus]; para o sistema Hg/ar/sólido, 140°;

r = Raio de garganta de poros [angstrons];

D = Diâmetro de garganta de poros [angstrons].

6 Síntese dos Resultados e Considerações Finais

1) Os principais constituintes primários dos reservatórios carbonáticos da Formação Coqueiros são bioclastos de bivalves, com quantidades subordinadas de gastrópodes e ostracodes. Ocasionalmente são encontrados intraclastos lamosos, e grãos siliciclásticos e vulcanoclásticos. Os bioclastos de bivalves formam comumente *rudstones* com diferentes graus de seleção e orientação da fábrica, enquanto que os ostracodes formam *grainstones*. Oóides de argilas magnesianas formam arenitos bem selecionados, provavelmente *in situ*, ou arenitos mal selecionados, misturados com pelóides retrabalhados de diferentes dimensões.

2) Os reservatórios da Formação Macabu são compostos dominantemente por precipitados químicos abióticos, formados por agregados cristalinos de calcita fascicular, associados a argilas singenéticas magnesianas na forma de laminações e pelóides/oóides. Os agregados fasciculares de calcita tendem a coalescer e formar crostas, que podem conter porosidade primária de crescimento do arcabouço (*growth framework*) entre os agregados cristalinos. Onde pouco desenvolvidos, os agregados fasciculares podem ocorrer intercalados com matriz argilosa, isolados ou parcialmente coalescentes, com reduzida porosidade primária. *Grainstones* e *rudstones* de intraclastos fragmentados das crostas de calcita são comuns, contendo quantidades subordinadas de oóides e pelóides argilosos. As rochas intraclásticas frequentemente apresentam matriz singenética argilosa. Laminitos de provável

origem microbial e siltitos híbridos, formados por lama e areia siliciclásticas, matriz argilosa singenética e fragmentos de carbonatos, são comuns.

3) A calcita é o principal constituinte diagenético dos reservatórios bioclásticos da Formação Coqueiros, na forma de cimento drusiforme preenchendo poros móldicos e como franjas cobrindo os bioclastos. Dolomita ocorre de forma mais restrita, preenchendo poros móldicos ou interpartícula remanescentes, ou substituindo completamente os constituintes primários e diagenéticos de forma pervasiva. Sílica e traços de barita e celestina ocorrem localmente, cimentando e/ou substituindo constituintes primários e diagenéticos. Svanbergita/goyazita ocorrem substituindo intraclastos lamosos. Arenitos ooidais “estevensíticos” foram normalmente substituídos totalmente por sílica e/ou dolomita associadas no mesmo nível ou segundo camadas intercaladas.

4) A calcita é igualmente o constituinte diagenético mais comum da Formação Macabu, entretanto na forma de esferulitos que ocorrem substituindo e deslocando laminações argilosas singenéticas e lama terrígena. A dolomita é um constituinte muito comum, ocorrendo em diversos hábitos, substituindo argilas magnesianas e cimentando poros primários e secundários. Silicificação é comum, substituindo as argilas ou a calcita de forma localizada ou pervasivamente em níveis milimétricos a centimétricos, comumente associada com dolomita. Pirita está presente como traços, normalmente substituindo diversos constituintes. Fluorita, barita, celestina, anidrita, apatita, svanbergita/goyazita são ocasionalmente observados como substituição ou como cimentos nos poros. Argilominerais interestratificados I/S são observados com frequência, como traços, geralmente substituindo argilas e micas detríticas.

5) A maior parte das alterações diagenéticas dos reservatórios bioclásticos ocorreu durante a eodiagênese, relacionada principalmente ao neomorfismo e dissolução da aragonita e à precipitação de cimento de calcita. A relação entre dissolução e neomorfismo dos bioclastos aragoníticos de bivalves foi controlada provavelmente pelas condições ambientais durante a eodiagênese. Ambientes com alta circulação de fluidos tenderiam a dissolver bioclastos, favorecendo a formação de porosidade

móldica, enquanto que ambientes mais estagnados estariam relacionados à preservação dos bioclastos e predomínio de neomorfismo da aragonita, com preservação da porosidade interpartícula. Os principais processos mesodiagenéticos foram a precipitação de dolomita em sela, cimentação por calcita blocosa e formação de poros vugulares por dissolução.

6) A diagênese dos arenitos ooidais ocorreu, da mesma forma, dominantemente durante a eodiagênese, com cimentação por sílica ou dolomita anterior à compactação. A intensa silicificação observada em arenitos ooidais ocorrentes no topo da seção rifte está provavelmente relacionada à sua exposição ao longo da superfície de discordância regional, que os teria submetido à um ambiente freático meteórico, com intensa dissolução dos argilominerais magnesianos e precipitação de sílica e dolomita, formando dolocretes e silcretes.

7) A diagênese da Formação Macabu foi fortemente influenciada pela reatividade dos argilominerais magnesianos e sua interação com agregados esferulíticos e com crostas de calcita fascicular durante a eodiagênese. As argilas foram precocemente dissolvidas ou substituídas por calcita, dolomita e sílica.

8) Os esferulitos de calcita foram precipitados deslocando e substituindo laminações de argilas magnesianas singenéticas e lama terrígena, crescendo como concreções no interior dos sedimentos inconsolidados.

9) Aparentemente o desenvolvimento das crostas de calcita fascicular ocorreu concomitante ao crescimento dos esferulitos, controlado por variações nas condições químicas lacustres. Ao menos parte das crostas de calcita fascicular ocorre como substituição de matriz argilosa, o que é evidenciado pela ocorrência comum de restos de laminações argilosas truncadas e/ou deformadas entre os agregados cristalinos.

10) A ocorrência de depósitos microbiais é limitada, correspondendo a níveis e lentes em laminitos com porosidade restrita, que não constituem reservatórios.

11) Alterações hidrotermais de diferentes depósitos de ambas formações ocorrem

de forma restrita, associados à intenso faturamento, dolomitização e silicificação. As rochas afetadas por alteração hidrotermal comumente contém barita, fluorita, caulim, pirita e outros sulfetos.

12) O estudo petrográfico e petrofísico permitiu o estabelecimento de diversos tipos de petrofácies de reservatório na Formação Coqueiros, com base na composição e textura primárias e no tipo de porosidade predominante: *rudstones* com porosidade interpartícula, *rudstones* com porosidade móldica/intrapartícula, *rudstones* com porosidade vugular, *grainstones* com porosidade móldica, arenitos ooidais “estevensíticos”, espatitos e doloespatitos.

13) *Rudstones* bioclásticos com porosidade interpartícula apresentam porosidades intermediárias e boas permeabilidades, graças à boa conectividade dos poros. Por outro lado, *rudstones* e *grainstones* com poros móldicos apresentam a melhor porosidade entre as rochas bioclásticas, porém com baixas permeabilidades, devido à pobre conectividade do sistema poroso. A associação de poros vugulares com outros tipos de poros tende a amplificar a permeabilidade, pela conexão dos poros e alargamento das gargantas de poros.

14) Arenitos ooidais argilosos possuem múltiplos tipos de poros, incluindo poros móldicos/intrapartícula, poros intercristalinos associados à dolomitização, e poros microcristalinos e vugulares associados à silicificação. Como resultado, observa-se grande heterogeneidade petrofísica. Regiões onde ocorre predomínio de dolomita apresentam as melhores condições de permeabilidade, enquanto que intervalos com porosidade móldica e microcristalina mostram boa porosidade, porém com baixas permeabilidades.

15) A dolomitização da Formação Coqueiros tendeu a ser mimética, com formação de doloespatitos, onde parte da porosidade móldica e interpartícula foi preservada, promovendo a conexão da porosidade anterior pela porosidade intercristalina, contribuindo para o aumento da permeabilidade. Tais doloespatitos constituem a petrofácies com melhor qualidade dentre os reservatórios da Formação Coqueiros.

16) As crostas coalescentes de calcita representam os melhores reservatórios

primários da Formação Macabu, com boa porosidade primária entre os agregados fasciculares, com gargantas largas que resultam em boa permeabilidade. Processos de cimentação por dolomita comumente reduzem a porosidade, enquanto que o alargamento dos poros por dissolução de materiais argilosos intersticiais e dos agregados ocorre como agente amplificador da permeabilidade. Crostas não-coalescentes com abundantes argilas e sedimentos detríticos entre os agregados fasciculares somente constituem reservatórios quando ocorre dissolução de matriz, gerando alguma porosidade intercristalina, porém com baixas permeabilidades.

17) As petrofácies intraclásticas normalmente ocorrem compactadas e cimentadas por calcita ou possuem matriz argilosa. Onde possuem porosidade primária preservada, ou de dissolução de matriz, podem apresentar boas porosidades e permeabilidades.

18) Os doloespatitos da Formação Macabu que foram formados por substituição precoce, influenciada pelo ambiente alcalino lacustre com alta disponibilidade de magnésio, sofreram total destruição da porosidade original. Doloespatitos com formação tardia, associados a faturamento e minerais com provável origem hidrotermal, frequentemente associados à silicificação, apresentam altas porosidades e permeabilidades.

19) A presença de áreas e intervalos com altas permeabilidades é de especial interesse para a indústria do petróleo, face ao fato de apresentarem alto impacto na migração de hidrocarbonetos e água. Trabalhos futuros deveriam focar no entendimento da influência e controles dos processos hidrotermais que modificaram as rochas do Grupo Lagoa Feia.

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DEPOSITIONAL AND DIAGENETIC CONTROLS ON THE QUALITY OF LACUSTRINE PRE-SALT CARBONATE RESERVOIRS FROM NORTHERN CAMPOS BASIN, OFFSHORE BRAZIL

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ABSTRACT

Lacustrine carbonate reservoirs from the Lagoa Feia Group, rift section of Campos Basin, offshore eastern Brazil, have sustained for decades a significant production from shallow water oil fields. The discovery of giant accumulations in the rift and sag (pre-salt) section of the adjacent Santos Basin has reactivated the exploration for equivalent reservoirs in the Campos Basin and in other marginal basins. A systematic petrographic and petrophysical study has been performed on the rift Coqueiros Formation and the sag Macabu Formation carbonates from the Lagoa Feia Group in northern Campos Basin, in order to characterize the main controls on the origin and evolution of those unconventional reservoirs and their pore systems. The main types of reservoir petrofacies recognized were grainstones and bioclastic rudstones, magnesian clay ooidal arenites and dolostones from the Coqueiros Formation; coalescent and non-coalescent crusts of fascicular calcite, intraclastic rudstones and grainstones, and dolostones from the Macabu Formation. The evolution of bioclastic reservoirs was controlled by the balance between dissolution and neomorphism of the aragonitic bivalve bioclasts, favoring the generation of poorly-connected moldic porosity or the preservation of well-connected interparticle porosity, which controlled the permeability of the reservoirs. The magnesian clay ooidal arenites suffered dissolution and replacement by dolomite and silica, what generated highly heterogeneous pore systems, composed by moldic, intercrystalline, vugular and microcrystalline pores. The growth of crystal shrubs in coalescent calcite crusts generated growth-framework primary porosity, which was reduced mostly by dolomite cementation, or enlarged by dissolution, what enhanced their permeability. Non-coalescent calcite crusts usually show strong interaction with syngenetic magnesian clay minerals, which fill interstices and/or are interbedded with the crusts. Their porosity is related to dissolution of the clays, what generated poor permeability. Intraclastic rocks usually display clay matrix, or are compacted and cemented. Where they show interparticle primary porosity or matrix dissolution, they may have good porosities and permeabilities. The heterogeneous dolomitization of both formations, either destroyed the primary or early diagenetic porosity, or generated high porosity and permeability values in the dolostones. Relationships of replacement and compaction indicate that most of the diagenetic processes occurred during eodiagenesis, controlled mostly by the instability of the aragonite in the bioclastic Coqueiros reservoirs, and of the magnesian clay minerals in the Macabu Formation. This study represents the first published petrographic characterization of the unconventional sag carbonate reservoirs, and stresses the crucial importance of systematic petrographic studies for the understanding and prediction of the quality of complex reservoirs.

Key words: Lagoa Feia Group, lacustrine carbonates, diagenesis, permeability, porosity

1 INTRODUCTION

Lacustrine carbonate reservoirs from the Lagoa Feia Group of the Campos Basin have been yielding significant production from oilfields in the shallow offshore portion of the basin (Bruhn *et al*., 2003). Such reservoirs belong to the Coqueiros Formation (Barremian, rift stage of Campos Basin), corresponding to bioclastic rudstones and grainstones, and have been intensively studied during previous decades (Baumgarten, 1985; Bertani and Carozzi, 1985a, 1985b; Baumgarten *et al*., 1988; Dias *et al*., 1988; Abrahão and Warme, 1990; Horschutz and Scuta, 1992; Carvalho *et al*., 2000; Rangel and Carminatti, 2000). On the other hand, the Macabu Formation reservoirs (Aptian, post-rift sag stage) have not attracted much attention till recently. The Macabu Formation is composed mainly by a succession of syngenetic magnesian clays and calcite crusts, which microbial or abiotic genesis is controversial (Dias, 2005; Terra and Lemos, 2008; Terra *et al*., 2010; Dorobek *et al*., 2012; Wright, 2012, 2013; Devaux, 2014; Tosca and Wright, 2014; Wright and Barnett, 2015). After the discovery of significant light oil accumulations in the sag and rift carbonates of the so-called Pre-salt province of the Santos Basin, new studies of potential analogs from the Campos Basin and other basins have been performed (Freire *et al*., 2011; Sebastien *et al*., 2011; Beglinger *et al*., 2012; Hamon *et al*., 2012; Altenhofen, 2013; Buchheim and Awramik, 2013; Jahnert and Collins, 2013; Perri *et al*., 2013; Muniz, 2014; Thompson *et al*., 2015; Armelenti *et al*., in press).

Lacustrine carbonates strongly contrast with marine carbonates, because the latter have a much greater control of geological age on their environmental formation conditions (Moore, 2001). Lacustrine carbonates have their genesis controlled by local geological settings, such as tectonics, climate, fauna and specific geochemical conditions that can form very particular deposits (Davis and Wilkinson, 1983; De Wett *et al*., 2002; Dunagan and Turner, 2004; Gierlowski-Kordesch, 2009; Fedorchuk, 2014). The deposition of lacustrine carbonates is controlled by several factors, including hydrological (input and output of surface waters, precipitation and groundwater flow), sediment input and temperature variations (Platt and Wright, 1991; Tucker and Wright, 1991). Water geochemistry plays a major importance on lacustrine environments, as it controls a number of processes, including development of microbial communities (Riding and Liang, 2005; Vasconcelos *et al*., 2006; Spadafora *et* *al*., 2010), abiotic precipitation of carbonates (Riding, 2008) and associated clay minerals (Calvo *et al*., 1999; Pozo and Casas, 1999; Furquim *et al*., 2008), and the development of algae, ostracodes, mollusks (bivalves or gastropods), commonly found in these settings (Casanova, 1986; Renaut *et al*., 1986; Vincens *et al*., 1986; Harris *et al*., 1994; Frogley *et al*., 2002).

The pore systems of carbonate reservoirs, both continental and marine, show great diversity of pore types, resulting in very heterogeneous reservoirs (Choquette and Pray, 1970; Mazzullo and Harris, 1991; Lucia, 1995; Moore, 2001; Mazzullo, 2004; Lønøy, 2006; Ahr, 2008). Porosity is controlled by many factors, including heterogeneous deposicional processes, controlled by the overprint of biological and chemistry activity, intense diagenesis due to the chemical reactivity of carbonates, and hydrothermal and fracturing processes (Ehrenberg *et al*., 2006, 2012; Chafetz, 2013; Brigaud *et al*., 2014). Usually, the evolution of the pore system tends to be controlled by the depositional fabric, while the superimposed diagenetic processes evolve accordingly to specific patterns for each type of petrofacies. The permeability is usually closely related to the porosity, being frequently controlled by the fabric and the diagenetic patterns (Dürrast and Siegesmund, 1999; Weger *et al*., 2009; van der Land *et al*., 2013; Rezende and Pope, 2015). The study of porosity geometry is very important in carbonate reservoirs characterization, supporting petrophysical interpretation and improving predictability of reservoir quality and heterogeneity (Basan *et al*., 1997; Mountjoy and Marquez, 1997; Eichenseer *et al*., 1999; Machel, 2005).

The aim of this work is to assess the diagenesis and porosity of the Lagoa Feia Group reservoirs in the northern part of Campos Basin through the combination of petrographic and petrophysical characterization of both the bioclastic deposits from the Coqueiros Formation and the unconventional deposits from the Macabu Formation, which genesis remains highly controversial, and still poorly understood. We intend to characterize the pore systems of the reservoirs in order to interpret their genesis and evolution, searching for better understanding and predictability of the petrophysical characteristics in the Lagoa Feia Group reservoirs.

2 GEOLOGICAL SETTING

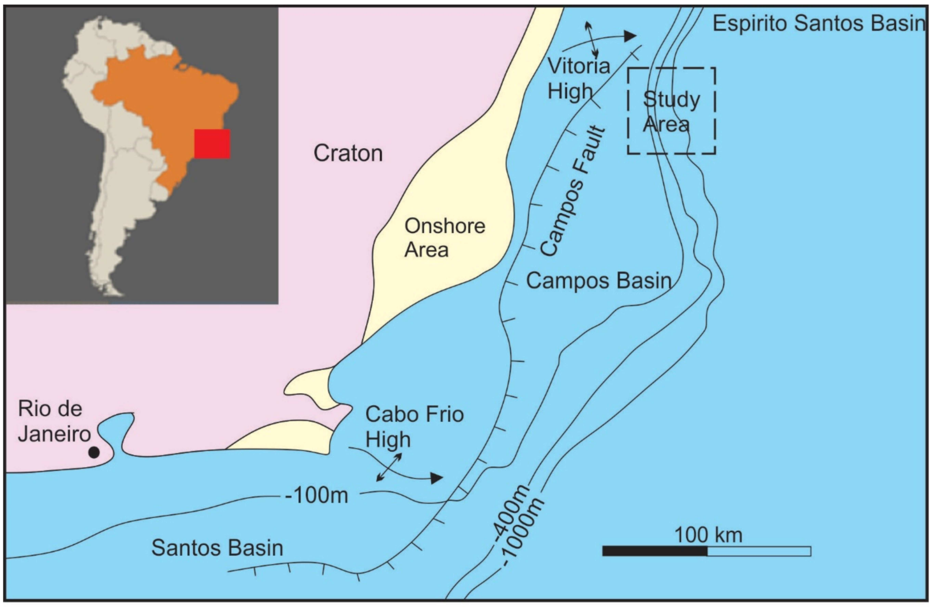
Campos Basin is a passive margin basin situated on the northern coast of the Rio de Janeiro State, Brazil (Fig. 1), extending to the southern coast of the Espírito Santo State, being geographically located between parallels 21° and 23° South. It is located between the Santos Basin to the South, bounded by the Cabo Frio volcanic high, and the Espírito Santo Basin to the North, bounded by the Vitória High.

The Campos Basin was formed under an extensional tectonic regime during the breakup of the Gondwana Continent during Late Jurassic/Early Cretaceous, preceding the final separation of South America and Africa, and the formation of the South Atlantic Ocean (Rabinowitz and LaBrecque, 1979; Austin and Uchupi, 1982; Nürnberg and Müller, 1991; Cainelli and Mohriak, 1999; Guardado *et al*., 2000). The formation of the Atlantic Ocean starts with crustal thinning and breakup of San Francisco-Congo-Rio de La Plata-Kalahari Cratons, accreted during the Eoproterozoic orogeny. South Atlantic formation begins during Jurassic with rifting between Argentina and southern Africa, and through the Equatorial Margin (Szatmari, 2000; Meisling *et al*., 2001; Mohriak *et al*., 2008). The rupture of the central portion, controlled by a resistant cratonic core (São Francisco – Congo Craton), occurred later, in the Hauveterivian (Clemson *et al*., 1997; Karner and Driscoll, 1999), resulting in the development of a narrow rift zone in this region, where Campos Basin is located. This contrasted with rift regions controlled by Proterozoic fold belt, which developed wider basins (Rosendahl *et al*., 2005).

The Barremian to Aptian Rift Stage is characterized by increasing lithospheric extension and asthenospheric uplift (McKenzie, 1978; White and McKenzie, 1988), associated with an extensive intracratonic tholeiitic volcanism (Turner *et al*., 1994; Mohriak *et al*., 2008; Torsvik *et al*., 2009). Elongated deep lakes were formed parallel to basement lineaments, filled with both volcanic and sedimentary rocks (Dias *et al*., 1988; Mizusaki *et al*., 1988). The presence of transfer zones played an important role in the partitioning of the rifted margin in sub-basins (Meisling *et al*., 2001; Muniz, 2014). The Aptian sag stage is characterized initially by uplift and erosion of the rift section, resulting in a regional discordance (Karner and Driscoll, 1999; Dias, 2005; Winter *et al*., 2007). The initial carbonate, argillaceous and clastic sediments were overlain by evaporites formed in arid climate under influx of marine incursions (Abrahão and Warme, 1990; Karner and Gambôa, 2007). The last stage (drift phase) is characterized by the development of oceanic crust and passive margin (Chang *et al*., 1988, 1992; Aslanian *et al*., 2009).

The Lagoa Feia Group, originally defined by Schaller (1973) as Lagoa Feia Formation, unconformably overlies pre-rift basaltic floods from the Cabiúnas Formation, comprising the non-marine to transitional sequences from the Campos Basin. Subsequently, the Lagoa Feia Formation has been subdivided into four distinct depositional sequences bounded by unconformities, including: the Basal Clastic Sequence; the Talc-Stevensitic Sequence; the Coquinas Sequence; and the Clastic-Evaporitic Sequence (Dias *et al*., 1988). The two last sequences separated by the pre-Alagoas unconformity. More recently, Winter *et al*. (2007) have subdivided Lagoa Feia Formation into several formations, elevating it to a group status (Fig. 1).

The Coqueiros Formation is composed mainly by bioclastic rudstones and grainstones, informally known as “coquinas”, interfingered with proximal clastic sediments of the Itabapoana Formation towards the western margin of the basin. The Coqueiros Formation is dominantly constituted of bioclasts of bivalve mollusks, with subordinate amounts of gastropods and ostracods. Castro *et al*. (1981) separated the bioclastic deposits into two types: “detrital coquinas” consisting of bioclasts and non-carbonate grains (talc-stevensite ooids and peloids, and mud intraclasts); and “pure coquinas”, essentially composed by bioclasts. Bertani and Carozzi (1985a) recognized several microfacies, separating sequences dominated by bivalves and by ostracods, and according to variations on composition, grain size and matrix/cement content. These authors related the distribution of microfacies to two distinct depositional models: playa-lake settings dominated by ostracods, and pluvial-lake settings dominated by bivalve mollusks. Carvalho *et al*. (2000) recognized seven main depositional environments based on facies associations, including bioclastic calcarenite beaches, bioclastic sandy beaches and marginal lacustrine settings, bioclastic sheet / bar fringes, consisting of shell debris deposited by storm events, bioaccumulation banks and deep lacustrine settings.



PERIOD/ EPOCH	STAGE	LOCAL STAGE	OSTRACOD BIOZONES		STRATIGRAPHY			BASIN STAGE
LOWER CRETACEOUS	113Ma	ALAGOAS	-	LAGOA FEIA GROUP	RETIRO Fm			SAG
	APTIAN		011		ITABAPOANA Fm	GARGAU Fm	MACABU Fm	
		JIQUIÁ	010		ITABAPOANA Fm	COQUEIROS Fm		RIFT
	125Ma		1100					
	BARREMIAN		1020					
			1010					
			009.3					
			009.2			ATAFONA Fm		
			009.1					
		BURACICA	008					
			007					
		ARATU	006					
	129.4Ma		005.5					

Figure 1. Localization of the Campos Basin and the study area (Modified from Dias *et al*., 1988), biozones (Moura and Praça, 1985; Silva-Telles, 1992) and stratigraphy of the Lagoa Feia Group (Winter *et al*., 2007).

Deposits from the Macabu Formation were referred by Dias *et al*. (1988) as carbonate facies consisting of nodular diagenetic limestones and laminated limestones, “probably algal in origin (stromatolites)”. Later, Dias (2005) interpreted laminites and microbial stromatolites as supratidal and upper intertidal facies, overlaid by lower intertidal clay rich facies and infratidal mudstones deposited in an epicontinental context with extensive pre-evaporitic clastics located in shallow proximal regions. Muniz and Bosence (2015) interpreted the absence of both marine biota and normal freshwater biota, such as charophytes, and the presence of non-marine ostracods, as evidence of a brackish-water–lacustrine environment, refuting the presence of tidal influence on microbial development. On the other hand, recent interpretation suggests that most deposits from the Macabu Formation are chemical precipitates, controlled by the geochemistry of alkaline lacustrine waters (Wright, 2011, 2012, 2013; Tosca and Wright, 2014). Wright and Barnett (2015) identified a typical cyclothem of the Macabu Formation as composed by three main components: mud-grade laminated carbonates, accumulated in flooding phase, calcite spherulites within a matrix of Mg-silicates and calcitic shrub-like growths trigged by evaporation, controlled by pH and ionic activity.

3 METHODS

This study has been based on 780 samples from 3 cores and sidewall cores collected from five wells (Fig. 2). The methods applied included optical petrography, ultra-violet fluorescence (UVF), cathodoluminescence (CL), scanning electron microscopy (SEM), X-ray diffraction (XRD) and petrophysical analysis. In order to characterize the primary and diagenetic constituents, the pore types, and their relationships, optical petrography has been performed on thin sections impregnated with blue epoxy resin. Pore types were classified according to Choquette and Pray (1970). Thin sections were stained for carbonate identification with an acid solution of alizarine red and potassium ferrocyanide (Dickson, 1965). A Zeiss Colibri fluorescence light-emitting diode (LED) UV microscope was used to evaluate organic matter presence and relations of replacement between fluorescent minerals. Cathodoluminescence (CL) microscopy was performed in a Cambridge Image Technology Ltd. (CITL) luminoscope in order to identify diagenetic carbonate phases. CL analyses were performed in CL8200 MK5-2 equipment coupled to Leica microscope, under operating conditions of 300 mA and 17 kV. SEM analyses using backscattered and secondary electrons were executed in ZEISS EVO LS15 and JEOL JSM 6490LV electron microscopes, both equipped with OXFORD Inca/Aztec energy dispersed spectrometers EDS, in order to investigate paragenetic relationships between primary and diagenetic constituents, and to identify the elemental mineral composition in representative rock fragments and thin sections. XRD analysis of clay fraction, including air dried, glycol saturated and heating treatment were used to confirm the clay mineral species. Petrophysical analyses were performed on plugs cut from cores and on sidewall cores in order to determine porosity and permeability according to norm API RP 40:1998. Mercury Injection Capillary Pressure (MICP) was performed on a Micrometrics AutoPore IV 9500 equipment to determine the pore throat distribution on representative samples.

4 RESULTS

4.1 Primary Texture and Composition

4.1.1 Coqueiros Formation

The primary carbonate composition of deposits from the Coqueiros Formation includes bioclasts of bivalves, gastropods and ostracods. Bivalve bioclasts are the main component of the carbonate reservoirs of the Coqueiros Formation (Fig. 3A). Bivalves always occur disarticulated and reworking is very variable. Bivalve bioclasts are frequently broken, with dimensions ranging from 0.07 to 25 mm (average = 2.3 mm) in the thin sections. Dissolution is very frequent, indicating an originally aragonitic composition of the shells. Drusiform and mosaic calcite cement usually fills intraparticle and moldic porosity.

Neomorphism, characterized by mimetic replacement of the original foliated structure of the aragonite shells by calcite mosaic is common. Bivalves are usually the main component of rudstones and grainstones in the studied wells, composing up to 100% of primary constituents. These bioclastic limestones are mostly massive, with a parallel to sub-parallel, occasionally chaotic fabric, and poor to good sorting.

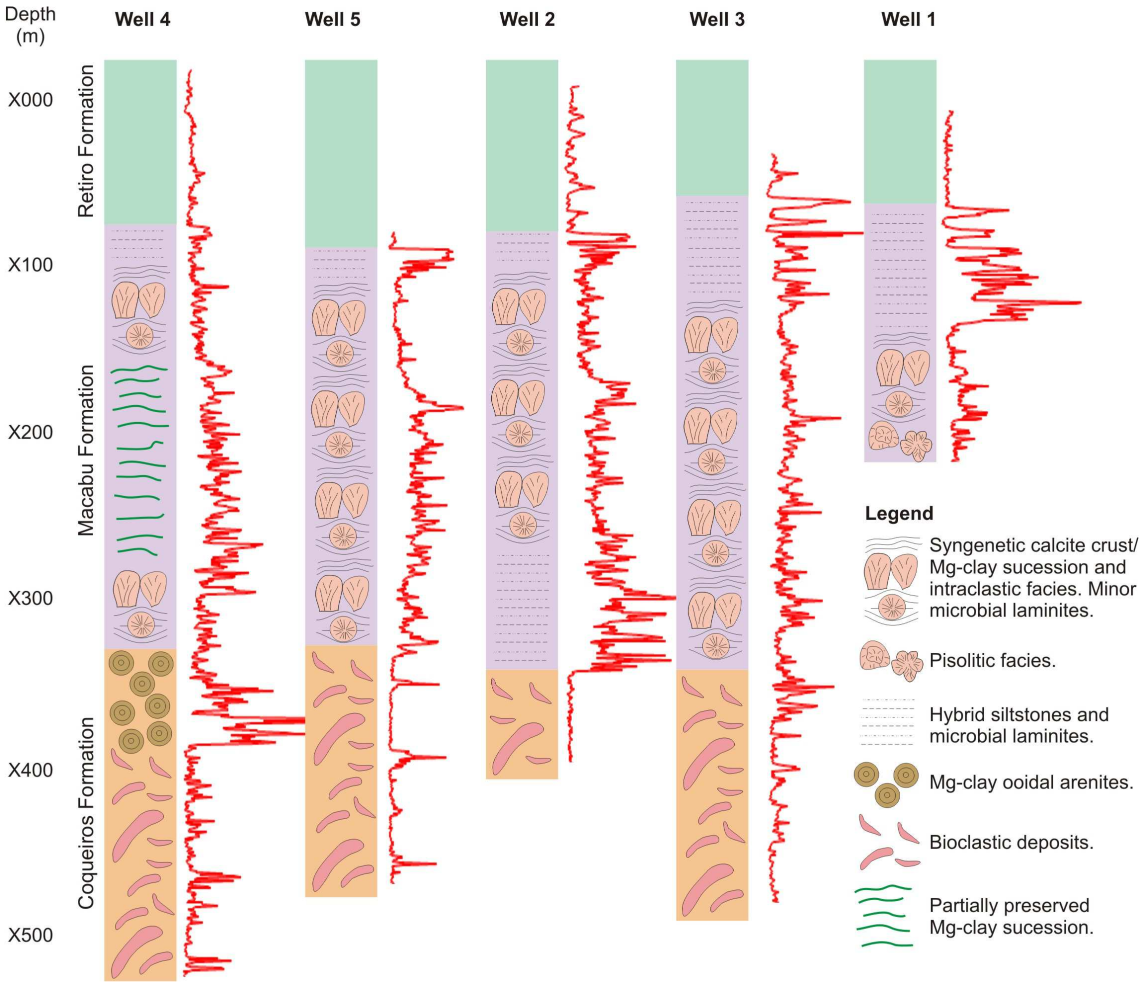


Figure 2. Well correlation showing Gamma Ray log and facies succession.

Gastropods are occasionally observed and typically are whole and not abraded (Fig. 3B). Their neomorphism and dissolution are frequent, also indicating an original aragonitic composition. Intraparticle porosity within the gastropods shells is common. The ostracod bioclasts are commonly whole, sometimes articulated, and occasionally recrystallized. In some cases, ostracods are the main primary component of grainstones (Fig. 3C).

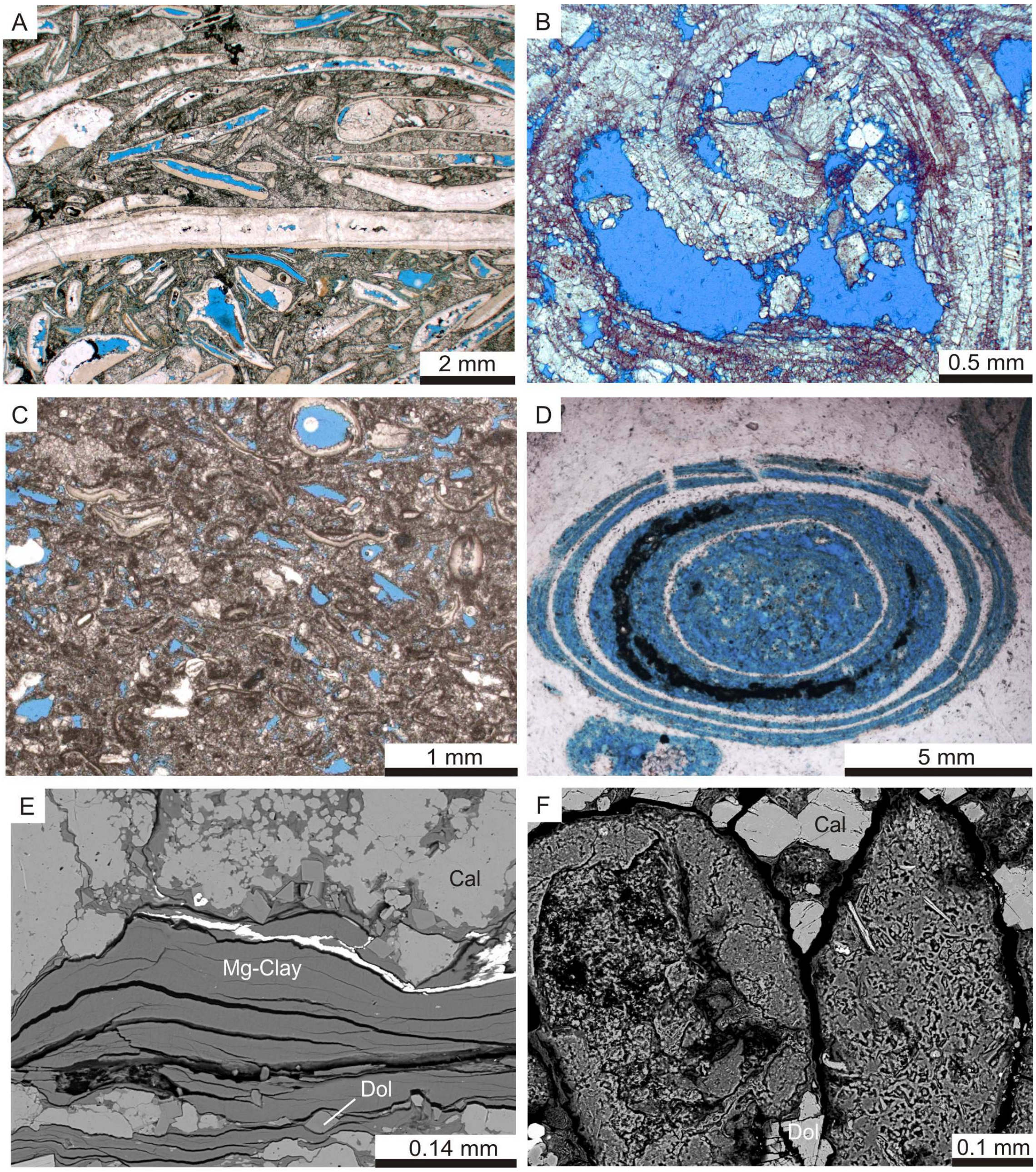


Figure 3. Photomicrographs with uncrossed polarizers (//P) and backscattered electrons microscopy (BSE) of common constituents of Coqueiros Formation and magnesian clays from Macabu Formation: (A) Poorly sorted rudstone composed by bioclasts of bivalves (//P). (B) Neomorphised gastropod bioclast (//P). (C) Recrystallized grainstone composed mainly by ostracod bioclasts (//P). (D) Ooids of Mg-clay dissolved/replaced and cemented by silica (//P). (E) Mg-clay laminated aggregates partially replaced by dolomite and calcite and shrinked

(BSE). (F) Mg-clay ooid and peloid. Ooid on the left has a partially dissolved nucleus and an outer massive texture, whereas ooid on the right shows a dominantly massive texture (BSE).

Mg-clay ooids occur only in Well 4 on the top of Coqueiros Formation, accumulated apparently *in situ*, in an about 50 meters thick layer of ooidal arenites (Fig. 3D). Previous studies pointed to a composition of stevensite, kerolite and talc for such ooids (Armelenti *et al*., in press; Bertani and Carozzi, 1985a, 1985b; Rehim *et al*., 1986; Abrahão and Warme, 1990). These clay ooids occur commonly dissolved and replaced by dolomite and quartz, and often deformed. They are rarely observed in the bioclastic rudstones and grainstones. Intraclasts of mud, volcanic and siliciclastic grains are occasionally observed in small amounts in the bioclastic rudstones and grainstones.

4.1.2 Macabu Formation

The primary composition of the Macabu Formation rocks consists predominantly of intrabasinal constituents, including syngenetic magnesian clays and calcite crusts, intraclasts, bioclasts of ostracods and fish. Siliciclastic silt-sized grains and detrital clays occur in subordinate amounts.

Syngenetic Mg-clays occur in different habits, often dissolved and/or replaced in variable intensity by dolomite, silica and calcite (Fig. 3E). Magnesian clays are relatively more abundant in Well 3, where they are preserved in an interval of about a hundred meters thick. In the other wells, they are poorly preserved. Mg-clays occur as laminated or massive aggregates, as peloids with very variable size (Fig. 3F), as ooids with up to 2 mm of diameter, as thin coatings covering intraclasts, and as a matrix in hybrid siltstones and intraclastic rocks. Laminated and massive aggregates show often contraction. Clay peloids and ooids are often deformed. Mg-clays are pale to dark brown in color, indicating variations in the organic matter content. They show blue, yellow and light brown to orange UV fluorescence (Fig. 4A and 4B). Even when partially dissolved or replaced by carbonates or silica, clay particles and laminated aggregates show intense fluorescence, indicating that at least part of the fluorescence is related to organic matter. SEM analyses show very low crystallinity, even in high magnification. EDS analyses indicate considerable amounts of SiO_2 (about 65%), MgO (about 30%) and lower contents of Al_2O_3 (about 2%). FeO, K_2O, CaO_2 and NaO occur in trace amounts, usually below 1%. The elemental composition of the clays and the XRD analysis indicate a dominant mineralogy made of stevensite. The Mg-clay syngenetic laminations were formed probably in low-energy environments, whilst peloids and ooids were formed in more energetic conditions.

Non-magnesian syngenetic calcite occurs as divergent crystal aggregates exhibiting radial-fibrous and more commonly fascicular-optic texture (*sensu* Kendall, 1977; *i.e.* plumose or increasingly divergent). Calcite crystals grew predominantly in a vertical to near-vertical orientation, with average individual length of 1.55 mm (up to 12 mm). The fascicular aggregates display commonly a shrub shape that coalesce both vertically and horizontally, forming calcite crusts of variable thickness (Fig. 4C). In some cases, fibrous calcite aggregates tend to form millimetric continuous and isopachous palisade crusts (Fig. 4D), intercalated with microcrystalline calcite/dolomite, siliciclastic mud or syngenetic Mg-clays. Recrystallization is frequent, generating apparently massive forms, although with remnants of the original fibrous crystal fabric recognizable in polarized light and SEM images (Fig. 4E). The engulfment and replacement of siliciclastic grains, Mg-clay peloids, and ostracod bioclasts is frequent. In some cases, thin laminations occur within the aggregates, which can be associated to interruptions in crystal growth, allowing accumulation of fine particles or development of microbial films. CL images show a homogeneous luminescence pattern, which indicates the total recrystallization of original aggregates (Fig. 4F). Micropores and fluid inclusions are commonly observed as defects formed by recrystallization or by dissolution of included Mg-clay peloids or other particles (Fig. 5A). UV fluorescence is very weak or absent, indicating that no or only minor amounts of organic matter were included in syngenetic calcite aggregates. Crenulated laminated structures are rare, occurring at the base of some calcite aggregates, indicating possible nucleation of few of the fibrous calcite aggregates by microbial deposits. Microbial remnants were, however, not identified in SEM analyses.

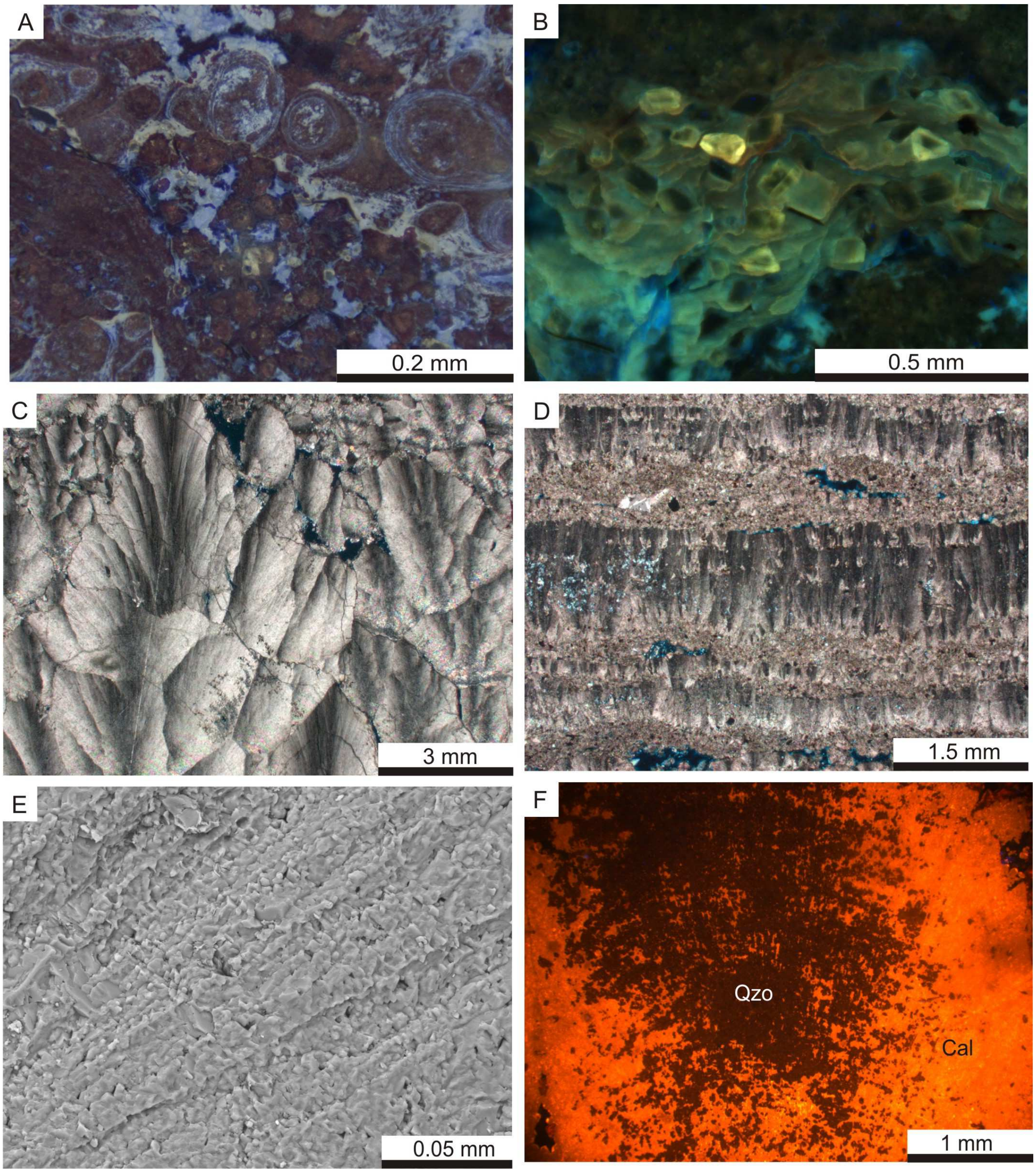


Figure 4. Photomicrographs with crossed polarizers (XP), secondary electrons microscopy (SEM), ultra-violet fluorescence (UVF) and cathodoluminescence (CL) images showing different aspects of syngenetic Mg-clays and calcite crusts: (A) Mg-clays as ooids and as coatings on intraclasts (UVF). (B) Laminated Mg-clay aggregates partially replaced by dolomite (UVF). (C) Calcite crusts formed by coalesced shrubs of fascicular aggregates (XP). (D) Millimetric, continuous, isopachous fibrous calcite palisade crusts (XP). (E) Detail of the mimetic recrystallization of fibrous calcite, showing large amount of micropores (SEM). (F) Shrub with massive calcite recrystallization and replacement by quartz (CL).

Spherical to sub-spherical non-magnesian calcite pisoliths (1 mm to 5 cm) with divergent growth pattern are abundant in Well 1 (Fig. 2), showing fascicular-optic and/or radial-fibrous texture, similar to that observed in the crusts, but with concentric growth (Fig. 5B). Some pisoliths are formed by thin concentric laminae (~0.5 mm), while others are formed by divergent coalescence of fascicular-optic calcite aggregates, reaching up to 2 mm in diameter. The pisoliths were commonly deformed and recrystallized. No fluorescence is observed in the pisolithic aggregates, indicating that their growth was not directly promoted by any microbial colony. Most pisoliths seem to occur *in situ*, but in some cases, they are mixed with intraclasts and other particles.

Phosphatic bioclasts are common in Well 1, but show a more erratic distribution in the other wells. Phosphatic bioclasts, such as fish scales, teeth, vertebrae and other bones, occur scattered or concentrated, locally together with undifferentiated phosphatic grains. Ostracod bioclasts are frequently concentrated in millimetric levels (Fig. 5C), and commonly occur as articulated shells, with limited reworking. Their partial to total replacement by quartz or dolomite is common. Dissolution is uncommon, indicating an original low-Mg calcitic composition.

Intraclastic grainstones and rudstones, constituted by fragments of reworked calcite crusts, Mg-clay laminated aggregates, calcite spherulites, laminated microbial carbonates, and microcrystalline calcite peloids are common (Fig. 5D). The average diameter of such particles is 1.56 mm (up to 25 mm). Calcite intraclasts are commonly rounded, with shapes controlled either by their internal crystal habit or by their laminated fabrics. Their recrystallization and replacement by quartz and dolomite is common. Mechanical compaction through fracturing and deformation, and chemical compaction through pressure dissolution along interparticle contacts or stylolites is frequently intense. Syngenetic clay matrix is a common constituent in intraclastic rocks, so that facies with this combined composition were considered as hybrid “packstones”, due to lack of nomenclature.

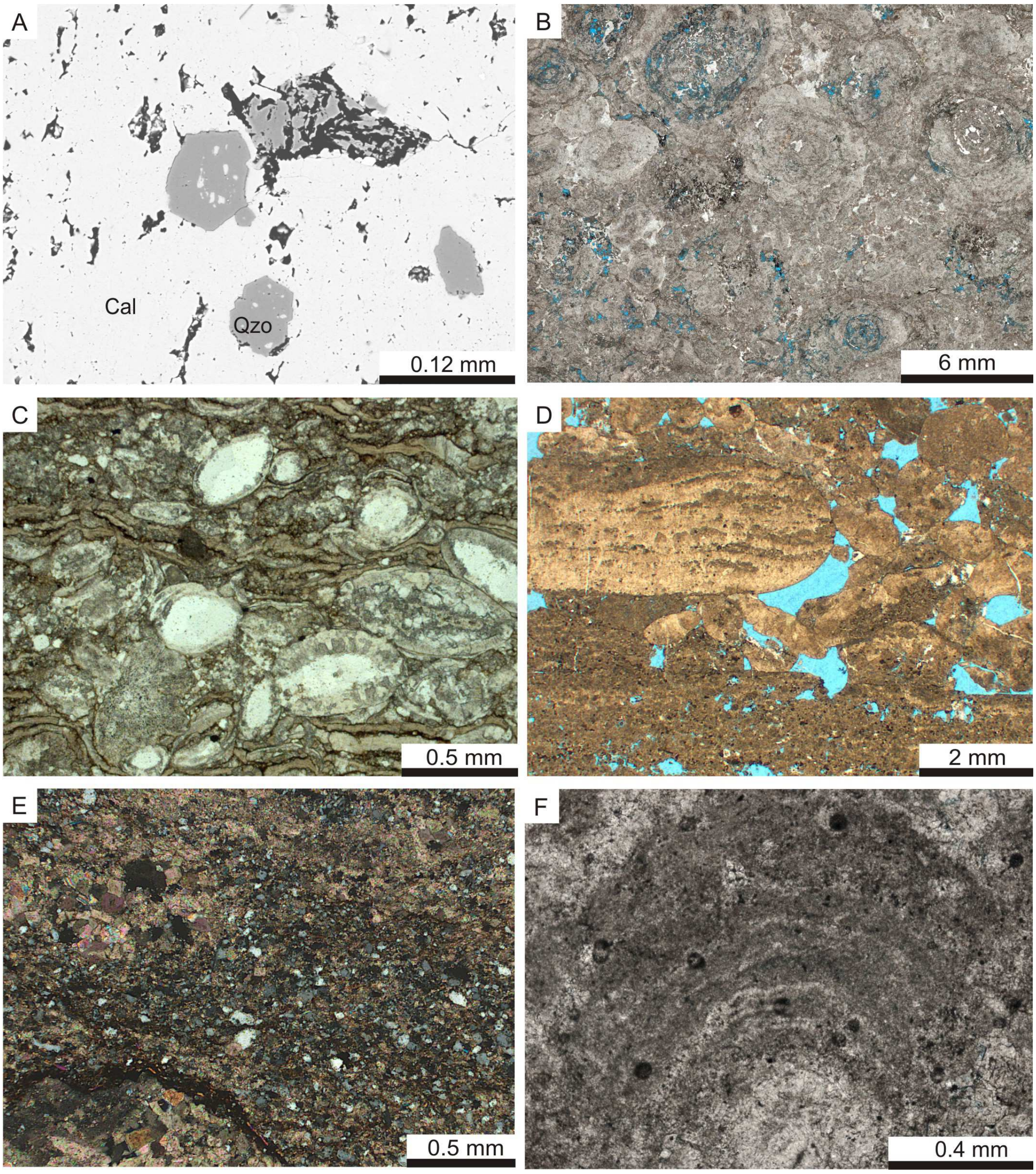


Figure 5. Photomicrographs and BSE images showing common primary components of the Macabu deposits: (A) Detail of the internal structure of a fascicular-optic calcite aggregate, replaced by quartz and showing abundant microporosity (BSE). (B) Low-Mg calcite pisoliths recrystallized and partially dissolved (//P). (C) Articulated ostracod bioclasts filled and replaced by quartz within deformed Mg-clay laminations replaced by dolomite (//P). (D) Intraclasts of fibrous and microcrystalline calcite crusts (//P). (E) Hybrid siltstone composed mainly by quartz grains and by dolomite and calcite replacing syngenetic Mg-clays and detrital clays (XP). (F) Crenulated, irregular laminations probably formed by the mineralization of microbial communities (//P).

Silt to sand-sized siliciclastic grains occur in all analyzed wells, usually occurring in small percentages of less than 1%, either included within or among the calcite crusts aggregates, or mixed with intraclasts or with Mg-clay matrix. Quartz, biotite, muscovite, orthoclase and plagioclase occasionally compose hybrid and siliciclastic siltstones and arenites (Fig. 5E). They are particularly abundant at the base of Macabu Formation in Well 2, within an interval about 100 meters thick and at the top of Macabu Formation in all wells (Fig. 2).

Millimetric, roughly crenulated calcite laminations, which could be related to the mineralization of microbial deposits, are occasionally observed. Intense recrystallization and replacement usually render difficult the identification of the microbial origin of these deposits (Fig. 5F).

4.2 Diagenetic Composition

4.2.1 Coqueiros Formation

The diagenetic processes observed in the Coqueiros Formation were directly controlled by the composition of the primary constituents. The main diagenetic constituent of bioclastic rudstones and grainstones is calcite, derived from dissolution of the bioclasts and precipitated with different habits. Silica and dolomite show a subordinate, more localized occurrence. The diagenetic patterns of the Mg-clay ooidal arenites are very heterogeneous. Frequently the ooids were cemented by dolomite or quartz, almost completely silicified, dolomitized and/or dissolved. Ooids that were not cemented or replaced during eodiagenesis show intense deformation.

Calcite

Bivalve bioclasts were either dissolved or neomorphised, commonly showing a heterogeneous range from partial to total dissolution, with variable intensity of cementation of the intraparticle and moldic porosity, to total neomorphism (Fig. 6A). The distinction between neomorphism and intraparticle/moldic pores cementation is, in many cases, difficult. Neomorphism of aragonitic bivalve and gastropod bioclasts often forms a mosaic of calcite crystals elongated along the original shell structure. In contrast, the intraparticle and moldic pores were filled by drusiform (Fig. 6B), and blocky calcite. The bioclasts were commonly covered by continuous rims of microcrystalline to prismatic calcite (Fig. 6A and 6B). Drusiform calcite frequently represents the main interparticle cement. Subordinate cements filling interparticle porosity include blocky, coarse mosaic and microcrystalline calcite. Dull CL indicates a more ferroan composition of neomorphised calcite than of interparticle cement that is red in CL (Fig. 6C). Recrystallization of early calcite to blocky and microcrystalline calcite is common, masking the original fabric. Calcite is rare in stevensitic ooidal arenites, occurring mainly as coarse mosaic fracture filling.

Dolomite

Dolomite occurs as a minor constituent in bioclastic rudstones and grainstones of Wells 3 and 5 and more rarely in Well 2. Blocky dolomite fills locally interparticle and moldic porosity, occasionally replacing diagenetic calcite. Saddle dolomite was observed filling interparticle, intraparticle, moldic and vugular porosity (Fig. 6D). Dolostones and totally dolomitized bioclastic rocks up to 10 meters thick were observed in Well 5. These rocks are composed of small rhombohedral and more rarely of saddle dolomite crystals. Dolomitization usually is mimetic, preserving part of the original fabric and porosity (Fig. 6E).

Blocky or mosaic dolomite replaces clay ooids and drusiform dolomite fills interparticle porosity in ooidal arenites (Figs. 7C, 7D and 7F). In some cases, dolomite selectively replaced only the ooids, while in other cases only cemented the interparticle porosity. In other cases, dolomitization was complete, mimetic or totally pervasive and obliterative. Rarely, saddle dolomite replaced diagenetic silica. CL images indicate that precipitation of dolomite occurs at various stages, showing zoned red luminescence patterns related to variations in the Fe/Mn ratio. Dolomite shows dark to light blue and light green UV fluorescence in ooidal sandstones. Millimetric to centimetric levels of pervasively dolomitized or silicified arenites occur intercalated, commonly through stylolitic contacts, heterogeneously organized.

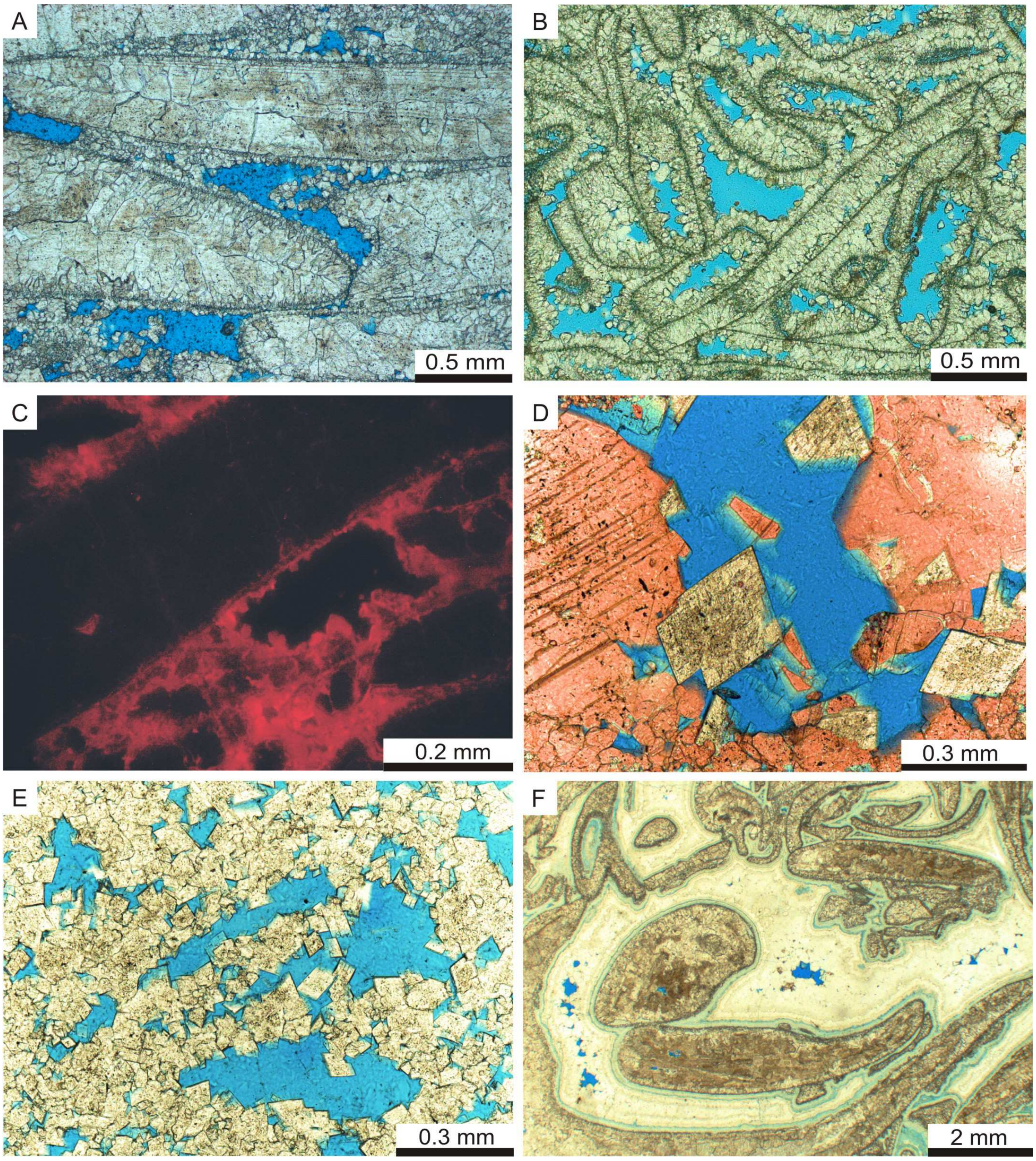


Figure 6. Photomicrographs showing main diagenetic constituents of bioclastic deposits from Coqueiros Formation: (A) Rudstone composed by neomorphised bioclasts with interparticle fine rim of calcite pore-lining (//P). (B) Rudstone with dissolved bioclasts filled by drusiform calcite and interparticle porosity cemented by bladed rim (//P). (C) CL micrograph showing neomorphised bioclasts (black) with interparticle pore lined by rim and fine mosaic of calcite

(red). (D) Saddle dolomite filling and replacing calcite in vug (//P). (E) Mimetic dolostone (//P). (F) Bioclastic rudstone cemented by displacive silica (//P).

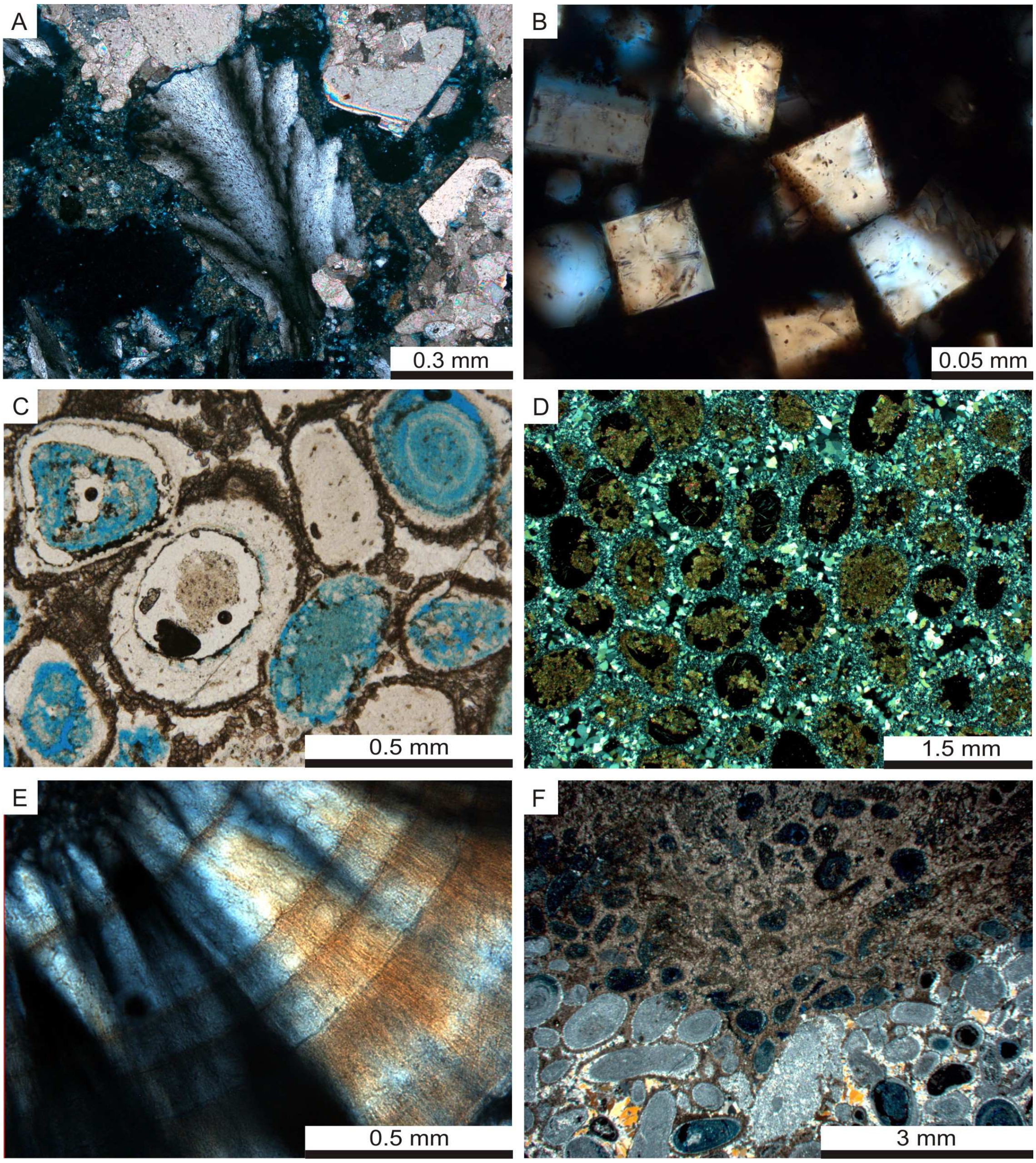


Figure 7. Minor diagenetic constituents of bioclastic rudstones and grainstones: (A) Celestine filling vug porosity (XP). (B) Svanbergite replacing mud intraclast (XP); and common diagenetic constituents of ooidal arenites: (C) Ooids partially dissolved/replaced by silica (white) and covered by dolomite rim (//P). (D) Ooids replaced by dolomite and cemented by chalcedony rim and quartz mosaic (XP). (E) Banded chalcedony marked by fine brown inclusions (XP). (F) Contact between mostly silicified ooidal arenite and totally dolomitized/dissolved arenite (XP).

Silica

Chalcedony and quartz (as blocky, prismatic crystals, coarse mosaic, and microcrystalline), often occur in small amounts mainly replacing calcite and rarely filling interparticle, intraparticle, moldic and vugular porosity in bioclastic rudstones and grainstones. Microcrystalline quartz rims locally covered bioclasts. Drusiform quartz locally filled moldic and vugular porosity. Displacive quartz interparticle cement was observed in some samples (Fig. 6F).

In ooidal arenites, authigenic silica usually lined particles as fibrous and microcrystalline rims, followed by pore-filling fine to coarse, drusiform mosaic (Fig. 7D). Total to partial replacement of ooids by microcrystalline silica is frequent (Fig. 7C). Microcrystalline silica, botryoidal or drusiform are observed filling vugular, interparticle and fracture pores. Botryoidal chalcedony commonly shows small brown inclusions that mark its growth (Fig. 7E), and green and blue UV fluorescence. Microcrystalline silica shows abundant microporosity and fracturing.

Minor Constituents

Bioclastic deposits have a poor diversity of diagenetic species. Barite and celestine ($SrSO_4$) were rarely observed, filling interparticle porosity or replacing diagenetic constituents (Fig. 7A). The aluminum-phosphate-sulphate (APS) minerals svanbergite and goyazite ($SrAl_3(PO_4)(SO_4)(OH)_6$ and $SrAl_3(PO_4)(PO_3OH)(OH)_6$) were locally described as pseudocubic crystals replacing mud intraclasts (Fig. 7B) in bioclastic rudstones. Traces of fluorite, svanbergite-goyazite, and pyrite were described replacing silica in ooidal arenites.

4.2.2 Macabu Formation

The diagenesis was very heterogeneous in the Macabu Formation. Processes and constituents vary significantly even in the millimetric scale, directly impacting the porosity and permeability of rocks. The main volumetric authigenic constituents are calcite and dolomite, followed by silica.

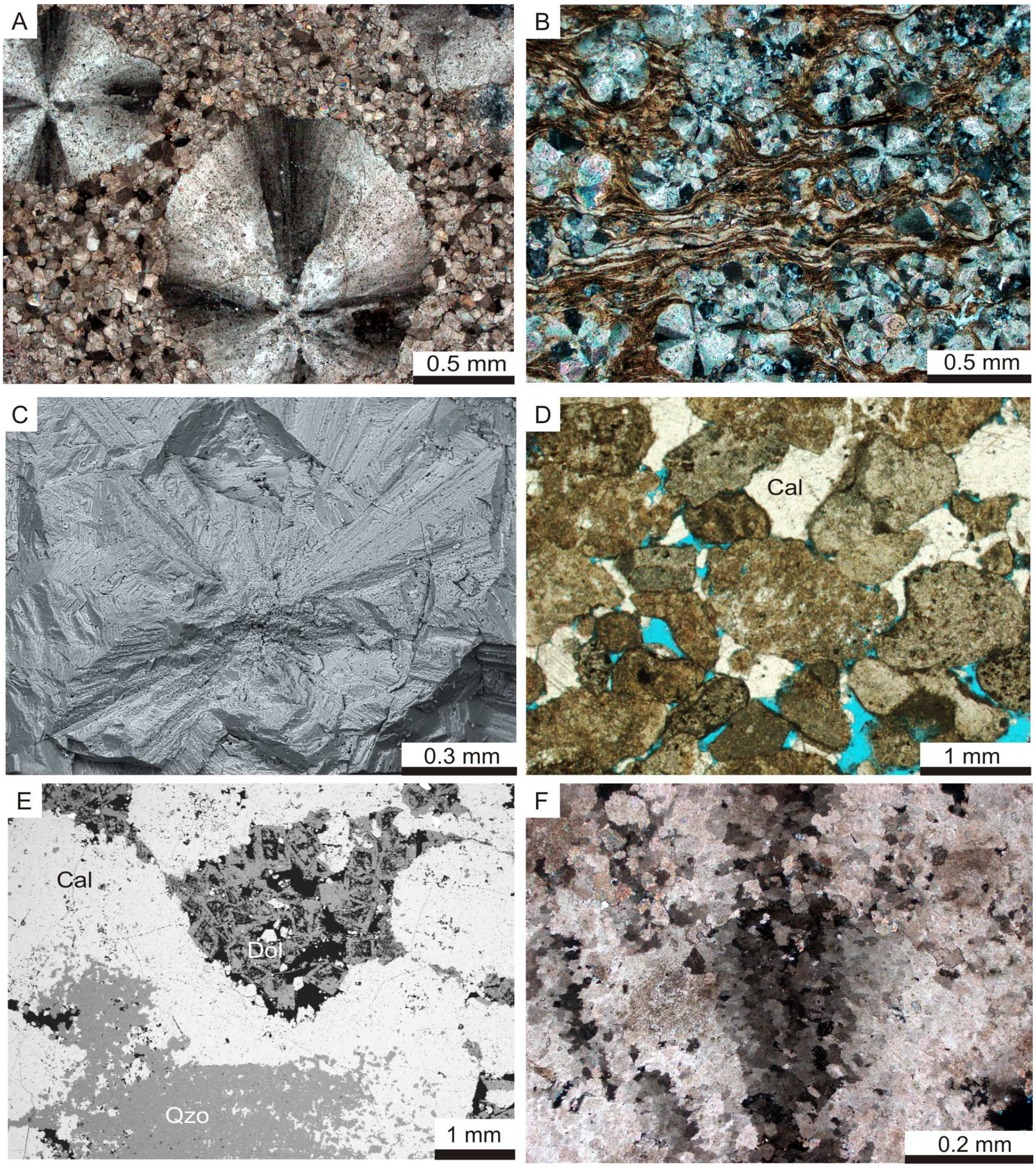


Figure 8. Photomicrographs, SEM and BSE images showing important aspects of diagenetic calcite and dolomite from Macabu Formation: (A) Spherulites replacing dolomitized syngenetic clay matrix (XP). (B) Spherulites partially silicified, displacing and replacing syngenetic dolomitized laminations (XP). (C) Detail of preserved original radial-fibrous structure on spherulite (SEM). (D) Coarse mosaic of calcite filling interparticle porosity on intraclastic grainstone. (//P). (E) Calcite crust partially replaced by microcrystalline quartz and cemented by partially dissolved dolomite (BSE). (F) Mimetic dolomitization replacing fascicular-optic calcite crust (XP).

Calcite

Calcite spherulites and hemispherulites are the most common diagenetic constituent of the Macabu Formation, occurring as an important component in all studied wells (Fig. 8A). Spherulites of 0.15 to 2.5 mm of diameter (mean = 0.8mm) frequently replaced Mg-clays and hybrid siltstones, displacing and deforming unconsolidated sediments (Fig. 8B). Most spherulites do not display a nucleus. Recognized nuclei include clay peloids, ostracods, intraclasts and lumps of microcrystalline calcite, which may be microbial in origin. Spherulites are often very abundant, coalescing as irregular levels (Fig. 8B), commonly with stylolitic contacts. Although spherulites were commonly recrystallized, some show a well-preserved original structure (Fig. 8C). Partial replacement by chalcedony and quartz is frequent.

Blocky, microcrystalline, and coarse mosaic calcite commonly occur as cement filling interparticle pore on intraclastic grainstones and rudstones (Fig. 8D) and more localized filling growth-framework porosity in calcite crusts. Microcrystalline calcite occasionally replaces dolomite. Calcite rims of scalenohedral, prismatic and bladed crystals usually cover calcite intraclasts and rarely calcite fascicular aggregates in the crusts. Coarse mosaic calcite is observed filling vugs, fractures and channels (Fig. 9B). Partial or total recrystallization of primary constituents to microcrystalline calcite is common.

Dolomite

Dolomite as blocky, microcrystalline and more rarely mosaic dolomite are the most common diagenetic constituent filling growth-framework pores in calcite crusts (Fig. 8E and 9C) and interparticle pores in grainstones and intraclastic rudstones. Dolomite has a great diversity of habits mainly replacing Mg-Clays (Fig. 4B, 8A and 8B) and often all primary constituents. Blocky dolomite that replaced syngenetic clays often displaced and deformed the laminations (Figs. 3E and 4B). Saddle dolomite is common filling growth-framework, interparticle and vugular porosity, as well as replacing the primary and diagenetic constituents. Partial to total replacement of the original deposits (protoliths) is common, in some cases mimetic (Fig. 8F) and another destroying totally the original fabrics. Overdolomitization, *sensu* Halley and Schmoker (1983), meaning the precipitation of dolomite cementing pore spaces, and not only replacing Mg clays and calcite, is very common. Intercalations of millimetric to centimetric levels of cherts and dolostones, usually with stylolitic contacts, are observed (Fig. 9A). Dolomite often occurs zoned, showing variable red intensity pattern on CL, and/or partially dissolved, forming intracrystalline porosity (Fig. 8E). SEM images reveal abundant microporosity within the dolomite crystals. UV blue, green or brown is always observed (Fig. 4B), mostly where replacing syngenetic clays, probably due to the presence of organic matter inclusions.

Silica

Authigenic fibrous and microcrystalline silica selectively replaced syngenetic calcite crusts and calcite spherulites (Figs. 4F, 5A 8B and 8E). Silica frequently replaced carbonate intraclasts, syngenetic matrix and clay peloids and occasionally filled interparticle and growth-framework porosity. Brown to green UV is observed in silica replacing clay syngenetic matrix, probably due to organic matter. Pervasive replacement commonly promotes the formation of microporosity and/or vugular porosity. Coarse mosaic or prismatic quartz occasionally occurs replacing and cementing crusts and intraclastic grainstones, replacing syngenetic clay matrix, and cementing vugs and fractures (Fig. 9B).

Minor Constituents

Cubic, blocky and microcrystalline pyrite, although scarce (<1%), occurs in most described thin sections, mainly replacing magnesian clays and other primary or diagenetic constituents. Corse mosaic, prismatic or radial-fibrous barite and celestine were described in a few samples in most wells, usually filling fractures or growth-framework porosity (Fig. 9C). Pseudocubic-blocky, microcrystalline svanbergite-goyazite ($SrAl_3(PO_4)(SO_4)(OH)_6$ - $SrAl_3(PO_4)(PO_3OH)(OH)_6$) are recognized in SEM analyses mainly replacing Mg-clays, associated with silicification and rarely related to dolomitization. Fluorite, sphalerite, apatite and anatase were identified in SEM analyses, replacing different constituents. Coarse mosaic of anhydrite fills a fracture in one sample.

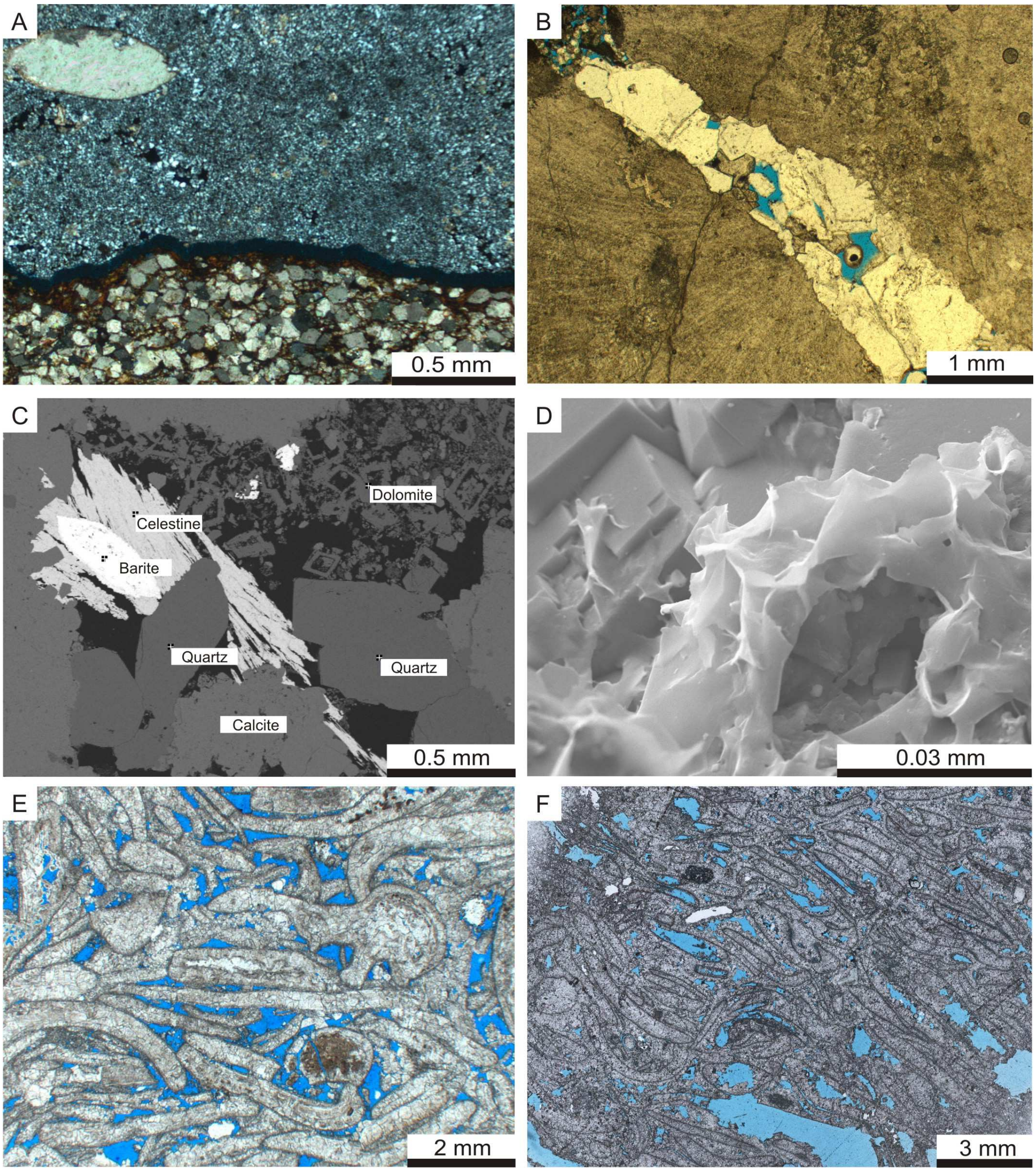


Figure 9. Diagenetic constituents of Macabu Formation: (A) Intercalation of microcrystalline chert and dolostones (XP). (B) Coarse-mosaic of silica filling calcite crust fracture (//P). (C) Quartz, celestine, barite and dolomite filling growth-framework enlarged porosity and replacing calcite crust (BSE). (D) IS covering dolomite filling growth-framework porosity (SEM); and common porous type on Coqueiros Formation: (E) Interparticle porosity in bioclastic rudstone (//P). (F) Predominant moldic-vugular porosity in bioclastic rudstone (//P).

Authigenic clay minerals are not significant in volume, but occur very frequently in the studied samples. Illite, I/S and kaolinite clays identified in MEV and XRD (Fig. 9D), commonly occur in calcite crusts mainly replacing detrital interstitial components and seem to remain as residual dissolution of syngenetic clays. In hybrid siltstones, authigenic clays occur replacing micas and detrital/syngenetic clays. SEM images show flake, scale and ribbon habits. Kaolinite occurs localized as calcite replacement in association with fluorite and celestine/barite.

4.3 Porosity

4.3.1 Coqueiros Formation

The main pore types in bioclastic rudstones are interparticle, intraparticle, moldic and vugular (Figs. 6A, 6B, 9E and 9F). Bioclastic grainstones have predominantly moldic and intercrystalline porosity (Figs. 3C and 10A). Compaction patterns tend to be controlled by the pore type. Interparticle porosity preservation favors porosity reduction by particle rearrangement and deformation, while bioclast molds can collapse during burial. Both dolomitized rudstones and grainstones are rich in intercrystalline porosity (Fig. 6E).

Moldic and intraparticle porosity due to dissolution of Mg-clay ooids (Fig. 3D and 7C) and intercrystalline porosity due to dolomitization are the most common type of porosity in ooidal arenites. Interparticle porosity is usually reduced by cementation and compaction, but remnants of primary porosity were observed in few samples (Fig. 10B). Silicified ooidal arenites show significant microcrystalline porosity (Fig. 10B) and localized vugs.

4.3.2 Macabu Formation

Growth-framework porosity in the interstices among coalescent fascicular calcite aggregates is the main type of primary porosity within calcite crusts of the Macabu Formation (Fig. 10C). Dolomite cement is the main reducer of growth-framework porosity (Figs. 8E). Pore enlargement is a common porosity enhancement within calcite crusts (Figs. 8E, 9C and 10C). Compaction does not significantly affect growth-framework porosity. Primary interparticle porosity occurs in intraclastic grainstones and rudstones and hybrid arenites (Figs. 5D, 8D and 10D), reduced by mechanical and chemical compaction.

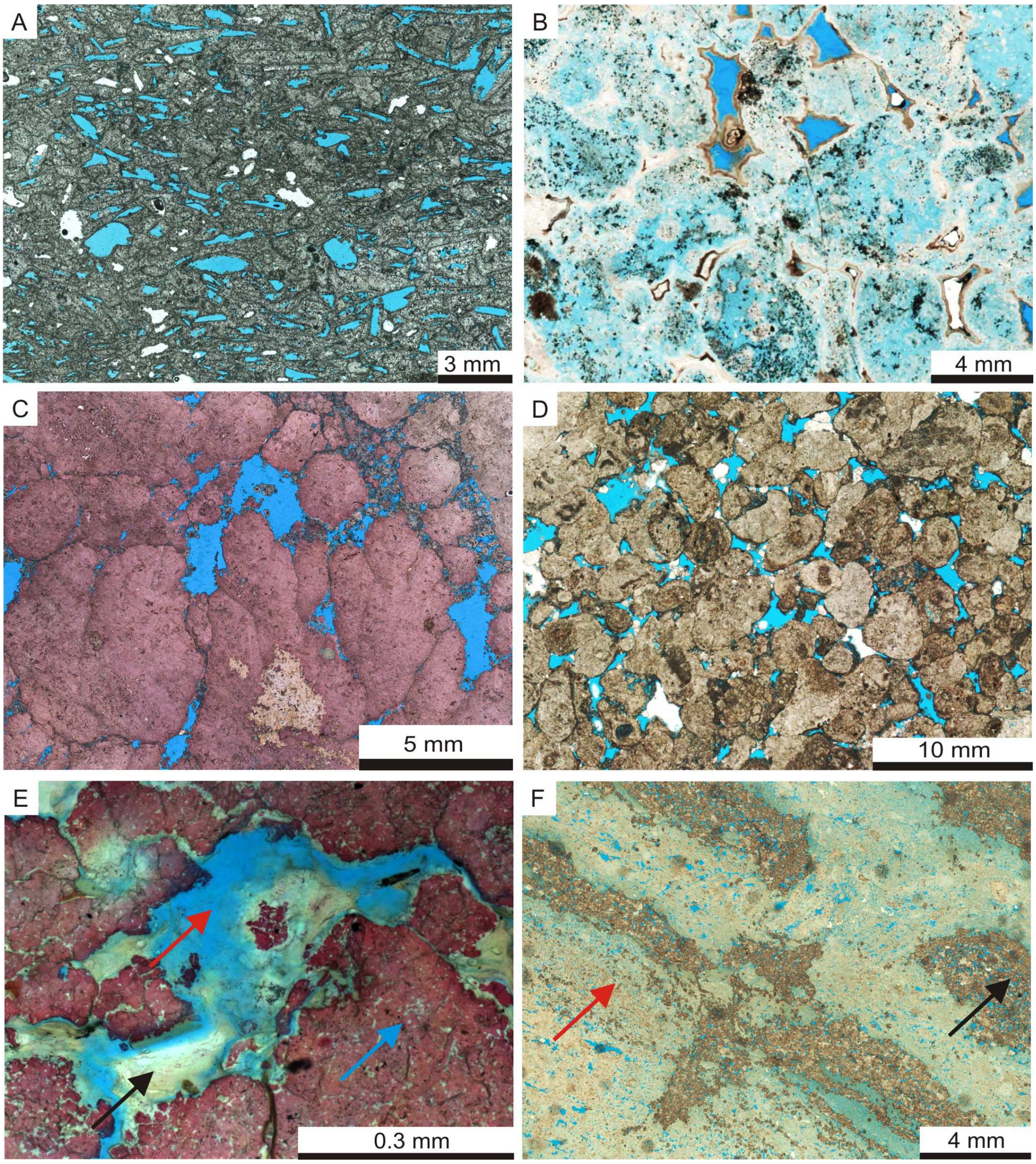


Figure 10. Common porous type on Coqueiros and Macabu Formation: (A) Moldic porosity in bioclastic grainstone (//P). (B) Microporous chert replacing ooidal arenite with remnants of primary porosity (//P). (C) Primary growth-framework porosity within calcite crust reduced by dolomite filling, followed by partial dissolution of cement and pore enlargement (//P). (D) Interparticle porosity in intraclastic grainstone (//P). (E) Mg-clay (black arrow) partially dissolved

(red arrow) replaced by calcite (blue arrow) (//P). (F) Bioturbated laminations exhibiting intercrystalline porosity within dolostone laminae (black arrow), and vug-microcrystalline in chert laminae (red arrow).

Porosity generated by partial dissolution and contraction of syngenetic clay matrix occasionally occurs within preserved Mg-clays (Fig. 10E). Intracrystalline porosity caused by dissolution of dolomite that filled interparticle and growth-framework porosity is common (Figs. 8E and 9C). Intercrystalline porosity is observed in dolostones and sparites. Microporosity was observed in SEM and optical petrography in several syngenetic and diagenetic phases, including calcite crusts (Fig. 5A), dolomite and silica (Fig. 10B). Significant microporosity is indicated by nuclear magnetic resonance (NMR) logs within intervals rich in magnesian clays. Vugs are commonly seen in cherts (Fig. 10F), and scarcer in other facies.

4.4 Petrophysical Analysis

4.4.1 Coqueiros Formation

The results of petrophysical analyses of the Coqueiros Formation show significant variability among the bioclastic reservoirs, the Mg-clay ooidal arenites and the dolostones (Table 1). The overall porosity ranges between 1.8 and 31.2 % (mean = 11.54 %) and the mean permeability is 13.43 mD (maximum = 280 mD). Bioclastic rudstones show significant difference in permeability according to the predominant type of porosity. The porosity of rudstones with interparticle porosity varies between 1.8 and 14.6 % (mean = 8.68 %) and mean permeability is 7.2 mD (maximum = 74.7 mD). Rudstones with moldic and intraparticle porosity have higher porosities than those with predominantly interparticle porosity (mean = 10.32 %; 5.1-25.8 %), but lower permeability (mean = 1.1; <0.001-10.3 mD). Rudstones with predominant vugular porosity have the lowest permeability and porosity values (mean = 7.91%; 2.8-15.1 %; mean permeability = 0.14 mD; <0.001-0.76 mD). The porosity of bioclastic grainstones with predominantly moldic porosity range between 8.5 and 19.9% (mean = 14.46 %) and their average permeability is

1.54 mD (0.019-5.97 mD). Mg-clay ooidal arenites have significant variability in porous types, as previously discussed, which corresponds to the large variation in porosity and permeability of Coqueiros Formation. Their porosity varies between 5.1 and 31.2 % (mean = 18.74 %) and the mean permeability is 21.4 mD (<0.001-267 mD). Dolostones have porosities between 10.1 and 23.1 % (mean = 16.6 %), and permeabilities ranging between 13.5 and 280 mD (mean = 87.18 mD).

4.4.2 Macabu Formation

The samples from the Macabu Formation show high variability of permeability and porosity (Table 2). Their average porosity is 6.22 % (0.1-37 %), and their average permeability is 17.29 mD (<0.001-1193 mD). About 40 % of the samples are not considered reservoirs, including syngenetic argillaceous rocks, hybrid siltstones, microbial laminites, pisolithic rudstones, hybrid “packstones”, cherts and sparites, which usually show low porosities and permeabilities (<0.001 mD). Well-developed coalescent calcite crusts have porosities ranging between 0.20 and 16.10 % (mean = 6.85 %) and permeability average of 28.94 mD (<0.001-295 mD), while poorly-developed non-coalescent crusts have porosity between 1.30 and 13.6 % (mean = 6.14 %) and average permeability of 0.8 mD (<0.001-16.2 mD). Intraclastic grainstones and rudstones have average porosities of 6.34 % (0.1-19 %) and average permeabilities of 10.45 mD (<0.001-523 mD). Dolostones show the best values of porosity and permeability, with porosity ranging between 1.3 and 32.4 % (mean = 15.27 %) and maximum permeability reaching 1193 mD (mean = 126.94 mD).

4.5 Mercury Injection Capillary Pressure

4.5.1 Coqueiros Formation

Mercury injection capillary pressure (MICP) analyses of bioclastic rudstones and grainstones showed throat pore mean of 4.51 μm (median = 3.05 μm) with multiple modes, reflecting the heterogeneity of the porous system. The distribution of pore sizes shows predominance of microporosity (micro = 47.1; meso = 28.7; macro = 24.2 %), according to Hassall *et al*. (2004) classification. The dolostones from the Coqueiros Formation show better permeability conditions, considering pore throats (median = 5.75; mean = 6.32 μm) and predominance of macroporosity (micro = 8.92; meso = 42.23; macro = 48.85 %).

4.5.2 Macabu Formation

The results of MICP indicate high variability of pore throats in the studied samples. Coalescent calcite crusts often show multiple modes, reflecting the heterogeneous distribution of diagenetic processes. The distribution of pore sizes presents a wide variation (micro = 37.5; meso = 31; macro = 31.5 %). The average pore throats diameter is 5.83 μm (median = 2.96 μm). Non-coalescent crusts tend to have narrower throats (mean = 1.92 μm; median = 0.52 μm) with dominance of micro and mesoporosity (micro = 51.14; meso = 40.49; macro = 8.38 %). Most of dolostones samples tend to have a narrow distribution with only one mode and average pore throat of 3.74 μm (median = 2.54 μm). The distribution of pore throat shows wide prevalence of mesoporosity (micro = 24.7; meso = 52.0; macro = 23.2 %).

5 DISCUSSION

5.1 Diagenetic Evolution and Paragenesis

The diagenetic evolution of the main studied rock types was determined according to the paragenetic relationships between diagenetic mineral phases and primary constituents, among each other, in relation to mechanical and chemical compaction, and with porosity, as observed in optical microscopy, CL, UVF, BSE and SEM. The studied samples show intense modifications during early diagenesis, owing to the instability and reactivity of primary constituents, in both the Coqueiros and the Macabu Formation.

Table 1. Statistical summary of petrophysical analysis from the Coqueiros Formation.

Petrofacies	Bioclastic Rudstones						Bioclastic Grainstones		Ooidal Arenites		Dolostones		All Samples	
Predominant Pore Type	Interparticle		Moldic		Vug		Moldic		Multiple Types		Intercrystalline			
	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)
Mean (Arithmetic)	8.68	7.209	10.32	1.055	7.91	0.142	14.46	1.540	18.74	21.400	16.60	87.188	11.54	13.431
Median	8.60	1.760	9.60	0.198	7.20	0.078	14.65	0.561	17.95	0.432	16.80	72.000	10.10	0.534
Standard Deviation	2.34	15.348	4.27	2.427	2.85	0.185	3.21	1.961	7.22	62.825	4.06	71.990	5.45	39.944
Variance	5.46	235.548	18.20	5.888	8.13	0.034	10.28	3.846	52.18	3947.036	16.52	5182.532	29.66	1595.485
Kurtosis	1.10	13.235	4.58	11.087	1.58	4.024	-0.81	1.006	-0.57	15.983	-0.95	2.095	1.32	23.789
Range	12.80	74.699	20.70	10.299	12.30	0.759	11.40	5.951	26.10	266.999	13.00	266.500	29.40	279.999
Minimum	1.80	<0.001	5.10	<0.001	2.80	<0.001	8.50	0.019	5.10	<0.001	10.10	13.500	1.80	<0.001
Maximum	14.60	74.700	25.80	10.300	15.10	0.760	19.90	5.970	31.20	267.000	23.10	280.000	31.20	280.000
n	45	45	31	31	27	27	18	18	18	18	16	16	163	163

Table 2. Statistical summary of petrophysical analysis from the Macabu Formation.

Petrofacies	Coalescent Crust		Non-Coalescent Crust		Grainstones/Rudstones		Dolostones		All Samples	
Predominant Pore Type	Growth-framework		Intercrystalline		Interparticle		Intercrystalline			
	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)	Ø (%)	k (mD)
Mean (Arithmetic)	6.85	28.937	6.14	0.809	6.34	10.445	15.27	126.942	6.22	17.292
Median	7.00	0.435	5.50	0.055	6.00	0.010	15.10	12.850	4.95	0.011
Standard Deviation	4.09	64.646	3.47	2.400	4.68	61.224	9.06	291.946	5.18	87.071
Variance	16.69	4179.164	12.07	5.759	21.91	3748.384	82.06	85232.246	26.79	7581.391
Kurtosis	-1.25	6.818	-1.14	28.798	-0.57	66.974	-0.99	9.293	5.81	104.253
Range	15.90	294.999	12.30	16.199	18.90	522.999	31.10	1192.999	36.90	1192.999
Minimum	0.20	<0.001	1.30	<0.001	0.10	<0.001	1.30	<0.001	0.10	<0.001
Maximum	16.10	295.000	13.60	16.200	19.00	523.000	32.40	1193.000	37.00	1193.000
n	105	105	67	67	77	77	24	24	436	436

5.1.1 Coqueiros Formation

The paragenetic evolution in bioclastic rudstones and grainstones ("coquinas"; Fig. 11) seems to differ from equivalent rocks in other regions of the basin, where there is a predominance of aragonite neomorphism in relation to dissolution (Bertani and Carozzi, 1985b; Altenhofen, 2013; Muniz, 2014), reflecting different diagenetic conditions.

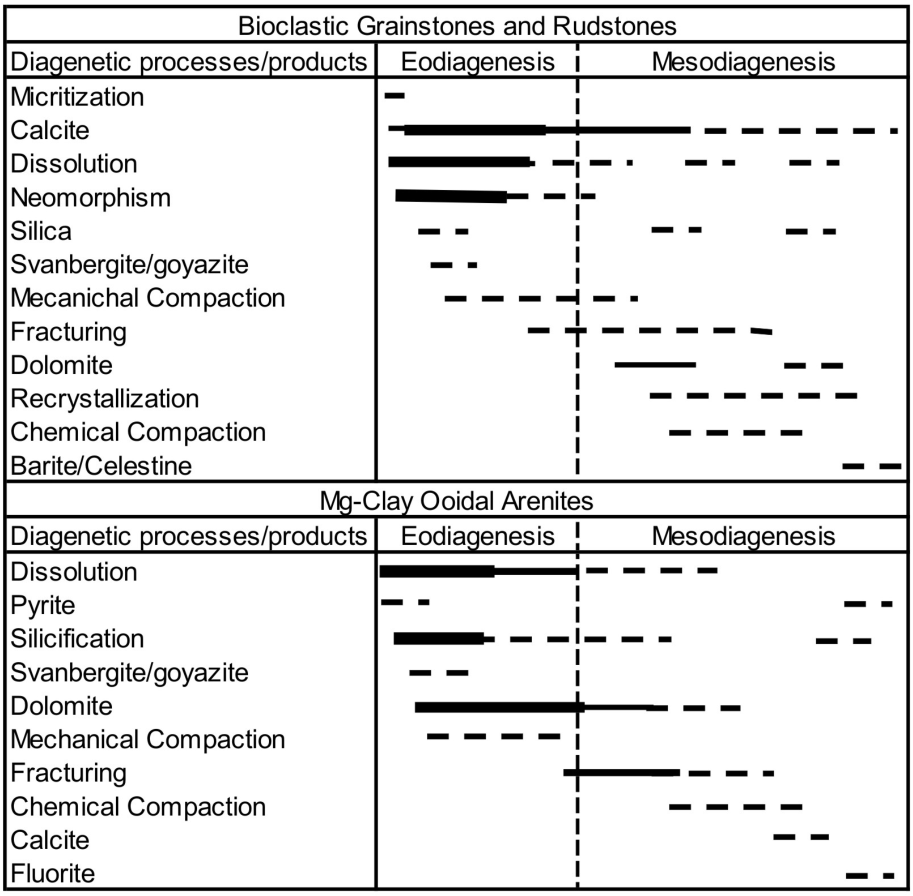


Figure 11. Diagenetic sequence for bioclastic grainstones and rudstones and Mg-clay ooidal arenites.

The bioclastic rocks show three main distinct evolutionary pathways, controlled by eodiagenetic conditions (Fig. 12). The circulation of meteoric groundwater promoted the dissolution of aragonitic bivalves (*cf.* Morse and Mackenzie, 1990; Moore and Wade, 2013), as well as the precipitation of calcite prismatic rims in interparticle and intraparticle pores (Fig. 6B). The early preservation of such bioclasts occurred under stagnant eodiagenetic conditions (*cf.* James and Choquette, 1984). Conditions for the precipitation of calcite prismatic rims, followed by total dissolution of bioclasts and occlusion of interparticle porosity, preserved moldic porosity (Fig. 10A). In slightly supersaturated environments, thin calcite rims were formed, interparticle calcite precipitation was inhibited, favoring neomorphism and limited dissolution (Fig. 6A and 9E). Typical vadose features (*cf.* James and Choquette, 1984) were not observed, which may indicate that exposure was not significant.

The common association of neomorphism and dissolution indicates gradation between the interpreted environmental conditions, which controls may be tectonic and/or climatic. Well 2 shows clear cyclical variation along the studied interval, suggesting a possible climatic control. However, the porosity types distribution could not be correlated with other wells. Well 3 shows predominance of moldic porosity, and Well 5 of interparticle porosity. Well 1 was not sampled in Coqueiros Formation and Well 3 was poorly sampled. The generation and destruction of accommodation space and/or variation in hydraulic gradient, probably driven by fault movements may have differently affected the bioclastic reservoirs in different regions of the studied area, generating patterns which cannot be easily correlated among the studied wells.

The lack of typical exposure features suggests that the observed silicification cannot be directly related to the formation of groundwater silcretes under arid/semi-arid conditions (Ross and Chiarenzelli, 1985; Khalaf, 1988; Hesse, 1989; Murray, 1990). Since the bioclastic deposits are poor in siliciclastics, it is possible that the silica was derived from dissolution of the magnesian clay ooids.

The record of mesodiagenetic processes is highly heterogeneous. In many cases, mesodiagenesis was incipient, and in other cases it completely destroyed the primary fabrics and constituents and/or the eodiagenetic constituents. Cementation by calcite filling remaining interparticle and intraparticle porosity is the most abundant constituent. Dolomite observed in bioclastic rocks is essentially mesodiagenetic. Blocky and saddle dolomite occurs mainly filling moldic, interparticle or vugular porosity. Mimetic pervasive dolomitization is frequent, preserving part of moldic and/or interparticle porosity. Saddle dolomite often filled such porosity in dolostones, indicating that the precipitation of dolomite occurred recurrently during mesodiagenesis. Late quartz occasionally replaced primary and diagenetic constituents or filled remaining porosity. Recrystallization of early calcite to microcrystalline or blocky calcite is common, destroying primary or eodiagenetic modified fabric.

Celestine, barite and anhydrite occur as late cements filling interparticle porosity, or replacing diagenetic phases and mud intraclasts. Circulation of fluids strongly influenced by the dissolution of Retiro Formation evaporites could favor the precipitation of these minerals, even though these minerals are scarce in overlaid Macabu Formation. On the other hand, these minerals are related to fracturing, indicating that these sulfates can be associated to hydrothermalism. Traces of svanbergite and goyazite occur replacing mud intraclasts, probably derived from the interaction between aluminum from the clays, Sr^{+2} from aragonite dissolution and available SO_4^{-2} on interstitial waters. The source of PO_4^{-2} cannot be easily explained. Detrital apatite and phosphatic bioclasts are not commonly observed in bioclastic rocks of the Coqueiros Formation.

Scarce mesodiagenetic dissolution of eodiagenetic constituents was observed, what enhanced porosity through generation of channel and vugular pores. Such dissolution may have been generated by acid pore waters related to hydrocarbon migration from rift shales (Moore, 2001; Racey *et al*., 2001; Beavington-Penney *et al*., 2008). “Exotic” mineralogy, such as epidotes, prehnite, amphiboles and metallic sulphides, was not observed within channel and vugular porosity, indicating that the influence of hydrothermal fluids probably was not an important control on mesodiagenetic dissolution at pore scale (Bakalowicz *et al*., 1987; Hulen *et al*., 1994). Brittle mechanical compaction by collapse of moldic pores and fractured neomorphised particles is occasionally observed. Ductile deformation occurs preferable in neomorphised particles. Stylolites are uncommon, probably because extensive chemical compaction was inhibited by the presence of rims and by the heterogeneous framework of the bioclastic rocks.

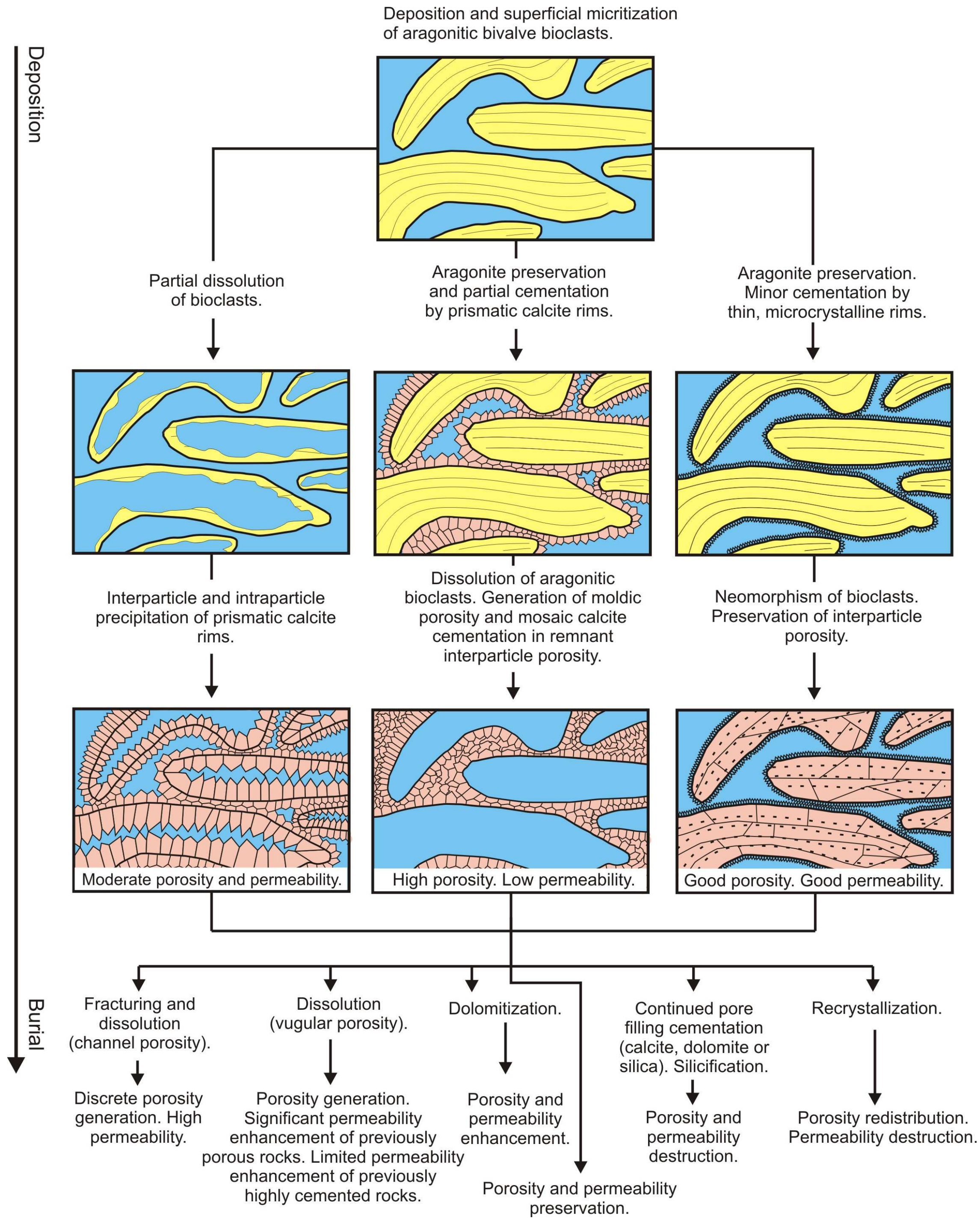


Figure 12. Schematic diagenetic pathways of bioclastic rudstones and grainstones from Coqueiros Formation.

The diagenesis of the Mg-clay ooidal arenites is controlled by the high reactivity of primary constituents. According to Tosca and Wright (In press), the high sensitivity to $p\mathrm{CO_2}$, pH and fast dissolution kinetics due to the high specific surface area favors the eodiagenetic dissolution of Mg-clays. As a result, their dissolution, silicification and dolomitization are commonly observed. Their dissolution generates abundant moldic and intraparticle porosity (Figs. 3D and 7C). Intense dissolution, silicification and dolomitization are observed in Mg-clay ooidal arenites just below the pre-Alagoas regional unconformity, what could suggest leaching and formation of silcretes and dolocretes in a meteoric vadose and phreatic environment during post-rift uplift and exposure. Evidences of vadose cements were, however, not found.

5.1.2 Macabu Formation

The diagenetic evolution of the Macabu Formation reservoirs was strongly influenced by the reactivity of syngenetic Mg-clay minerals and their interaction with the syngenetic crusts of calcite (Figs. 13 and 14). Previous studies suggested that Mg-clays may evolve from a Si-Mg hydrated gel precursor (Tosca and Wright, 2014; Wright and Barnett, 2015), which would evolve to stevensite, kerolite, talc and other Mg-clay mineral phases (Rehim *et al*., 1986; Pozo and Casas, 1999). Experimental studies showed that the stability and evolution of such gel depends upon a number of environmental factors, including pH, salinity, temperature, and ionic content (Jones, 1986; Tosca and Masterson, 2014).

The most abundant eodiagenetic constituents are spherulites of calcite, which occur throughout the Macabu Formation (Figs. 8A, 8B and 8C). The formation of spherulites is interpreted to be favored in the interior of inorganic gels (Beck and Andreassen, 2010; Shtukenberg *et al*., 2012). Petrographic evidence shows that the development of spherulites occurs inside the syngenetic or hybrid matrix, both displacing and replacing the protolith, indicating early diagenetic conditions. Although it is reported in several studies that the development of calcite spherulites is favored by high levels of silica and magnesium in alkaline environments (García-Ruiz, 2000; Meister *et al*., 2011), diagenetic spherulites of different compositions are found in many geological contexts, including sandstones, with highly variable dimensions (Browne and Kingston, 1993; McBride *et al*., 2003; Rodrigues *et al*., 2015). Even though few of the spherulites show nuclei with microcrystalline textures that could correspond to mineralized microbial colonies, many show no nuclei, or several other types of materials as nuclei, such as clay peloids, siliciclastic grains and ostracod bioclasts. Natural and synthetic spherulites interpreted as formed by direct microbial activity (Braissant *et al*., 2003; Spadafora *et al*., 2010) have sub-micrometric dimensions, differing significantly from those found in the Macabu Formation. Replacement of the syngenetic matrix by dolomite commonly occurred after the precipitation of the spherulites (Figs. 8A and 8B).

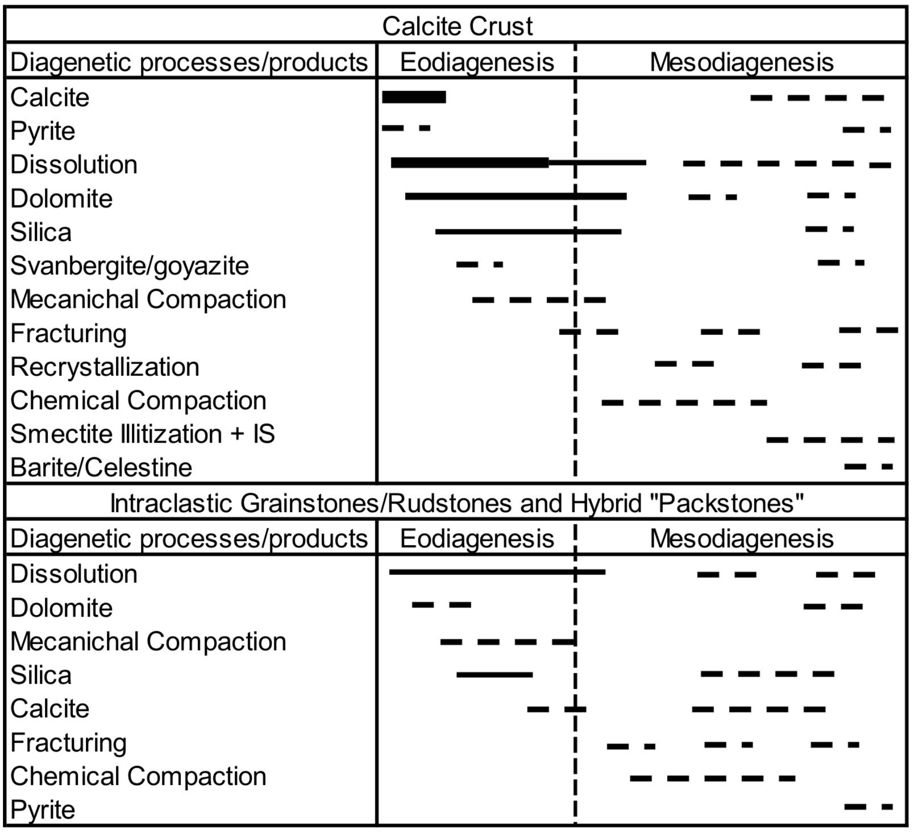


Figure 13. Diagenetic sequence for calcite crusts and associated syngenetic matrix and intraclastic rocks from Macabu Formation.

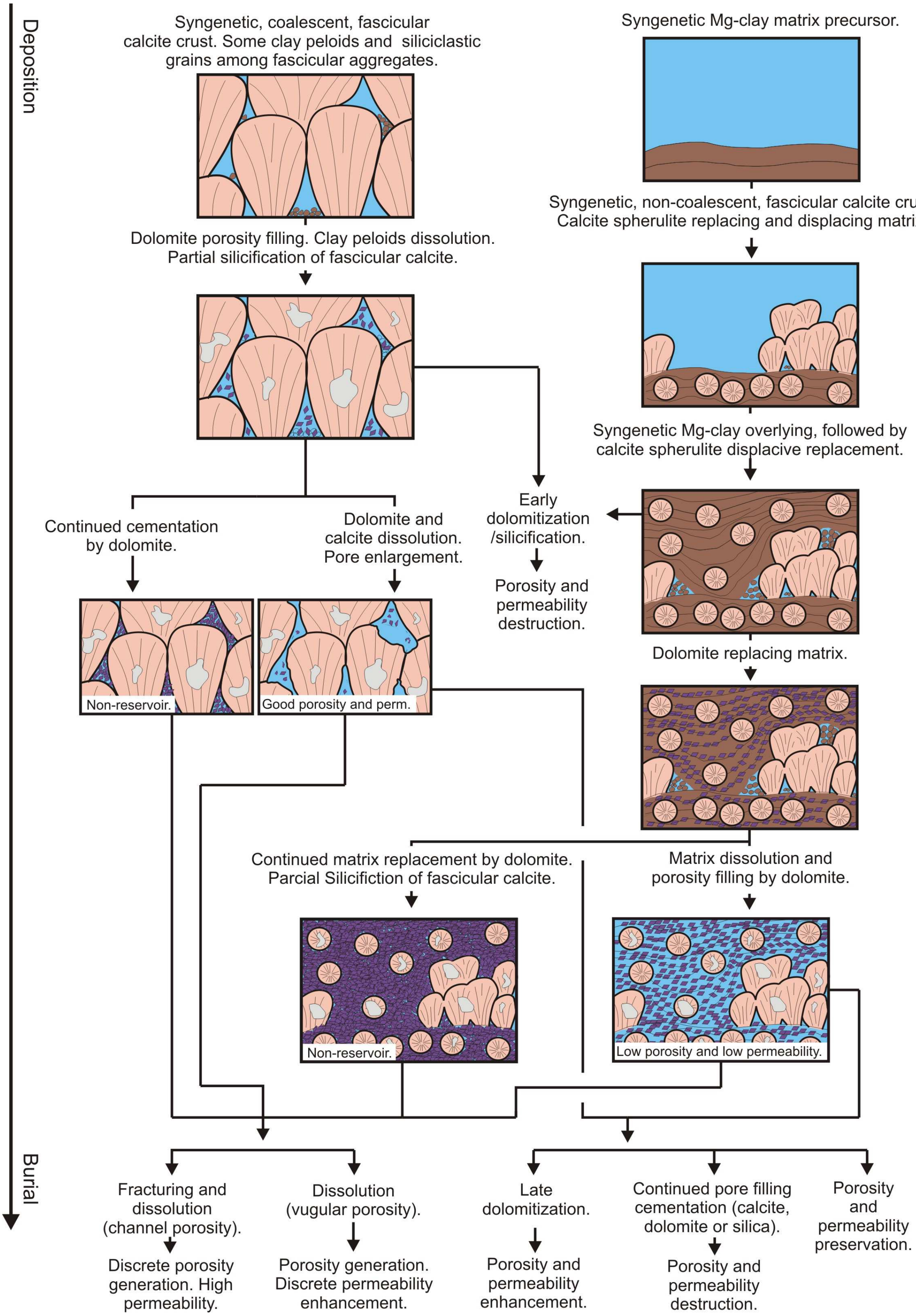


Figure 14. Schematic evolution of calcite crusts and their relationship with syngenetic matrix.

Apparently, part of the fascicular-optic calcite aggregates was precipitated replacing the background argillaceous sediments, instead of only growing on them as crusts. This is indicated by the occurrence of substantial amounts of Mg-clay peloids, laminations and other particles in the interstices and as inclusions in some shrubs. The truncation of clay laminations by some shrubs indicates that they grew as diagenetic replacement. Conversely, non-replaced clay peloids are very commonly included within the crusts. Therefore, the precipitation of syngenetic calcite and Mg-clay apparently occurred simultaneously in some cases. The combined evidences indicate that the abiotic precipitation of syngenetic calcite crusts and of Mg-clays was chemically controlled by the dynamic evolution of the lacustrine environmental conditions.

The dissolution of the syngenetic Mg-clay laminations and peloids/ooids probably began at the lake bottom, continuing during eodiagenesis, controlled by variations on geochemistry of the lacustrine waters. In most cases, the clay laminations and peloids were replaced by silica or dolomite, which were derived from the dissolution of the Mg-clays. Svanbergite/goyazite occurs as traces associated with early silica, although they have been described in association with late dolomitization.

Besides replacing the Mg-clays, eodiagenetic dolomite is observed filling primary porosity in the calcite crusts (Figs. 8E e 9C) and in intraclastic grainstones and rudstones. In some cases, the precipitation of dolomite continued during mesodiagenesis, as indicated by optical and CL microscopy. Total dolomitization is a common process in the Macabu Formation that can increase or destroy porosity. Porosity enhancing dolomitization apparently occurred during mesodiagenesis, since relicts of eodiagenetic components can be observed. Overdolomitization (*sensu* Halley and Schmoker, 1983) is more likely to have been developed during eodiagenesis, due to the high availability of magnesium in the lacustrine environment.

Coarsely-crystalline quartz and calcite that cement pores or replace primary and eodiagenetic constituents, were heterogeneously precipitated during mesodiagenesis. Sulfates and part of APS minerals were derived either from the dissolution of overlying evaporites, or from hydrothermal fluids, as discussed in the next section. Coarse pyrite that replaced primary and diagenetic phases was related to thermal sulfate reduction (Machel *et al*., 1995; Cai *et al*., 2001). The enlargement of pores in the calcite crusts, accompanied by partial dissolution of dolomite cements (Fig. 10C), is ascribed to mesodiagenetic dissolution. Intracrystalline dissolution on dolomite crystals is common (Figs. 8E and 9C). Stylolites are commonly cutting throughout syngenetic matrix, with organic matter residue concentrated along the surfaces of dissolution. Compaction has not significantly affected the calcite crusts. Illite-smectite interstratified mesodiagenetic clays were formed in small amounts, commonly in the interstices of crusts and replacing detrital matrix.

Intraclastic grainstones and rudstones of the Macabu Formation were formed by fragmentation of the calcite crusts (Fig. 5D) and erosion and re-deposition of spherulites and Mg-clay ooids. Most of the recrystallization of intraclasts and replacement of ooids by microcrystalline calcite probably occurred during eodiagenesis. Two main eodiagenetic patterns of relationship between compaction and cementation were recognized (Fig. 15): grainstones without significant cementation and grainstones partially cemented by dolomite, calcite or silica. Rocks poorly cemented during eodiagenesis developed ductile and chemical compaction, followed by mesodiagenetic calcite cementation. The heterogeneous eodiagenetic cementation by calcite or dolomite promoted the preservation of primary porosity, inhibiting pressure solution. Some samples shows eodiagenetic precipitation of displacive silica, filling the porosity and inhibited the mesodiagenetic processes. Mesodiagenetic dissolution of particles after compaction is locally observed.

Hybrid “packstones” mostly formed by intraclastic particles and clay matrix occasionally show early dissolution of the matrix, enabling a diagenetic evolution similar to that observed in the grainstones (Fig. 15). The distinction between grainstones and hybrid “packstones” with dissolved matrix is often difficult.

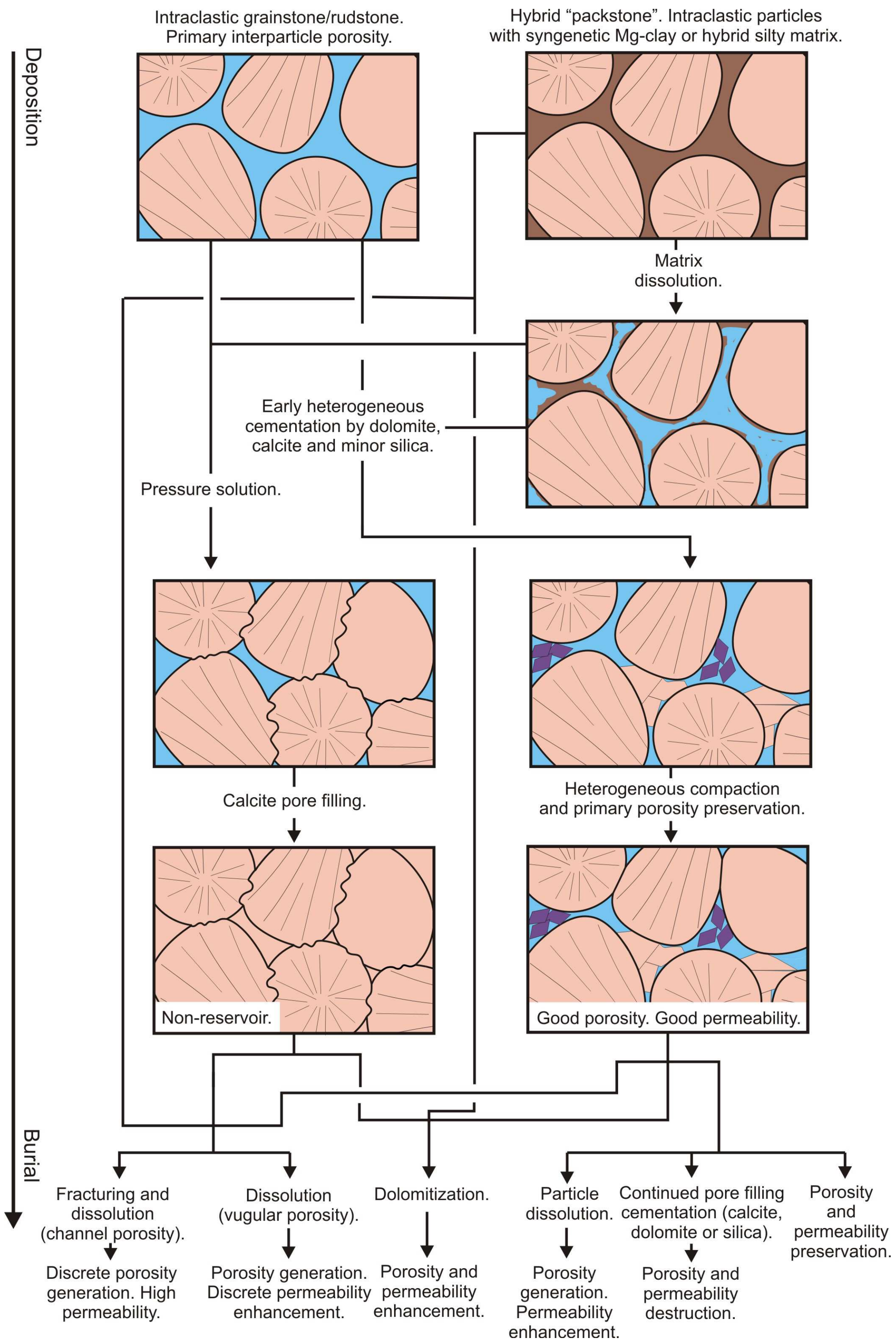


Figure 15. Schematic diagenetic evolution of intraclastic rocks from the Macabu Formation.

5.2 Hydrothermalism

Although lacking sampling of the most affected intervals, areas with intense dissolution were interpreted in well logs and formation tests as probably related to hydrothermal karst (*cf.* Dublyansky, 1995; Heward *et al*., 2000). Even though not representative in volume, such areas may be extremely important for the flow of fluids and well production. These dissolution zones are characterized by intense fracturing, silicification, dolomitization, as well as concentration of barite, fluorite, kaolinite/dickite, pyrite and other sulfides, indicating the hydrothermal influence on these areas.

5.3 Porosity and Permeability of Reservoir Petrofacies

The evolution of porosity and permeability on the studied rocks was strongly controlled by facies, original fabric and composition, which conditioned the impact of mainly eodiagenetic processes on the quality of the Lagoa Feia Group reservoirs.

5.3.1 Coqueiros Formation

The quality of bioclastic reservoirs from the Coqueiros Formation was directly controlled by the evolution of the aragonitic bioclasts during eodiagenesis, which conditioned the types of pores generated (Fig. 12). Their porosity was strongly controlled by the pattern of dissolution of aragonite and precipitation of low-Mg calcite, either as neomorphism, or as dissolution-cementation. The porosity of the ooidal arenites was controlled by the dissolution of the Mg-clays and their replacement by silica and dolomite. Seven reservoir petrofacies were defined for the studied Coqueiros Formation lithologies, based on the depositional facies, diagenetic patterns and relationship between porosity and permeability, according to the De Ros and Goldberg (2007) concept.

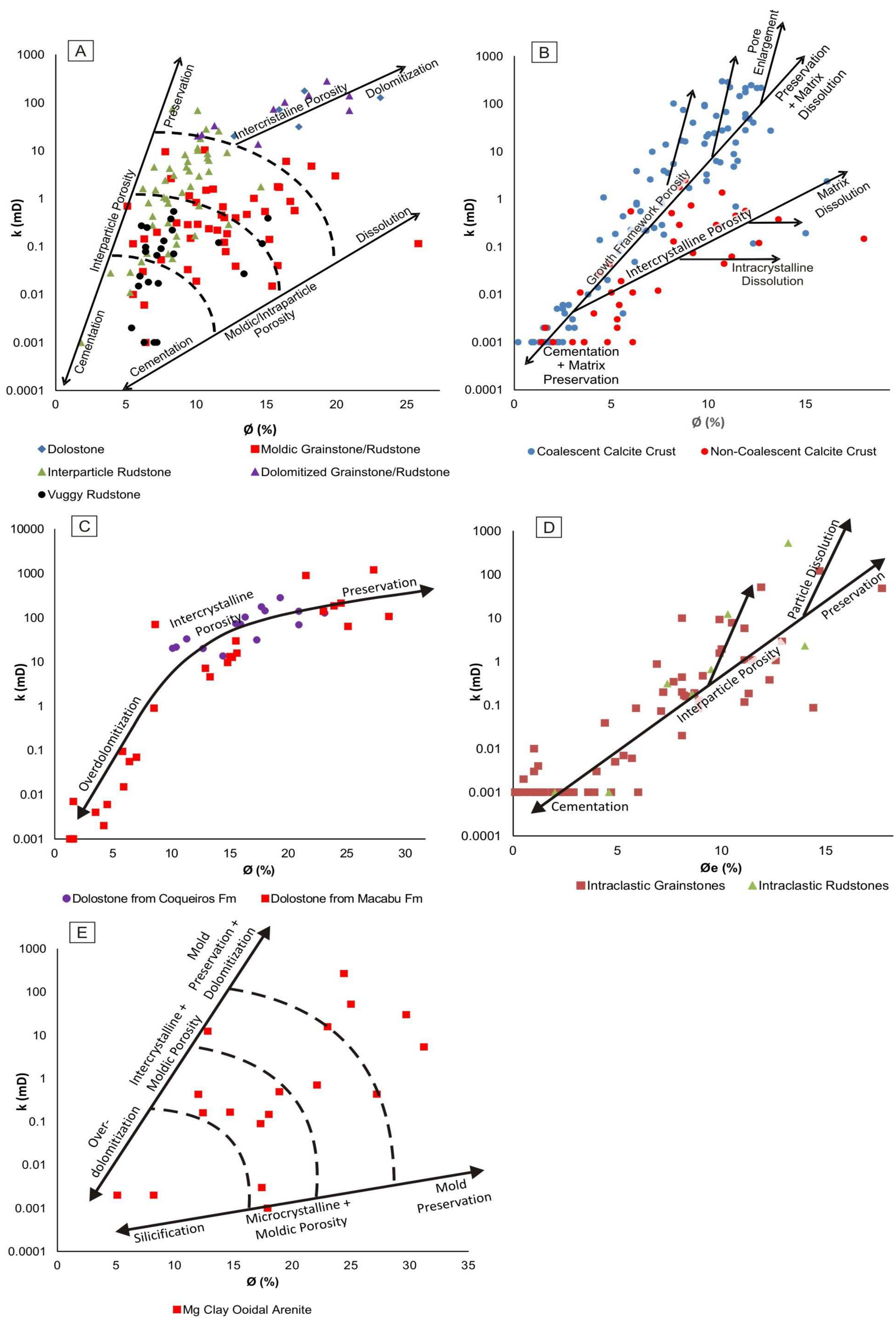

A
k (mD)
Ø (%)
Preservation
Interparticle Porosity
Cementation
Intercristaline Porosity
Dolomitization
Dissolution
Moldic/Intraparticle Porosity
Cementation
Dolostone
Moldic Grainstone/Rudstone
Interparticle Rudstone
Dolomitized Grainstone/Rudstone
Vuggy Rudstone
B
k (mD)
Ø (%)
Pore Enlargement
Preservation + Matrix Dissolution
Growth Framework Porosity
Intercrystalline Porosity
Matrix Dissolution
Intracrystalline Dissolution
Cementation + Matrix Preservation
Coalescent Calcite Crust
Non-Coalescent Calcite Crust
C
k (mD)
Ø (%)
Intercrystalline Porosity
Preservation
Overdolomitization
Dolostone from Coqueiros Fm
Dolostone from Macabu Fm
D
k (mD)
Øe (%)
Particle Dissolution
Preservation
Interparticle Porosity
Cementation
Intraclastic Grainstones
Intraclastic Rudstones
E
k (mD)
Ø (%)
Mold Preservation + Dolomitization
Intercrystalline + Moldic Porosity
Over-dolomitization
Mold Preservation
Microcrystalline + Moldic Porosity
Silicification
Mg Clay Ooidal Arenite

Figure 16. Correlation between porosity and permeability, showing the effect of diagenesis on porous type and permeability through studied reservoir petrofacies: (A) Bioclastic rudstones/grainstones. (B) Calcite crusts. (C) Dolostones. (D) Intraclastic rudstones/grainstones. (E) Mg-clay ooidal arenites.

Bioclastic rudstones with interpaticle porosity. Preservation of interparticle porosity occurred mostly where neomorphism predominated over dissolution (Figs. 6A and 9E). This petrofacies shows the best permeability conditions among the bioclastic reservoirs, even considering the samples with lower porosity (Fig. 16A, Table 1). MICP analyses indicate that the pore throats are significantly larger than in others petrofacies (Fig. 17A). In some cases, however, continued cementation narrowed the pore throats, decreasing permeability. Subordinated moldic pores increased the porosity, but with little effect on permeability. On the other hand, the local development of vugular pores enlarged the primary porosity enhancing permeability.

Bioclastic rudstones with moldic porosity. Bioclastic rocks with predominantly moldic or intraparticle porosity, combined with some interparticle porosity are common (Fig. 6B). These rocks usually have good porosity, but low permeability (Table 1, Fig. 16A). The cementation of the moldic and intraparticle pores is common, occasionally totally occluding the secondary porosity. The remaining interparticle pores contribute to the permeability, but as pore throats are usually controlled by microcrystalline and intercrystalline pores (Fig. 17A), permeabilities generally are low. Locally, vugular porosity was formed by enlargement and connection of molds, enlarging pore throats (Fig. 9F) and generating fair permeability conditions in previously low permeability rocks.

Bioclastic rudstones with vugular porosity. Vugular porosity is commonly scarce in the studied rocks. Where vugs are the predominant pore type, the porosities and permeabilities are generally low (Table 1, Fig. 16A), owing to the low connectivity of the pore system. The presence of remnants of interparticle porosity usually contributes to some permeability, though generally low. However, as cores with high vugular porosity were probably poorly recovered and sampled for thin sections preparation, and not suitable for petrophysical analyses, it is possible that the amount of vugular porosity was underestimated in the reservoirs due to these sampling limitations.

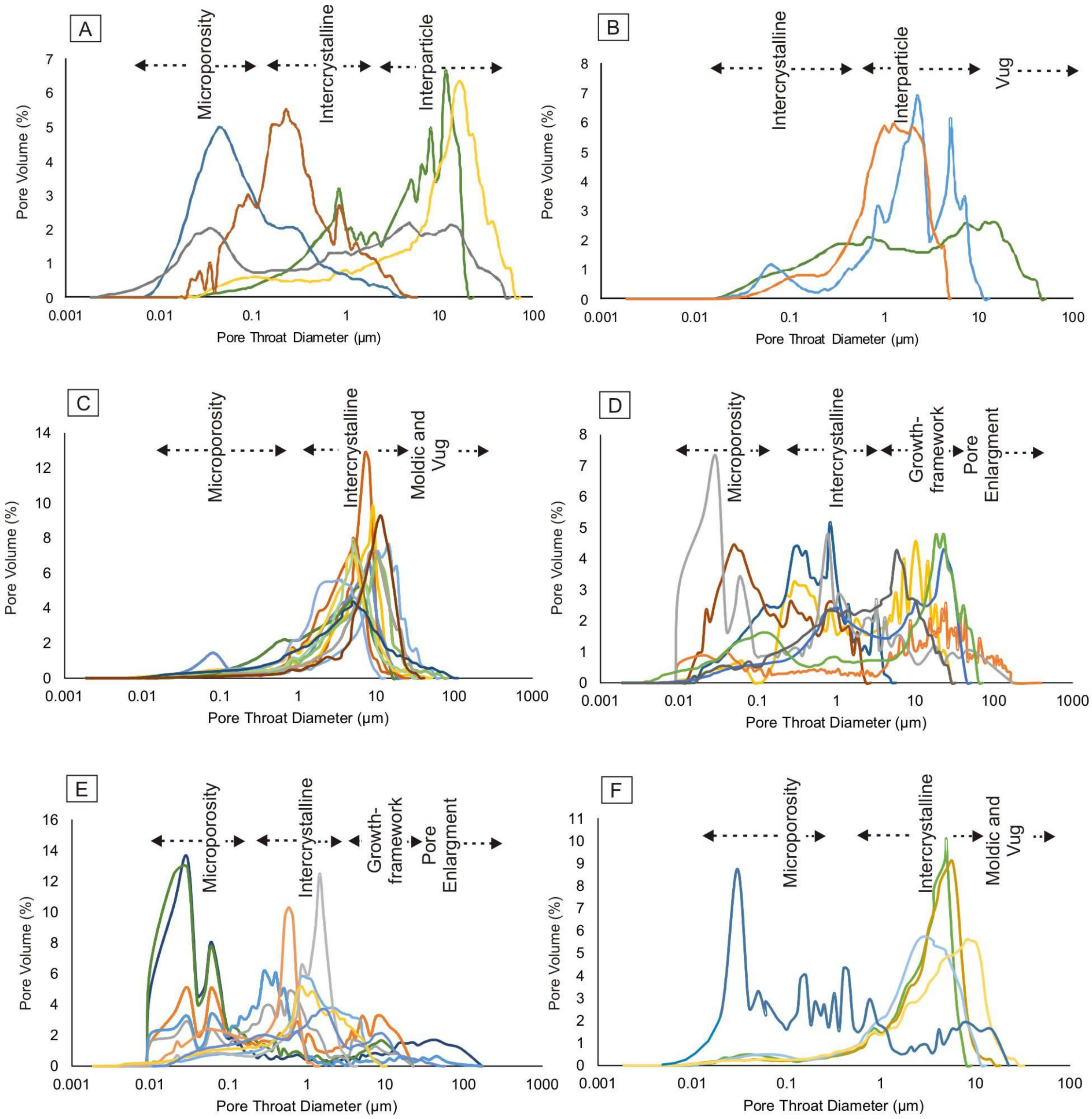


Figure 17. Distribution of pore throats of studied samples and their probable relationship with the pore type: (A) Bioclastic rudstones. (B) Bioclastic grainstones. (C) Dolostones from Coqueiros Formation. (D) Coalescent calcite crust. (E) Non-coalescent calcite crust. (F) Dolostones from Macabu Formation.

Bioclastic grainstones with moldic porosity. All the described grainstones show high content of moldic porosity, and the higher total porosity among the bioclastic petrofacies (Table 1, Fig. 16A). These rocks generally present remaining interparticle porosity, what gives them some permeability. Vugular porosity was generated from enlarged moldic pores, what corresponds to widened pores throats, increasing permeability. Often the grainstones present recrystallization, which generated intercrystalline porosity, but with narrow pore throats (Fig. 17B) and low permeability.

Sparites. Fully recrystallized bioclastic grainstones and rudstones occasionally occur and show redistribution of primary interparticle or secondary porosity as intercrystalline porosity and microporosity. This petrofacies has general low reservoir quality.

Dolostones. Dolomitization was a very important process for the quality of some reservoirs, generating rocks with the best porosity and permeability of the Coqueiros Formation (Table 1, Fig. 16C). In many cases, mimetic dolomitization added intercrystalline porosity, leaving relicts of previous primary interparticle or moldic/vugular secondary porosity. Rudstone protoliths with moldic and/or vugular porosity developed higher porosities and permeabilities after dolomitization through the connection of vugs and molds by intercrystalline porosity, which conditioned pore throat diameters (Fig. 17C). Late processes were limited to precipitation of minor amounts of saddle dolomite, preserving most of the intercrystalline porosity generated during dolomitization.

Mg-clays ooidal arenites. The high instability of Mg-clays (*cf.* Tosca and Wright, 2014) favored the dissolution of ooids and the precipitation of silica and dolomite (Figs. 3D and 7C), redistributing the original porosity and decreasing permeability. The pore system of ooidal arenites is extremely heterogeneous, generally presenting high porosities and variable permeabilities (Table 1, Fig. 16E). The dissolution of ooids, resulting in moldic and intraparticle porosity was important, generating significant increase in porosity. Microporosity within authigenic silica is commonly observed. Remnants of primary interparticle porosity are occasionally observed, although the original interstitial porosity of most of the samples was completely filled by dolomite and quartz. Dolomitization generated abundant intercrystalline porosity, occasionally connecting moldic and intercrystalline pores and supplying permeability to Mg-clay ooidal arenites.

5.3.2 Macabu Formation

The reservoir petrofacies from Macabu Formation differ significantly from those of the Coqueiros Formation. Most of the reservoirs correspond to calcite crusts or to intraclastic grainstones/rudstones composed mainly by fragments from crusts and spherulites. Five reservoir petrofacies were defined for the Macabu Formation, based on depositional facies, composition, fabric, diagenetic evolution and porosity:

Coalescent calcite crusts. Coalescent calcite crusts are composed of well-developed calcite shrubs, which coalesced vertically and/or horizontally (Fig. 4C). The *in situ* growth of these chemical precipitates generated a growth-framework primary pore system with moderate porosity and good permeability, controlled by the wide pore throats produced by the irregular precipitation process (Table 1, Fig. 16B). Rezende and Pope (2015) on relating aspects of the depositional texture, including shrub size, sorting and packing to the porosity and permeability of analog reservoirs from the Santos Basin, interpreted the rocks as microbialites. According to those authors, the size of individual shrubs exerts a primary control on pore size, affecting both porosity and permeability. Probably, these were major controls on the primary petrophysical properties of the crusts, but commonly they were strongly modified during diagenesis by cementation and/or dissolution. The dissolution of remnants of clay laminations or peloids filling interstices among the fascicular aggregates or in adjacent layers favored the interstitial precipitation of dolomite cementing partially or totally the porosity, which represented the main primary porosity reducer (Fig. 8E). Commonly, coalescent crusts present porosity enhancement through partial dissolution of shrubs and cements (Fig. 10C), which increased the porosity and widened pore throats, with consequent increase of permeability (Fig. 17D). Intracrystalline dissolution of dolomite crystals (Figs. 8E and 9C) or of recrystallized shrubs is very common.

Petrophysical porosity is recorded even in rocks without petrographic porosity, owing to microporosity from shrubs dissolution, observed in SEM analyses (Fig. 5A).

Non-coalescent calcite crusts. Poorly-developed calcite crusts with non-coalescent calcite shrubs show interstices occupied by syngenetic laminations and clay peloids, or even siliciclastic matrix, resulting in little growth-framework primary porosity (Fig. 14). As result, these rocks show poor primary porosity and low permeability. During diagenetic evolution, the syngenetic laminations and peloids were partially replaced mainly by dolomite. Later dissolution generated intercrystalline porosity. The dominance of discontinuous intercrystalline porosity plus minor growth-framework porosity generated low-quality reservoirs with mostly narrow pore throats (Figs. 16B and 17E). Most non-coalescent crusts have low porosities/permeabilities and do not constitute reservoirs. Nevertheless, even though these reservoirs present poor quality (Table 2), they probably contribute to the production of the wells.

Intraclastic grainstones/rudstones. Although most intraclastic rocks do not exhibit good reservoir quality, some show partially preserved primary porosity, representing good quality reservoirs (Table 2, Fig. 16D). Probably several of the intraclastic grainstones and rudstones, formed by reworked calcite crusts and spherulites had good primary porosity. Heterogeneous cementation preserved part of the primary porosity, while the porosity of uncemented rocks was completely obliterated during mesodiagenesis by interparticle pressure dissolution and derived calcite cementation (Fig. 15). No clear particle size control is observed on porosity, indicating that diagenesis was the major control. Secondary intraparticle porosity from intraclasts dissolution is occasionally seen in rocks with some preserved interparticle porosity. When moldic or vugular porosity occur subordinately to interparticle porosity, permeability is higher, generating the best reservoir conditions among the intraclastic rocks.

Hybrid “packstones”. Intraclastic rocks with clay matrix are usually not reservoirs. In some cases, however, early dissolution of matrix occurred, leaving residual dolomite and/or silica. Some of such petrofacies evolved similarly to grainstones, with secondary porosity partially preserved by heterogeneous cementation (Fig. 15). Another possibility for the occurrence of secondary porosity in hybrid “packstones” was through early heterogeneous dolomitization or silicification, followed by dissolution.

Dolostones. Dolostones of Macabu Formation show significant variability of distribution of porosity (Table 2, Fig. 16C) and pore throats diameters. The magnesium availability in the depositional and diagenetic environments was high, and the frequent overdolomitization generated microporosity, decreasing significantly the pore throats (Fig. 17F). Apparently, early dolomitization promoted severe porosity reduction, while mesodiagenetic dolomitization, associated with fracturing and precipitation of barite, fluorite, kaolinite/dickite, pyrite and other sulfides, of probable hydrothermal origin, generated high porosity and permeability.

6 CONCLUSIONS

The study based on the petrographic characterization and petrophysical analysis revealed that the reservoirs of the Lagoa Feia Group have complex pore systems, resulting in highly heterogeneous permeability patterns. The primary composition and porosity had strong influence in the evolution of quality of the reservoirs.

The compositional and porosity evolution of the bioclastic reservoirs from the Coqueiros Formation was controlled by the dissolution or neomorphism of the aragonitic bioclasts. Where the circulation and geochemistry of eodiagenetic interstitial fluids favored the dissolution of aragonite and precipitation of low magnesian calcite, moldic porosity was generated and primary interparticle porosity was reduced, resulting in low permeability. The favoring of neomorphism over dissolution resulted in higher permeability owing to the preservation of primary porosity. Mg-clay ooidal arenites were strongly affected by meteoric phreatic eodiagenetic conditions, which caused precipitation of dolomite and silica, and dissolution of ooids, locally resulting in rocks composed almost exclusively by silica interlayered or associated with dolomite. As a result, a complex porous system was generated, composed by moldic, microcrystalline and intergranular porosity, with highly heterogeneous patterns of permeability.

The main reservoirs from the Macabu Formation consist of abiotic low magnesium calcite crusts with growth-framework primary porosity, or interstitial porosity generated by the dissolution of syngenetic Mg-clays, which patterns of preservation, dissolution and replacement exerted an important control on the quality of reservoirs. Petrographic evidence indicates that most of the mineralogical and porosity evolution of these rocks occurred during eodiagenesis, under direct influence of compositional fluctuations of the lacustrine waters, mainly in response to the high reactivity of the Mg-clay minerals. Subordinately, rocks composed by intraclasts of crusts and spherulites of calcite containing minor amounts of magnesian clay ooids/peloids or clay matrix, may constitute the reservoir where interparticle porosity was preserved and/or clays were dissolved.

Areas strongly fractured, dissolved, dolomitized and silicified were affected by the circulation of hydrothermal fluids, which locally generated high porosity and permeability. The distribution of such intervals with high permeabilities represents special interest to the exploration and development of the reservoirs, because of their high impact on the migration of fluids, as well as on the communication of pressure. Future work should focus on understanding the influence and controls of hydrothermal processes modifying the rocks from the Lagoa Feia Group.

This study represents the first published petrographic characterization of the highly unconventional sag carbonate reservoirs, and stresses the crucial importance of systematic petrographic studies for the understanding and prediction of the quality of complex reservoirs.

7 ACKNOWLEDGMENTS

The authors would like to thank to Brazilian Petroleum Company PETROBRAS S.A. for the access to samples, well logs, petrophysical analysis, and for the license to publish this work.

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