UNIVERSIDADE FEDERAL DO RIO GRANDE DO SUL

INSTITUTO DE GEOCIÊNCIAS

PROGRAMA DE PÓS-GRADUAÇÃO EM GEOCIÊNCIAS

# DEPOSIÇÃO, DIAGÊNESE E POTENCIAL DE RESERVATÓRIO DAS ROCHAS SEDIMENTARES NÃO-CARBONÁTICAS DA SEÇÃO RIFTE DA BACIA DE CAMPOS

**Garibaldi Armelenti**

**Orientador: Prof. Dr. Luiz Fernando De Ros**

**Porto Alegre, 2014**

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**Dissertação de Mestrado apresentada como requisito para obtenção do Título de Mestre em Geociências.**

**Porto Alegre, 2014**

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## RESUMO

Um estudo petrográfico foi conduzido na seção rifte do Grupo Lagoa Feia, Cretáceo Inferior da Bacia de Campos, margem leste brasileira, como parte de um projeto regional integrado. Os principais constituintes das rochas analisadas são grãos siliciclásticos e vulcanoclásticos, oóides e pelóides estevensíticos, e bioclastos de bivalves e ostracodes. Este estudo foi focado nas rochas clásticas, estevensíticas e híbridas, desde que os estudos precedentes ficaram limitados às rochas carbonáticas que constituem os reservatórios produtores. A maior parte da sedimentação rifte foi intrabacial, com concentração da contribuição extrabacial na proximidade das falhas de borda dos blocos rifteados. Nos arenitos e conglomerados clásticos, a mistura de fragmentos vulcânicos arredondados com grãos de quartzo e feldspatos e fragmentos plutônicos angulosos indica a reciclagem de depósitos epiclásticos do início do rifte, combinada com sedimentos de primeiro ciclo erodidos de blocos soerguidos do embasamento granítico-gnáissico. Os oóides e pelóides estevensíticos foram formados em ambientes lacustres alcalinos rasos, levemente agitados por ondas ou correntes. Eles foram misturados com bioclastos de bivalves e ostracodes e com sedimentos clásticos em toda seção rifte. Esta re-deposição gravitacional foi promovida por movimentos tectônicos ao longo das margens dos blocos estruturais falhados. Os principais processos diagenéticos nos arenitos e conglomerados clásticos e nos arenitos híbridos são a cimentação e substituição de grãos por esmectita, zeolitas, calcita e dolomita, compactação limitada e dissolução de feldspatos, fragmentos vulcânicos e bioclastos. Os arenitos estevensíticos sofreram cimentação precoce e substituição dos oóides e pelóides por quartzo, calcita e dolomita, ou intensa compactação dos grãos estevensíticos dúcteis nas áreas não cimentadas. Arenitos e conglomerados vulcanoclásticos com porosidade intergranular parcialmente reduzida por franjas de esmectita e com alguma dissolução de grãos podem constituir reservatórios de qualidade regular. Arenitos estevensíticos e híbridos com dissolução de grãos estevensíticos, bioclastos e de cimento de calcita podem também constituir reservatórios, com qualidade potencial limitada pela conexão restrita de tais sistemas porosos. A compreensão dos controles espaciais e temporais sobre a evolução deposicional e diagenética das litologias rifte carbonáticas e não-carbonáticas, gravitacionalmente re-depositadas, irá contribuir para o estabelecimento de novas estratégias de exploração para a Bacia de Campos.

## ABSTRACT

A petrographic study was conducted on the rift section of the Lagoa Feia Group, Lower Cretaceous of the Campos Basin, eastern Brazilian margin, as part of a regional, integrated project. The main primary constituents of the analyzed rocks are siliciclastic and volcaniclastic grains, stevensitic ooids and peloids, and bivalve and ostracod bioclasts. This study focused in the clastic, stevensitic and hybrid rocks, as previous studies were limited to the bioclastic rudstones and grainstones that constitute the producing reservoirs. Most rift sedimentation was intrabasinal, with extrabasinal contribution concentrated at the proximity of border faults along rifted blocks. In the clastic sandstones and conglomerates, the mixture of rounded volcanic fragments with angular quartz, feldspars and plutonic fragments indicates the recycling of early rift epiclastic deposits, combined with first-cycle sediments eroded from uplifted granitic-gneissic basement blocks. Stevensitic ooids and peloids were formed in shallow, alkaline lacustrine environments, slightly agitated by waves or currents. They were mixed throughout the rift section with bivalve and ostracod bioclasts, and with clastic sediments. This gravitational re-deposition was promoted by intense and recurrent tectonic movements along the margins of rifted structural blocks. The main diagenetic processes in clastic sandstones and conglomerates and hybrid arenites are cementation and grain replacement by smectite, zeolites, calcite and dolomite, limited compaction and dissolution of feldspars, volcanic fragments and bioclasts. Stevensitic arenites experienced early cementation and replacement of ooids and peloids by quartz, calcite and dolomite, or intense compaction of the ductile stevensitic grains in uncemented areas. Volcaniclastic sandstones and conglomerates with intergranular porosity partially reduced by smectite rims and some grain dissolution may constitute fair hydrocarbon reservoirs. Stevensitic and hybrid arenites with dissolution of stevensitic grains, bioclasts and calcite cement may also constitute reservoirs, with potential quality limited by the poor connection of such pore systems. The understanding of the space and time controls on the depositional and diagenetic evolution of the dominantly intrabasinal, gravitationally re-deposited rift carbonate and non-carbonate rocks will contribute to new exploration strategies for the Campos Basin.

# SUMÁRIO

## LISTA DE FIGURAS



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# 1. **TEXTO EXPLICATIVO**

Sobre a Estrutura desta Dissertação:

Esta dissertação de Mestrado está estruturada em torno do artigo: *Deposition, Diagenesis and Reservoir Potential of Non-Carbonate Sedimentary Rocks from the Rift Section of Campos Basin, Brazil*, submetido para publicação no periódico *Marine and Petroleum Geology*. Consequentemente, sua organização compreende as seguintes partes principais:

a) Introdução sobre o tema e descrição do objeto da pesquisa de mestrado, onde estão sumarizados os objetivos e a filosofia de pesquisa desenvolvidos, o estado da arte sobre o tema de pesquisa, e os principais resultados e interpretações do estudo.

b) Artigo: *Deposition, Diagenesis and Reservoir Potential of Non-Carbonate Sedimentary Rocks from the Rift Section of Campos Basin, Brazil*, submetido para publicação no periódico internacional *Marine and Petroleum Geology*, escrito pelo autor durante o desenvolvimento de seu Mestrado.

c) Anexos, compreendendo: tabela de resultados de petrografia quantitativa, descrições petrográficas individuais das amostras analisadas, documentação fotomicrográfica e resultados das análises de microscopia de elétrons retroespalhados (BSE) e de espectrometria de energia dispersada (EDS).

# 2. INTRODUÇÃO

A Bacia de Campos é a bacia brasileira mais prolífica, de acordo com a Agência Nacional de Petróleo (ANP). Os enormes volumes de óleo e gás descobertos correspondem a 84% de toda a produção de petróleo no Brasil. O petróleo descoberto na Bacia de Campos foi gerado na seção rifte do Grupo Lagoa Feia (Cretáceo Inferior; Guardado *et al.*, 2000). Nesta unidade, os principais reservatórios produtores correspondem a calcários lacustres bioclásticos, denominados "coquinas" (Bertani & Carozzi 1985a; 1985b; Dias *et al.*, 1988; Abrahão & Warme, 1990; Carvalho *et al.*, 2000). Devido à importância da produção de petróleo das "coquinas", os trabalhos sobre as características primárias e diagenéticas dos depósitos Lagoa Feia publicados até o momento foram direcionados unicamente para as rochas carbonáticas (Carvalho *et al.*, 2000; Castro, 2006). O presente trabalho tem como objetivo analisar os aspectos genéticos das rochas sedimentares não-carbonáticas do Grupo Lagoa Feia, incluindo rochas clásticas extrabaciais, rochas intrabaciais não-carbonáticas e rochas híbridas extra-intrabaciais. Através de um estudo petrográfico específico, integrado com análises sedimentológicas, petrológicas, estratigráficas e sísmicas como parte de um projeto financiado pelo BG Group, espera-se contribuir para uma melhor compreensão das condições deposicionais destas rochas, assim como dos processos e padrões da sua diagênese e sua potencial qualidade como reservatórios de hidrocarbonetos.

## 2.1. Localização da área de estudo

A Bacia de Campos situa-se na margem leste brasileira, ao longo do litoral norte do Estado do Rio de Janeiro, sendo limitada ao norte pelo Arco da Vitória, que a separa da Bacia do Espírito Santo, e ao sul pelo Arco de Cabo Frio, que a separa da Bacia de Santos. Abrange uma área de aproximadamente 120.000 km$^2$, sendo uma pequena área *onshore* (5800 km$^2$) e o restante *offshore*, com cotas batimétricas rasas a profundas (figura 1).

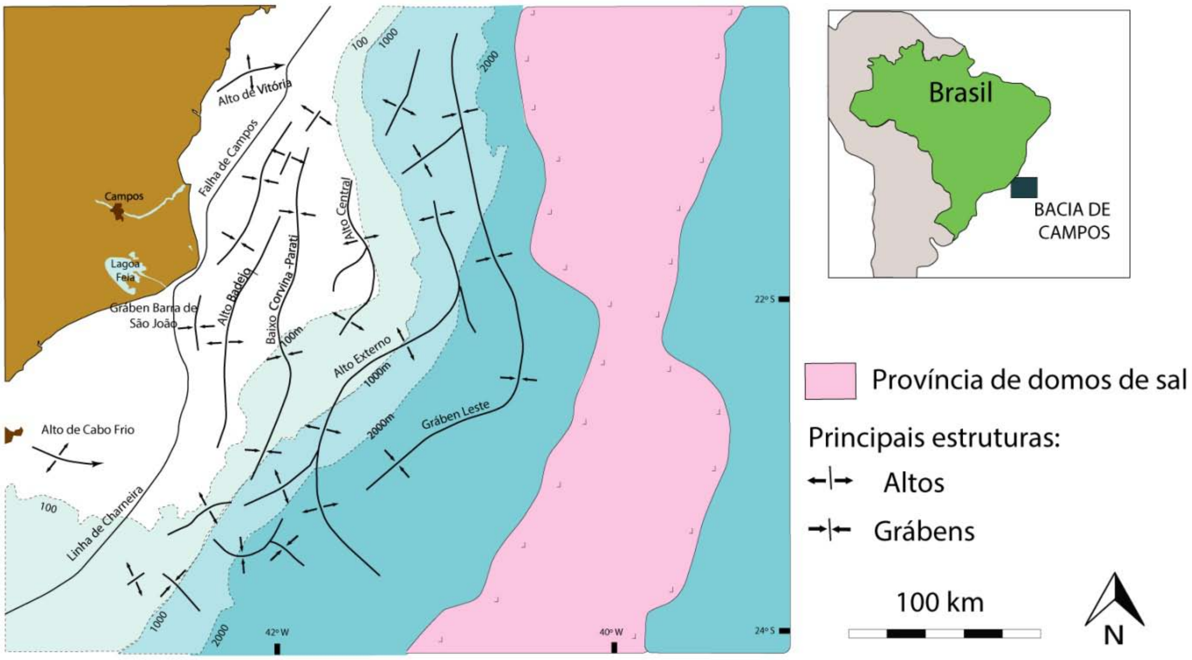



Figura 1. Mapa de localização da Bacia de Campos (modificado de Guardado *et al*., 2000).

# 3. GEOLOGIA REGIONAL DA BACIA DE CAMPOS

## 3.1. Contexto Geológico Regional

A primeira carta estratigráfica da Bacia de Campos foi proposta por Schaller (1973), e posteriormente modificada por Rangel *et al*. (1994). Na carta estratigráfica mais recente proposta por Winter *et al*. (2007) (figura 2), as formações Lagoa Feia e Macaé foram elevadas a Grupo, e os membros componentes das mesmas foram elevados a Formação. O embasamento cristalino da Bacia de Campos é constituído por gnaisses de idade pré-cambriana (Rangel *et al*., 1994; Winter *et al*., 2007). Sobrepostos a estes ocorre a Formação Cabiúnas, caracterizada por uma sucessão vulcano-sedimentar de idade neocomiana (Misuzaki *et al.*, 1988) com espessuras máximas perfuradas de 650 m (Winter *et al*., 2007). A Formação Cabiúnas é constituída por derrames de basaltos toleiíticos, tufos, brechas hidrovulcânicas, e rochas epiclásticas e de granulometria fina. Mizusaki *et al.* (1988) interpretaram a ocorrência de vulcanismo subaéreo nos campos de Badejo e Pampo, e subaquoso (lacustre) na região de Linguado. Idades K/Ar em rocha total (Mizusaki *et al*., 1992) em Badejo, Linguado, Pampo e no poço 1-RJS-0036-RJ variam de 134 a 111 Ma.

A sucessão rifte da Bacia de Campos corresponde à Formação Cabiúnas e à porção inferior do Grupo Lagoa Feia (formações Atafona, Coqueiros e Itapaboana; Winter *et al*., 2007). A Formação Atafona foi depositada no Andar Barremiano (andares locais Aratu superior e Buracica), a Formação Coqueiros tem idade do Barremiano Superior ao Aptiano Inferior (andar local Jiquiá), e a Formação Itapaboana do Barremiano ao Aptiano Inferior (Winter *et al.*, 2007).

O Grupo Lagoa Feia é caracterizado por sedimentos siliciclásticos, carbonáticos e evaporíticos depositados durante a fase rifte e pós-rifte. Segundo Guardado *et al*. (2000), as espessuras deste pacote variam de 200 m para mais de 1.500 m. O Grupo Lagoa Feia é composto pelas formações Atafona, Itapaboana, Coqueiros, Macabu, Gargaú e Retiro. A Formação Atafona é constituída por arenitos, siltitos e folhelhos depositados em ambiente lacustre alcalino onde precipitaram filossilicatos magnesianos identificados como kerolita, estevensita e talco (Bertani & Carozzi, 1985a; 1985b; Rehim *et al*., 1986; Winter *et al*., 2007). A Formação Coqueiros é representada por depósitos de coquinas compostos predominantemente por moluscos bivalves, intercalados com folhelhos e carbonatos lacustres (Winter *et al*., 2007). Segundo Baumgarten *et al.* (1988), os depósitos de coquinas são compostos por ciclos deposicionais de fácies de calcirrudito, calcarenito, calcilutito e aleatoriamente fácies de bioacumulados, podendo formar pacotes com espessuras maiores que 100 metros (Winter *et al*., 2007). A Formação Itapaboana é composta por conglomerados, arenitos, siltitos e folhelhos depositados nas porções proximais da bacia e ao longo da falha de borda (Winter *et al*., 2007). As formações Macabú e Gargaú são constituídas por sedimentos carbonáticos, margas e arenitos, depositados em ambiente transicional raso (Winter *et al*., 2007). A Formação Retiro é caracterizada por depósitos evaporíticos compostos por anidrita, halita e carnalita/silvita depositadas em ambiente marinho/lagunar sob clima árido (Winter *et al.*, 2007). As camadas de halita ocorrem remobilizadas, originando domos de sal que cortam as camadas sobrejacentes (Rangel *et al*., 1994). A porção superior do Grupo Lagoa Feia (formações Gargaú, Macabú, Retiro e Itapaboana superior) foi depositada já na fase pós-rifte (sag) da Bacia de Campos.

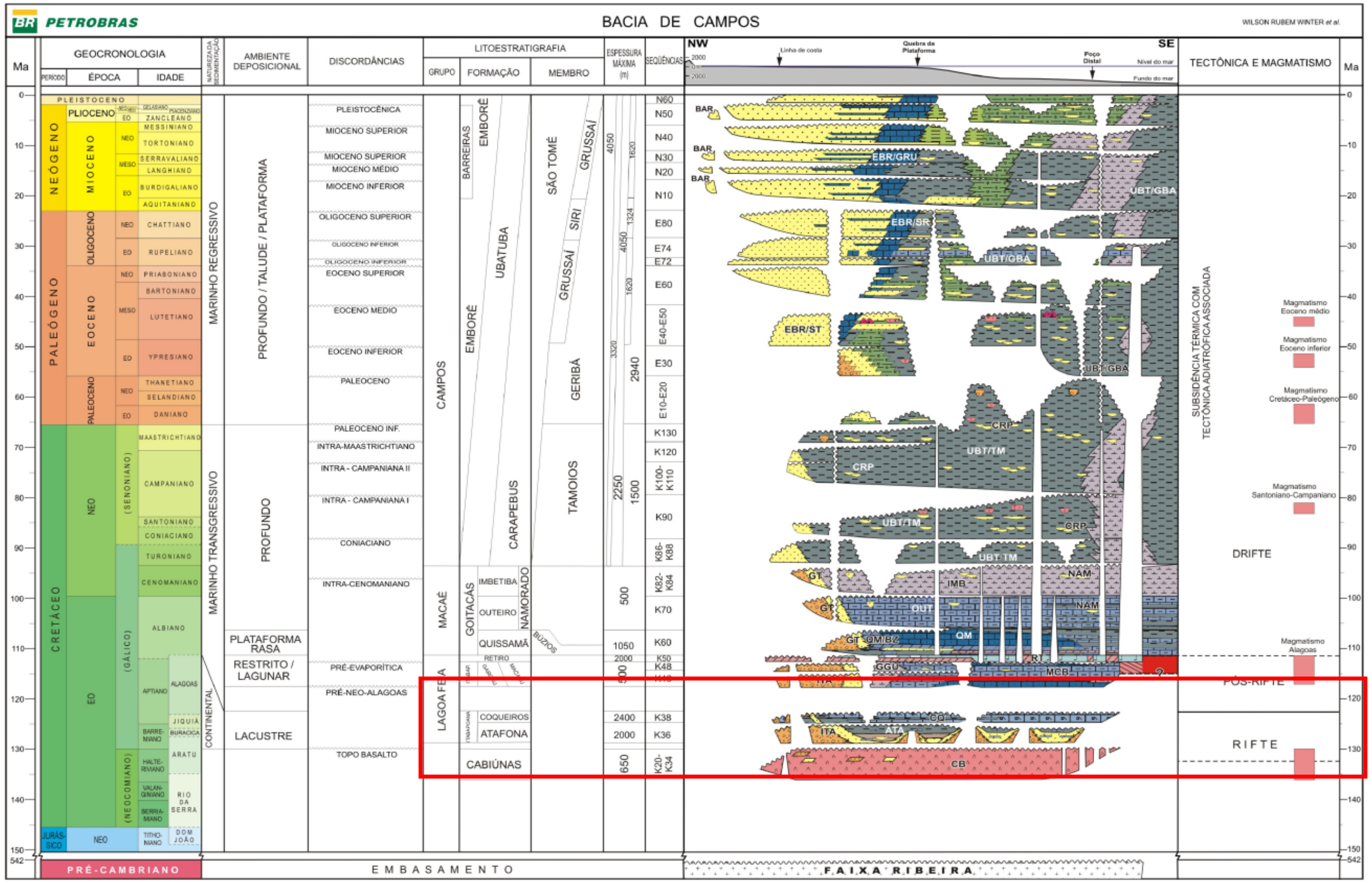



Figura 2. Coluna estratigráfica da Bacia de Campos (Winter *et al*., 2007), com o intervalo estudado marcado.

A sucessão pós-rifte e drifte da Bacia de Campos inclui os grupos Macaé e Campos. O Grupo Macaé é caracterizado por sedimentos carbonáticos depositados sobrepostos aos evaporitos da Formação Retiro (Winter *et al.*, 2007). Este grupo é constituído pelas formações Goitacás, Quissamã, Outeiro, Namorado e Imbetiba. A Formação Goitacás e a Formação Quissamã estão localizadas na porção proximal da Bacia de Campos, enquanto que as formações Outeiro e Namorado localizam-se na porção distal. A Formação Goitacás é representada por conglomerados e arenitos siliciclásticos. A Formação Quissamã, informalmente é chamada de "Macaé Inferior" ou também "Macaé Água Rasa" (Rangel *et al*., 1994), é representada por depósitos carbonáticos correspondentes a calcarenitos e calcirruditos oolíticos. Sobrepostos a estas camadas ocorre a Formação Outeiro, também conhecida por "Macaé Superior" ou "Seção Bota" devido a seu padrão em perfis elétricos. Esta unidade é constituída por calcilutitos, margas e folhelhos, por vezes com a intercalação de camadas arenosas isoladas da Formação Namorado (Rangel *et al*., 1994). A Formação Namorado é representada por depósitos arenosos originados por fluxos hiperpicnais, composto por arenitos feldspáticos turbidíticos (Winter *et al*., 2007). A Formação Imbetiba inclui depósitos de pelitos e margas de idade cenomaniana (Winter *et al*., 2007) com ocorrência isoladas dos arenitos turbidíticos da Formação Namorado. Os pelitos da Formação Imbetiba compõem uma grande cunha clástica que cobriu os depósitos carbonáticos de água rasa, cessando a formação dos carbonatos do Grupo Macaé (Winter *et al*., 2007).

O Grupo Campos consiste em sedimentos depositados do Turoniano ao Recente em ambiente marinho progressivamente mais profundo. Este grupo engloba as formações Ubatuba, Carapebus e Emborê (Winter *et al.*, 2007). A Formação Ubatuba é constituída por até milhares de metros de folhelhos depositados em ambiente marinho (Rangel *et al*., 1994). Intercalados com a Formação Ubatuba ocorrem arenitos finos a conglomerados siliciclásticos da Formação Carapebus, depositados por fluxos hiperpicnais em ambiente marinho profundo (Winter *et al*., 2007). A Formação Emborê é composta pelos membros São Tomé, constituída por sedimentos clásticos grossos; Membro Siri e Membro Grussai, constituídas por calcarenitos bioclásticos.

### 3.2. Arcabouço estrutural da Bacia de Campos

A Bacia de Campos apresenta dois estilos tectônicos distintos: tectônica diastrófica, que afeta os sedimentos da fase rifte, e tectônica adiastrófica, relacionada à halocinese, que atua sobre os sedimentos da fase transicional e drifte (Chang *et al*., 1990).

Nas estruturas da fase rifte observa-se um paralelismo entre os falhamentos da bacia e os principais alinhamentos do embasamento adjacente, com direção NE (Dias *et al.*, 1990). O padrão tectônico exibido na seção rifte é o de *horsts*, grabens e *half-grabens*, alongados na direção NE, limitados por falhas sintéticas e antitéticas. Falhamentos subordinados ocorrem nas direções NNW-SSE e E-W. A Charneira de Campos é uma feição importante na bacia, pois separa as áreas de embasamento raso e embasamento profundo, em cujo bloco baixo se depositou uma espessa seção rifte (figura 3).

Dentre os altos presentes na Bacia de Campos destaca-se o Alto Regional de Badejo, com mergulho para N (Chang *et al*., 1990). Nos altos contemporâneos à deposição do Grupo Lagoa Feia foram depositados os melhores reservatórios de coquinas (Baumgarten *et al.,* 1988). Por outro lado, nos baixos sindeposicionais de Corvina-Parati, São Tomé, Marlim e Norte de Albacora acumularam-se espessas seções de pelitos, que constituem as rochas geradoras da Bacia de Campos. Uma importante discordância de idade aptiana, conhecida como discordância pré-Alagoas, separa o padrão tectônico da fase rifte do padrão da fase pós-rifte.

O padrão tectônico da fase pós-rifte é caracterizado principalmente por falhas relacionadas ao fluxo de sal, de geometria lístrica, com anticlinais e calhas associadas, domos e diápiros de sal e estruturas geneticamente relacionadas. Algumas reativações de falhas do embasamento também afetaram os sedimentos da fase transicional e drifte (figura 3).

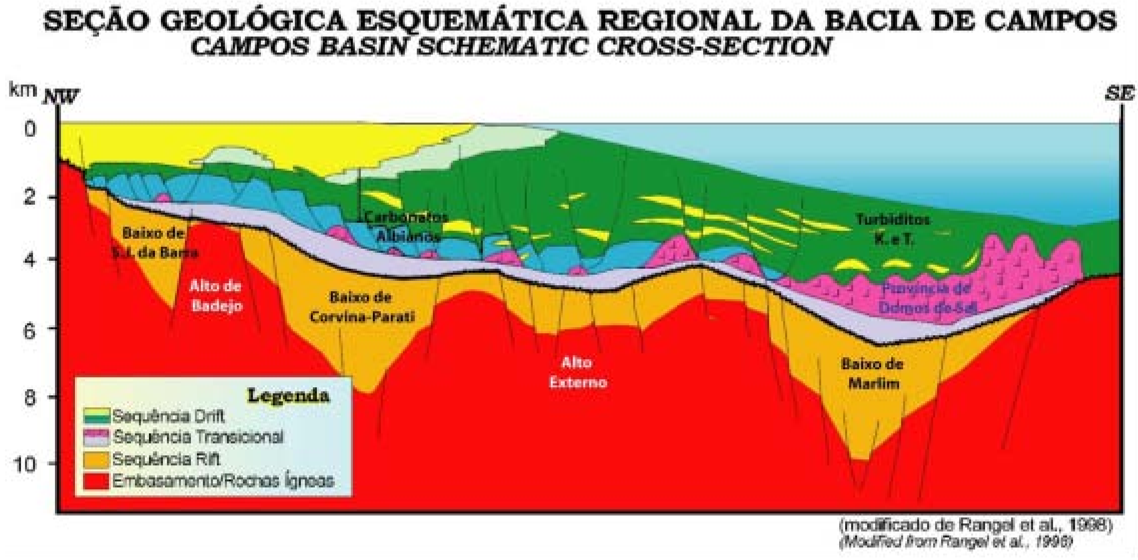



Figura 3. Seção geológica esquemática da Bacia de Campos. Extraído de Rangel *et al.* (1998).

### 3.3. Evolução tectono-sedimentar

Assim como as demais bacias da margem leste brasileira, a origem da Bacia de Campos está associada à quebra do Continente Gondwana e à abertura do Oceano Atlântico Sul. Winter *et al*. (2007) separou a evolução tectônica e estratigráfica da Bacia de Campos em três supersequências: Supersequência Rifte, Supersequência Pós-Rifte e Supersequência Drifte.

A Supersequência Rifte começou a ser depositada no Neocomiano. No início do rifteamento houve intensa atividade vulcânica, com extrusão de lavas basálticas e formação de rochas vulcanoclásticas, que constituem a Formação Cabiúnas (Ponte & Asmus, 1978). A continuação do processo de esforços distensivos produziu um sistema de grabens e horsts alongados na direção SW-NE, coincidentes com as direções de lineamentos do embasamento. Ao longo desses *rift valleys,* formados do Barremiano ao Aptiano, desenvolveu-se uma sedimentação lacustre (formações Atafona e Coqueiros), lateralmente associada a sedimentos continentais (leques aluviais) da porção inferior do Grupo Lagoa Feia (Formação Itapaboana).

A Supersequência Pós-Rifte ocorreu até o final do Aptiano, com deposição de sedimentos siliciclásticos, carbonáticos e evaporíticos. A sedimentação siliciclástica ocorreu nas porções proximais da bacia, com deposição típica segundo um padrão progradacional, enquanto que a sedimentação carbonática ocorreu dominantemente na porção superior deste intervalo, em padrão retrogradacional. O término desta supersequência ocorreu em ambiente marinho restrito associado a condições climáticas áridas a semi-áridas, correspondendo aos sedimentos evaporíticos da Formação Retiro (Winter *et al*., 2007).

A partir do Mesocretáceo (Turoniano–Coniaciano), se instaura na bacia um ambiente marinho franco, transgressivo. A transgressão resultou na deposição de uma espessa cunha de sedimentos siliciclásticos e carbonatos de plataforma rasa, que deram lugar a calcilutitos, margas e folhelhos de ambiente marinho cada vez mais profundo (Winter *et al*., 2007). A Supersequência Drifte, ou sequência marinha, iniciou-se após o término do evento de rifteamento e a abertura efetiva do Oceano Atlântico Sul, com a formação de uma rampa carbonática-clástica, sob clima quente e seco. Esta sedimentação desenvolveu-se durante o Albiano, acompanhando uma elevação eustática do nível do mar, e é representada pelos carbonatos oolíticos de água rasa da Fm. Quissamã do Gr. Macaé (Eo- Albiano), que gradam no topo da seqüência para calcilutitos, margas e folhelhos do Fm. Outeiro (Neo-Albiano/Turoniano; Winter *et al*., 2007). Concomitantemente à deposição destas duas unidades, em áreas mais proximais, foram depositados conglomerados polimíticos, arenitos e calcilutitos e margas da Fm. Goitacás. Ocasionalmente, em épocas de rebaixamento do nível do mar, foram gerados depósitos turbidíticos da Fm. Namorado. Nesta época começou a movimentação do sal depositado na fase anterior (halocinese), causada pelo basculamento progressivo da bacia para leste devido à subsidência térmica e à sobrecarga de sedimentos (Dias *et al.*, 1990). A partir do Neopaleoceno se instala na bacia uma configuração marinha regressiva, permanecendo até os dias atuais. Esta configuração se caracteriza por um conjunto de sedimentos clásticos progradantes, e compreende sistemas deposicionais que incluem desde leques costeiros e plataformas carbonáticas (Formação Emborê) até sistemas de talude e bacia profunda (Formação Ubatuba). Esta mudança no estilo sedimentar em relação à megasequência subjacente foi influenciada pela queda eustática de primeira ordem do nível do mar, aliada à baixa taxa de subsidência térmica e ao aumento do aporte sedimentar relacionado ao soerguimento da Serra do Mar durante o Terciário. A tectônica halocinética persistiu, gerando áreas rebaixadas que captaram sedimentos turbidíticos. Grandes sistemas turbidíticos estão presentes nesta megasequência e constituem importantes reservatórios de petróleo (Figueiredo & Mohriak, 1984; Guardado *et al*., 1990; Bruhn, 1998).

# 4. DIAGÊNESE: CONCEITOS BÁSICOS

Segundo uma definição geoquimicamente coerente, a diagênese compreende um campo de condições físicas e químicas que controla os processos geológicos atuantes sobre todos os tipos de materiais na superfície da crosta terrestre e nos primeiros milhares de metros de profundidade (engloba o intemperismo). Estes processos são controlados pela pressão, temperatura, composição dos fluidos intersticiais e pela composição química e mineralógica dos materiais. Os processos diagenéticos influenciam diretamente sobre a qualidade dos reservatórios, pois podem atuar tanto de maneira positiva, preservando e gerando porosidade, ou negativa, reduzindo ou destruindo totalmente a porosidade.

## 4.1. Estágios da diagênese

Morad *et al*. (2000) redefiniram os estágios da diagênese clástica, modificando as definições originais de Schmidt & McDonald (1979). A distribuição espacial dos estágios diagenéticos está esquematizada na figura 4.

Eodiagênese: atuante desde a superfície até profundidades em torno de 2 km, até 70°C de temperatura a baixa pressões; períodos de tempo muito variável. É influenciada pelo ambiente deposicional e/ou pela circulação de água superficial (marinha / meteórica).

Mesodiagênese rasa: atuante em profundidades que variam de 2 a 3 km, com temperaturas entre 70 e 100°C; pressão e temperaturas crescentes; fluidos diagenéticos modificados pelas reações com os minerais, circulando principalmente por compactação.

Mesodiagênese profunda: profundidades superiores a 3 km, temperaturas maiores que 100°C. Evolução: soerguimento/ telodiagênese/ soterramento crescente/ metamorfismo (gradação para anquimetamorfismo).

Telodiagênese: re-exposição de rochas previamente soterradas às condições superficiais por soerguimento e erosão de parte da seção (formação de discordâncias) ou infiltração de água meteórica a grandes profundidades.

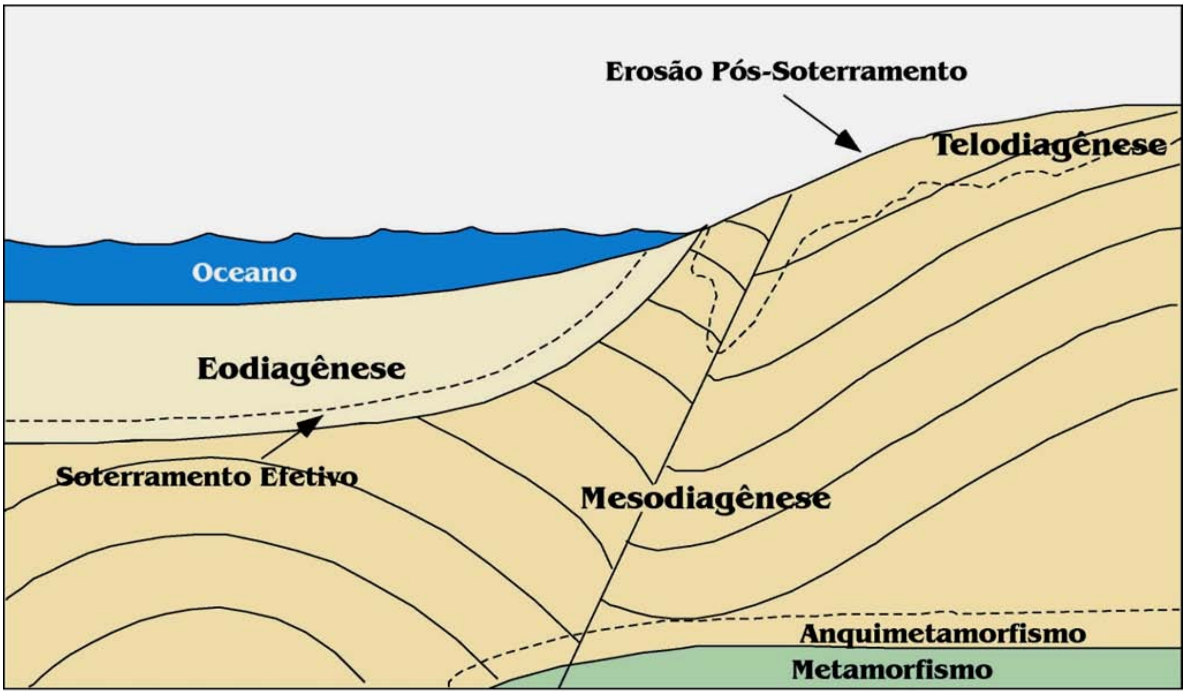



Figura 4. Representação da distribuição espacial dos estágios da diagênese (De Ros, 1996).

## 4.2. Principais processos diagenéticos

Os principais processos atuantes na diagênese podem ser sumarizados como segue:

Compactação: ocasionado pelo soterramento, com redução do espaço poroso. Pode ser física, através do rearranjo, fraturamento ou esmagamento dos grãos, ou química, através da dissolução por pressão nos contatos intergranulares ou ao longo de estilolitos.

Dissolução: pode afetar constituintes primários ou diagenéticos. Pode ser congruente, total, com total remoção dos materiais como íons em solução (ex: carbonatos); ou incongruente, incompleta (ex.: feldspatos → caulinita).

Autigênese: precipitação de novos minerais, cimentando os poros ou substituindo constituintes pré-existentes via dissolução e precipitação simultâneas.

Hidratação / desidratação: entrada ou saída de água da estrutura cristalina; ex.: anidrita ←→ gipsita.

Oxidação: remoção de elétrons dos elementos dos materiais na superfície ou próximo à superfície, sob influência de $O_2$ ou bactérias aeróbicas; ex.: $Fe^{2+} \rightarrow Fe^{3+}$, formando hematita.

Redução: adição de elétrons dos elementos dos materiais, sob influência da matéria orgânica e de bactérias anaeróbicas; ex.: $Fe^{3+} \rightarrow Fe^{2+}$, formando pirita, siderita.

Recristalização: Crescimento ou diminuição do tamanho cristalino, mantendo-se a mesma composição mineralógica.

Estabilização / neomorfismo / inversão: substituição de uma fase mineralógica de composição similar; ex.: aragonita → calcita.

## 4.3. Controles da diagênese clástica

Morad *et al*. (2012) definiram os principais controles atuantes sobre a diagênese como sendo a composição detrítica, a composição dos fluidos intersticiais, o fluxo dos fluidos e fatores físicos como pressão, temperatura e tempo (figura 5) A composição detrítica é definida em função essencialmente da proveniência, controlada pelos terrenos e rochas-fonte, pela geografia e pelo clima. A composição dos fluidos é controlada inicialmente pelo ambiente de deposição, que controla a composição das águas deposicionais, além das texturas, estruturas e geometria dos sedimentos, e, portanto o fluxo de fluidos, além das próprias litologias associadas. A composição dos constituintes diagenéticos anteriormente formados influencia as reações diagenéticas durante o soterramento ou soerguimento posteriores. A temperatura, pressão e tempo são parâmetros controlados pela história do soterramento, em função essencialmente do ambiente tectônico da sucessão sedimentar.


Figura 5. Representação das relações entre os parâmetros controladores da diagênese (modificado de Morad *et al*., 2012).

# 5. DIAGÊNESE E POTENCIAL DE RESERVATÓRIO DE ARENITOS E CONGLOMERADOS VULCANOCLÁSTICOS

Arenitos e conglomerados vulcanoclásticos por muitos anos não foram alvo de exploração de hidrocarbonetos, em razão de sua conhecida má qualidade como reservatórios (Surdam & Boles, 1979; Galloway, 1979). Estas rochas geralmente apresentam baixos valores de porosidade e permeabilidade, devido à destruição da porosidade logo após a deposição pela compactação e/ou pela precipitação de constituintes autigênicos. Isto se deve à composição imatura destas rochas, ricas em componentes altamente instáveis e reativos como feldspatos, anfibólios, piroxênios, vidro vulcânico e fragmentos de rochas vulcânicas. Os processos diagenéticos em arenitos e conglomerados vulcanoclásticos são controlados por diversos fatores, como o ambiente deposicional, a composição detrítica, a granulometria e a química dos fluidos intersticiais, além da pressão, temperatura e profundidade do soterramento (Surdam & Boles, 1979; Iijima, 1980). A descoberta de reservatórios de hidrocarbonetos nestas rochas em diferentes partes do mundo (Seemann & Scherer, 1984; Noh & Boles, 1993; Reed *et al*., 1993) tem atualmente ocasionado maior atenção da indústria exploratória para estas litologias.

A diagênese dos depósitos vulcanoclásticos é normalmente bem mais complexa do que a de outros depósitos siliciclásticos. Isto se deve aos componentes altamente reativos destas litologias, que sofrem alteração química, via dissolução e substituição, e física com efeitos da compactação mecânica, logo após a deposição, resultando em esmagamento e fraturamento de grãos mais frágeis (Mathisen & McPherson, 1991). Em razão da grande reatividade dos constituintes, a qualidade de reservatório dos arenitos e conglomerados vulcanoclásticos é controlada dominantemente por processos eodiagenéticos.

A profundidade de soterramento influencia diretamente sobre a qualidade reservatório das rochas vulcanoclásticas. Seemann & Scherer (1984) observaram que arenitos vulcanoclásticos de soterramento raso possuem melhor qualidade de reservatório dos que os de soterramento profundo a moderado, o que indica que, sob condições de soterramento crescente, a qualidade de reservatório dos arenitos vulcanoclásticos deteriora-se rapidamente, devido à ação da compactação mecânica e da cimentação crescentes. Remy (1994) mostrou que a qualidade de reservatório dos arenitos vulcanoclásticos da Bacia de Middle Park, Colorado (USA) foi prejudicada pela falta de um processo ativo de lixiviação por água meteórica na eodiagênese, por um prolongado episódio de compactação antes da precipitação de minerais autigênicos, e pela cimentação por argilominerais e zeolitas, que obstruiram os poros restantes, com fraca dissolução na fase de soterramento profundo.

Em alguns casos, entretanto, a cimentação eodiagenética impede a compactação mecânica, preservando parte da porosidade primária dos depósitos vulcanoclásticos, à qual pode vir a somar-se porosidade secundária gerada pela dissolução de cimentos e grãos. Nos arenitos vulcanoclásticos da Bacia de Junggar (China), Tang *et al.* (1994) verificaram que cutículas e franjas de esmectita formadas na eodiagênese sustentaram o arcabouço destes arenitos, preservando grande parte da porosidade primária original, que somada à secundária resultou em bons reservatórios.

Mathisen & McPherson (1991) observaram que a qualidade de reservatório de arenitos vulcanoclásticos estaria também relacionada a aspectos texturais como granulometria e seleção, por sua vez controlados, tal como em outros depósitos siliciclásticos, pelo ambiente deposicional, que também controla a composição e o fluxo de fluidos. A cimentação carbonática é mais comum nos depósitos vulcanoclásticos marinhos do que nos não-marinhos, onde o fluxo de água meteórica promove a remoção de íons dissolvidos (Stanley & Benson, 1979; Mathisen, 1984). Mathisen (1984) observou que os arenitos vulcanoclásticos não marinhos da Bacia de Cagayan, localizada no norte das Filipinas, possuem significativo volume de porosidade secundária originada por dissolução eodiagenética.

Muitos dos processos diagenéticos em sedimentos vulcanoclásticos envolvem reações de hidratação, como a dissolução de fragmentos vulcânicos e plagioclásios. Estas reações são importantes porque aumentam o pH das soluções intersticiais e liberam cátions em solução (Surdam & Boles, 1979). O aumento do pH e da salinidade tem um efeito direto sobre as reações diagenéticas subsequentes (Mathisen, 1984). A dissolução de silicatos pode também ocorrer como resultado da interação de ácidos orgânicos, oriundos de matéria orgânica presente em camadas adjacentes (Crossey *et al*., 1984). Os principais constituintes diagenéticos formados nos arenitos e conglomerados vulcanoclásticos são esmectitas, zeolitas, carbonatos, hematita, quartzo, K-feldspato, cloritas, ilita, caulinita, pirita e albita.

A autigênese de esmectita é o produto mais comum da alteração diagenética de materiais vulcânicos. A alteração de vidro vulcânico para esmectita é favorecida em ambientes de pH alcalino e elevada atividade de sílica (Jeans *et al*., 2000). As esmectitas substituem fragmentos vulcânicos, feldspatos e minerais pesados, e formam cutículas e franjas (Mathisen, 1984). A cobertura e substituição das cutículas e franjas de esmectita por zeolitas e calcita, comumente observada em arenitos vulcanoclásticos, é favorecida pela alteração das esmectitas em respostas a variações na composição e temperatura dos fluidos (McKinley *et al.*, 2003). A composição das esmectitas geradas na eodiagênese de sedimentos vulcanoclásticos é controlada pela composição dos materiais vulcânicos. A alteração de fragmentos de composição básica ou intermediária (basaltos, andesitos) tende a gerar esmectitas trioctaédricas, ricas em $Mg^{2+}$ e $Fe^{2+}$. Materiais de composição ácida ou alcalina félsica tendem a gerar esmectitas dioctaédricas aluminosas. Esmectitas dioctaédricas e trioctaédricas tem evolução diferenciada durante o soterramento com aumento da temperatura (Chang *et al*., 1986). Esmectitas dioctaédricas, ricas em $Al^{3+}$, evoluem para ilitas, através de interestratificados ilita-esmectita (Chang *et al*., 1996). A transformação de esmectitas em ilitas consome $K^+$ e libera $Si^{4+}$ para os fluidos intersticiais (Boles & Franks, 1979). O aumento da atividade de Si pode resultar na precipitação de quartzo (McKinley *et al.*, 2003). Esmectitas trioctaédricas, ricas em $Fe^{2+}$ e $Mg^{2+}$, evoluem para cloritas, através de interestratificados clorita-esmectita (Chang *et al*., 1986; Humphreys *et al*., 1994).

Zeolitas são comumente formadas em arenitos vulcanoclásticos pela interação de soluções alcalinas com fragmentos os vulcânicos instáveis, vidro vulcânico e plagioclásio (Hay, 1966; Iijima, 1980; Surdam & Boles, 1979; Noh & Boles, 1993; Tang *et al*., 1994; Hay & Sheppard, 2001). A dissolução de vidro vulcânico e plagioclásio e consequente formação de zeolitas é acelerada em ambientes com pH acima de 9 devido ao aumento da solubilidade de Si e Al (Taylor & Surdam, 1981). Zeolitas diagenéticas como a heulandita, clinoptilolita, analcima e laumontita são comuns em arenitos e conglomerados vulcanoclásticos originadas de grande variedade de materiais como vidro vulcânico, fragmentos de rocha vulcânica, feldspatos e argilominerais (Hay & Sheppard, 2001). A analcima é formada por precipitação direta ou através da reação entre salmouras intersticiais com argilominerais e plagioclásio (Chan, 1985). A presença de materiais vulcanoclásticos não é necessária para precipitar analcima, mas o ambiente deve ser alcalino e rico em Na, como em alguns lagos salinos, e preferencialmente com a presença de argilominerais ou plagioclásios (Hay, 1966). Alguns tipos de zeolitas formadas em baixa temperatura, como a clinoptilolita e a chabazita, são substituídas por analcima devido aumento de temperatura causado pelo soterramento (Aoyagi & Kazama, 1980). A formação de zeolitas pode estar também relacionada ao contato com intrusões (McKinley *et al*., 2001; Bernet & Gaupp *et al*., 2005). A laumontita forma-se em arenitos vulcanoclásticos normalmente acima de 50-60°C (Surdam & Boles, 1979).

A dissolução de feldspatos e vidro vulcânico aumenta a atividade de sílica nos fluidos intersticiais, o que pode ocasionar a precipitação de cimentos silicosos ou fornecer sílica para outras reações diagenéticas (Surdam & Boles, 1979). Reações de transformação de argilominerais também são fontes de Si, assim como dissolução por pressão ocasionada pela compactação. A ilitização de esmectita libera Si (Boles & Franks, 1979). Fluidos intersticiais enriquecidos em $Na^{+}$, $K^{+}$, $Si^{4+}$, $Al^{3+}$ promovem a precipitação de K-feldspato e albita. A albitização de feldspatos ocorre quando há alto teor de Na nos fluidos intersticiais. A substituição de esmectita ou zeolita por calcita ocasiona aumento de Na no meio circundante. A albitização de feldspatos é também favorecida pelo aumento de temperatura e pressão no soterramento (Tang *et al*., 1994).

A cimentação por calcita é comum em muitos arenitos vulcanoclásticos, principalmente em decorrência de sua composição detrítica rica em constituintes com Ca, como plagioclásios, piroxênios e fragmentos vulcânicos básicos e intermediários. A decomposição de matéria orgânica nos sedimentos associados favorece diretamente a precipitação de carbonatos. Bashari (1998) observou que a cimentação carbonática nos arenitos triássicos da Bacia de Bowen, Austrália, foi restrita aos intervalos ricos em fragmentos de rocha vulcânica, sugerindo que os íons de Ca derivados da alteração dos clastos vulcânicos propiciavam a cimentação. A precipitação precoce de calcita é mais comum em arenitos vulcanoclásticos marinhos do que os não-marinhos (Galloway, 1974). Ambientes salobros e alcalino ricos em Mg são propensos à precipitação de dolomita.

A precipitação de óxidos e hidróxidos de ferro é comum em arenitos vulcanoclásticos devido à presença de silicatos ferromagnesianos instáveis (Burns & Ethridge, 1979). Ambientes com clima árido ou semi-árido com fluxo de água meteórica reduzido são propícios à formação de hematita e outros óxidos e hidróxidos de ferro.

Os intensos processos diagenéticos nos sedimentos vulcanoclásticos podem afetar camadas de rochas adjacentes que não possuem grãos vulcanoclásticos em sua composição.

# 6. DEPOSIÇÃO E DIAGÊNESE DE ARENITOS ESTEVENSÍTICOS

Arenitos estevensíticos são formados por oóides e pelóides de estevensita, uma esmectita de composição fortemente magnesiana com fórmula geral: $(Ca_{0.5}Na)_{0.33}(Mg,Fe^{++})_3Si_4O_{10}(OH)_2 \bullet n(H_2O)$, ou: $Na_{0.2}(Mg_{2.3}Al_{0.3}Fe_{0.1})Si_4O_{10}(OH)_2 \bullet n(H_2O)$. A formação de estevensita ocorre caracteristicamente em lagos alcalinos e salinos, e em playas (Bradley & Fahey, 1962; Dyni, 1976; Tettenhorst & Moore, 1978; Eberl *et al*., 1982; Khoury *et al*., 1982; Jones & Weir, 1983; Jones, 1986; Darragi & Tardy, 1987; Martin de Vidales *et al*., 1991; Buch & Rose, 1996; Cerling, 1996; Chahi *et al*., 1997, 1999; Hover *et al*., 1999; Pozo & Casas, 1999; Mayayo *et al*., 2000; Yuretich & Ervin, 2002; Cuevas *et al*., 2003; Hover & Ashley, 2003; Furquin *et al*., 2008). Em diversas dessas ocorrências, a estevensita encontra-se interestatificada com talco ou com kerolita, sua versão hidratada. Bertani & Carozzi (1985a; 1985b) identificaram como kerolita a composição dos oóides e pelóides do Grupo Lagoa Feia, relacionando sua ocorrência à alteração de materiais vulcânicos. Rehim *et al.* (1986) caracterizaram mais precisamente sua composição como estevensita e camadas-mistas talco-estevensita, interpretando sua gênese como produto de precipitação direta em um ambiente lacustre alcalino rico em Mg e Si. No conjunto de trabalhos referidos, as condições para a formação desses filossilicatos magnesianos são genericamente consideradas como envolvendo alta atividade de Si e Mg e alto pH (acima de 9 para a estevensita).

Estudos experimentais desenvolvidos por Jones (1996) e por Tosca & Masterson (in press) mostraram que a salinidade, o pH e a razão Mg/Si do fluido, somados ao nível de energia e a outros fatores do ambiente deposicional, são controles determinantes sobre qual fase mineral ira precipitar, entre estevensita, talco, kerolita ou sepiolita. De acordo com Jones (1996), condições mais diluídas ou palustres favoreceriam a precipitação de sepiolita. O aumento do pH e da razão Mg/Si favoreceria a precipitação de kerolita. O aumento do pH e da salinidade favoreceriam a formação de estevensita. Tosca & Masterson (in press) reconheceram, de forma semelhante, a precipitação de sepiolita em condições de pH mais baixo e de menor razão Mg/Si, de estevensita em condições de alto pH (acima de 9) e salinidade e alta razão Mg/Si, com a formação de kerolita em condições intermediárias.

# 7. METODOLOGIA

Os métodos utilizados na execução deste trabalho foram: levantamento bibliográfico, análise petrográfica quantitativa, e análise de microscopia de varredura de elétrons retroespalhados (BSE) com suporte de espectrometria de energia dispersada (EDS).

## 7.1. Levantamento bibliográfico

Foi realizado uma compilação e avaliação de artigos referentes à sedimentação e evolução estratigráfica da Bacia de Campos, à diagênese em arenitos vulcanoclásticos e à deposição de arenitos estevensíticos.

## 7.2. Petrografia quantitativa

Foi realizada a descrição petrográfica quantitativa de 42 lâminas delgadas preparadas de amostras de testemunhos selecionados, impregnadas com resina epoxy azul, através do uso de microscópios de luz polarizada Zeiss, Leitz e Leica. Para a quantificação das lâminas delgadas, foram contados 300 pontos em cada, perpendicularmente à estrutura principal e/ou à orientação da fábrica das amostras, identificando os constituintes primários, diagenéticos e os tipos de porosidade, suas localizações e relações paragenéticas, com uso do software Petroledge® (De Ros *et al*., 2007). O método Gazzi-Dickinson (Zuffa, 1985) foi utilizado na contagem modal. Este método consiste em contar os cristais ou grãos de quartzo, feldspatos, etc. maiores que o tamanho silte (0,062 mm) no interior de fragmentos de rocha como seus constituintes mineralógicos. Somente fragmentos com textura fina são contados como tal, ex: vulcânicas afaníticas, fragmentos metamórficos de baixo grau, chert, rochas carbonáticas, lutitos, etc. (figura 6). O objetivo deste método é salientar a composição mineralógica / litológica das rochas-fonte, independente da granulometria dos sedimentos, desde que sedimentos mais grossos contém naturalmente mais fragmentos de rocha. A normatização do método Gazzi-Dickinson permite identificar as assinaturas composicionais dos principais ambientes tectônicos (crátons estáveis, riftes alimentados por blocos soerguidos do embasamento, arcos magmáticos e cinturões orogênicos com reciclagem de rochas sedimentares e meta-sedimentares; tabela 1) através do exame de diagramas ternários discriminantes (Dickinson, 1985; figura 7).

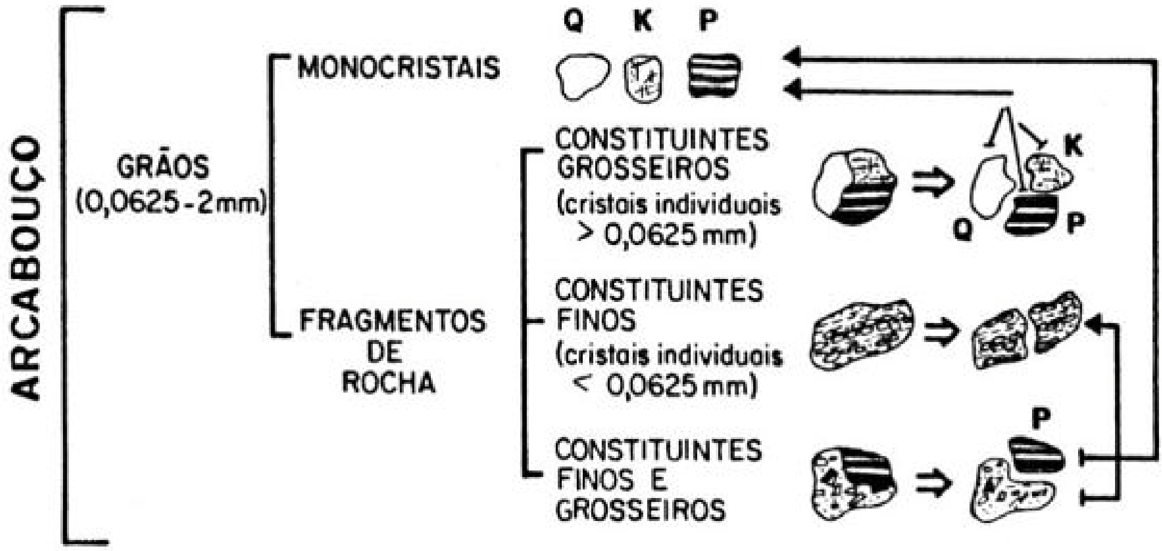
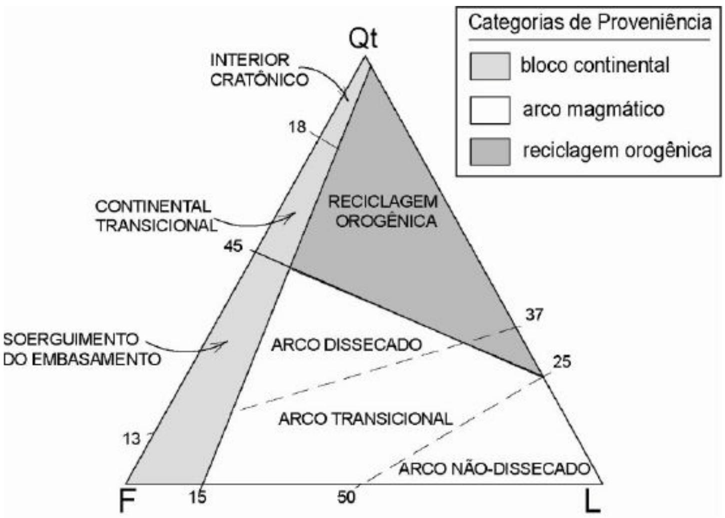



Figura 6. Representação do sistema de contagem Gazzi-Dickinson, utilizado para minimizar o efeito do tamanho de grão na composição do arcabouço de arenitos. Modificado de Zuffa (1985).

Tabela 1. Tipos de proveniência tectônica, ambientes tectônicos e composição das areias geradas (Dickinson, 1985).

| TIPOS DE PROVENIÊNCIA | AMBIENTE TECTÔNICO | COMPOSIÇÃO DAS AREIAS GERADAS |
|---|---|---|
| Cráton estável | Intracontinental ou plataforma passiva | Areias quartzosas (ricas em Qt) com altas razões de Qm/Qp e K/P |
| Soerguimento do embasamento | Rifte ou ruptura transformante | Areias quartzo- feldspáticas (Qm-F) pobres em Lt e Qp, similares à área fonte |
| Arco magmático | Arco de ilhas ou arco continental | Areias feldspato-líticas (F-L) vulcanoclásticas com altas razões P/K e Lv/Ls, gradando para areias quartzo-feldspáticas derivadas de batólitos |
| Reciclagem orogênica | Cinturão orogênico ou complexo de subducção | Areias quartzo-líticas (Qt-Lt) ricas em Ls (sedimentares e meta-sedimentares), pobres em F e Lv, com razões variáveis de Qm/Qp e Qp/L |




**Qt** - quartzo total, compreendendo grãos de quartzo macrocristalinos (cristais > 0,06 mm) mono ou policristalinos, isolados ou dentro de fragmentos de rocha plutônicas, sedimentares ou metamórficas;

**F** - feldspatos potássicos e plagioclásios isolados ou dentro de fragmentos de rocha;

**L** - fragmentos de rochas vulcânicas, hipoabissais, sedimentares e metamórficas.

Figura 7. Diagrama discriminante dos tipos de proveniência tectônica, de acordo com a composição das areias geradas (Dickinson, 1985).

Tingimento com uma solução ácida diluída de alizarina vermelha e ferrocianeto de potássio foi utilizado para determinar o tipo de carbonato presente em cada lâmina (cf. Dickson, 1965). O método de Kahn (1956) foi utilizado para avaliação do índice de empacotamento das amostras.

## 7.3. Microscopia eletrônica de varredura

Análises de microscopia eletrônica de varredura (MEV) por elétrons retroespalhados (backscattered electrons – BSE) foram executadas com um equipamento JEOL JSM-6610LV sobre 7 lâminas delgadas polidas e cobertas com carbono, para uma melhor definição das variações composicionais dos constituintes primários e diagenéticos, suas relações paragenéticas entre si e com a porosidade. As análises de BSE contaram com suporte de análises da composição elementar dos constituintes por espectrometria de energia dispersada (EDS), executadas em equipamento Brucker acoplado ao MEV.

# 8. RESUMO DOS PRINCIPAIS RESULTADOS E INTERPRETAÇÕES

1. Os sedimentos que compõem o Grupo Lagoa Feia da Bacia de Campos foram depositados diretamente sobre os basaltos da Formação Cabiúnas, extrudidos durante a fase inicial de separação dos continentes Sul-Americano e Africano.

2. A deposição inicial se deu nos baixos estruturais, onde diversos lagos pequenos, inicialmente isolados uns dos outros, se formaram. Os constituintes primários que compõem as rochas analisadas são: clásticos (grãos siliciclásticos e vulcanoclásticos, raramente lama siliciclástica) carbonáticos (bioclastos de ostracodes, pelecípodes e raramente gastrópodes, oóides, intraclastos e pelóides) e estevensíticos (oóides, pelóides e raramente intraclastos e laminações estevensíticas).

3. A sedimentação durante a fase de rifteamento foi predominantemente intrabacial, com a contribuição extrabacial restrita à proximidade das falhas de borda.

4. As rochas magmáticas presentes na seção analisada correspondem a basaltos, diabásios, brechas hidrotermais e tipos hidromagmáticos (peperitos).

5. A atividade tectônica foi intensa durante a sedimentação da sucessão analisada, assim como as atividades magmática e hidrotermal associadas.

6. A mistura de fragmentos de rochas vulcânicas arredondados com grãos de quartzo, feldspatos e fragmentos de rochas plutônicas angulosos indica a ocorrência de reciclagem de depósitos vulcanoclásticos epiclásticos formados na fase inicial do rifteamento, combinada com sedimentos de primeiro ciclo, erodidos do embasamento granito-gnáissico nas bordas dos blocos falhados.

7. A interação de fluidos hidrotermais com rochas vulcânicas e intrusivas básicas e com fragmentos destas litologias representou uma fonte de Mg e Ca para a precipitação de estevensita e de calcita.

8. Os oóides e pelóides estevensíticos foram formados em ambientes lacustres rasos de pH elevado, levemente agitados por ondas e correntes. A estevensita é abundante em toda seção rifte, não estando restrita a intervalos deposicionais determinados.

9. As rochas estudadas da seção rifte são caracterizadas por uma mistura generalizada e em diferentes proporções de grãos estevensíticos e bivalves, indicando sistemática re-deposição gravitacional de sedimentos lacustres rasos, formados em locais e tempos diversos. O ambiente fortemente alcalino de geração de estevensita seria totalmente intolerável pelos bivalves.

**10.** Esta mistura de grãos estevensíticos com bioclastos de bivalves estaria relacionada a processos de re-sedimentação ocasionados por atividade tectônica atuante na bacia durante a deposição dos sedimentos. Movimentos tectônicos de falhas ao longo das margens lacustres provocariam o colapso dos depósitos de bioclastos e de estevensita, misturando em proporções variadas estes sedimentos originalmente incompatíveis.

11. Os principais processos diagenéticos ocorridos nas litologias clásticas são a cimentação e a substituição de grãos por esmectitas, zeolitas, calcita e dolomita.

12. Os principais processos diagenéticos ocorridos nas litologias híbridas são a cimentação por esmectita, calcita e dolomita, além da substituição de grãos clásticos e estevensíticos, recristalização de bioclastos, e a dissolução de grãos e do cimento de calcita.

13. Os principais processos diagenéticos ocorridos nas rochas estevensíticas são a cimentação por quartzo, além da intensa substituição dos oóides e pelóides estevensíticos, por calcita e dolomita. Na ausência de cimentação precoce, a compactação mecânica pela deformação dos grãos estevensíticos dúcteis destruiu toda a porosidade nessas rochas.

14. Os arenitos e conglomerados vulcanoclásticos com porosidade intergranular remanescente, junto à porosidade intragranular secundária em grãos de

feldspatos e fragmentos vulcânicos, apresentam melhor potencial para constituírem reservatórios. Infelizmente, parte desta porosidade está comumente ocupada por betume/óleo pesado, produto de degradação provavelmente ocorrida durante o soerguimento pós-rifte, o que pode não ter ocorrido em áreas mais profundas e distantes das falhas marginais dos grabens.

15. Os arenitos híbridos e estevensíticos também apresentam potencial para constituírem reservatórios através do desenvolvimento de porosidade móldica e intrapartícula por dissolução de grãos estevensíticos e bioclastos, ou da contração de grãos estevensíticos, mas com qualidade limitada pelo caráter pouco conectado e efetivo desses sistemas porosos.

# 9. REFERÊNCIAS BIBLIOGRÁFICAS


Abrahão, D. & Warme, J.E. 1990. Lacustrine and associated deposits in a rifted continental margin – lower Cretaceous Lagoa Feia Formation, Campos Basin, offshore Brazil. In: Katz B.J. (Ed.). Lacustrine Basin Exploration: Case Studies and Modern Analogs, AAPG Memoir 50, p. 287-305.

Aoyagi, K. & Kazama T. 1980. Transformational changes of clay minerals, zeolites and silica minerals during diagenesis. Sedimentology, 27: 179-188.

Bashari, A. 1998. Diagenesis and reservoir development of sandstones in the Triassic Rewan Group, Bowen Basin, Australia. Journal of Petroleum Geology, 21: 445-465.

Baumgarten, C.S., Dultra, A.J.C., Scuta, M.S., Figueiredo, M.V.L. & Sequeira, M.F.P.B. 1988. Coquinas da formação Lagoa Feia, Bacia de Campos: evolução da geologia de desenvolvimento. Boletim de Geociências da Petrobras, 2(1): 27-36.

Bertani, R.T. & Carozzi A.V. 1985a. Lagoa Feia Formation (Lower Cretaceous) Campos Basin, offshore Brazil: Rift valley type lacustrine carbonate reservoirs – I. Journal of Petroleum Geology, 8(1): 37-58.

Bertani, R.T. & Carozzi A.V. 1985b. Lagoa Feia Formation (Lower Cretaceous) Campos Basin, offshore Brazil: Rift valley type lacustrine carbonate reservoirs – I. Journal of Petroleum Geology, 8(2): 199-220.

Bernet, M. & Gaupp, R. 2005. Diagenetic history of Triassic sandstone from the Beacon Supergroup in central Victoria Land, Antarctica. New Zealand Journal of Geology & Geophysics, 48: 447–458.

Blatt, H., 1979, Diagenetic processes in sandstone, in Scholle, P. A., and Schluger, P. R., eds., Aspects of Diagenesis, Volume 29: Tulsa, Ok, Society of Economic Paleontologists and Mineralogists, p. 141-158.

Boles, J.R. & Franks, S.G. 1979. Clay diagenesis in Wilcox sandstones of southwest Texas: implications of smectite diagenesis on sandstone cementation: Journal of Sedimentary Petrology, 49(1): 55-70.

Bradley, W.H. & Fahey, J.J. 1962. Occurrence of stevensite in the Green River Formation of Wyoming. American Mineralogist, 47: 996-998.

Bruhn, C.H.L. 1998. “Deep-Water Reservoirs from the Eastern Brazilian Rift and Passive Margin Basins”, AAPG International Conference & Exhibition, Rio de Janeiro, Brazil, Short Couse Notes, Part 2, p187.

Burns, L.K. & Ethridge, F.G. 1979. Petrology and diagenetic effects of lithic sandstones: Paleocene and Eocene Umpqua Formation, southwest Oregon. In: Scholle, P.A. & Schluger, P.R. (Eds.). Aspects of Diagenesis, SEPM Special Publication 26: Tulsa, OK, Society of Economic Paleontologists and Mineralogists, p. 307-317.

Buch M.W. & Rose D. 1996. Mineralogy and geochemistry of the sediments of the Etosha Pan Region in northern Namibia: a reconstruction of the depositional environment. Journal of African Earth Sciences, 22(3): 355-378.

Carvalho, M.D., Praça, U.M., Silva-Telles, A.C., Jahnert, R.J. & Dias, J.L. 2000. Bioclastic Carbonate Lacustrine Facies Models in the Campos Basin (Lower Cretaceous), Brazil. In: Gierlowski-Kordesch, E.H. & Kelts, K.R. (Eds.). Lake Basins Through Space and Time: AAPG Studies in Geology, p. 245-256.

Castro, J.C. 2006. Evolução dos conhecimentos sobre as coquinas-reservatório da Formação Lagoa Feia no trend Badejo-Linguado-Pampo, Bacia de Campos. Geociências, UNESP, 25(2): 175–186.

Cerling, T.E. 1996. Pore water chemistry of an alkaline lake; Lake Turkana, Kenya. In: Johnson, T.C. & Odada, E.O. (Eds.). The limnology, climatology and paleoclimatology of the East African lakes. Gordon and Breach Publishers, Australia, p. 225-240.

Chahi, A., Fritz-B., DuPlay, J., Weber, F. & Lucas, J. 1997. Textural transition and genetic relationship between precursor stevensite and sepiolite in lacustrine sediments (Jbel Rhassoul, Morocco). Clays and Clay Minerals, 45(3): 378-389.

Chahi, A., Duringer, P., Ais, M., Bouabdelli, M., Gauthier, L. F. and Fritz, B., 1999, Diagenetic transformation of dolomite into stevensite in lacustrine sediments from Jbel Rhassoul, Morocco. Journal of Sedimentary Research, v. 69, n. 5, p. 1123-1135.

Chan, M.A. 1985. Correlations of diagenesis with sedimentary facies in Eocene sandstones, western Oregon: Journal of Sedimentary Petrology, 55: 322-333.

Chang, H.K., Mackenzie, F.T. & Schoonmaker, J. 1986. Comparisons between the diagenesis of dioctahedral and trioctahedral smectite, Brazilian offshore basins: Clays and Clay Minerals, 34: 407-423.

Chang H.K., Kowsmanm R.O. & Figueiredo A.M.F. 1990. Novos conceitos sobre o desenvolvimento das bacias marginais do leste brasileiro. In: Raja Gabaglia G.P. & Milani E.J. (Eds.). Origem e Evolução de Bacias Sedimentares. Petrobras, Rio de Janeiro, p. 269-289.

Crossey, L.J., Frost, B.R. & Surdam, R.C. 1984. Secondary porosity in laumontite-bearing sandstones. In: Surdam, R.C. & McDonald, D.A. (Eds.). Clastic Diagenesis. AAPG Memoir 37, American Association of Petroleum Geologists, Tulsa, OK, p. 225-237.

Cuevas, J., de la Villa, R., Ramirez, S., Petit, S., Meunier, A. & Leguey, S. 2003. Chemistry of Mg smectites in lacustrine sediments from the Vicalvaro sepiolite deposit, Madrid Neogene Basin (Spain). Clays and Clay Minerals, 51(4) p. 457-472.

Darragi, F. & Tardy, Y. 1987. Authigenic trioctahedral smectites controlling pH, alkalinity, silica and magnesium concentrations in alkaline lakes. Chemical Geology, 63: 59-72.

De Ros, L.F. 1996. Compositional Controls on Sandstone Diagenesis: Compr. Summ. Uppsala Diss. Facul. Sci. Tech., 198: 1-24.

De Ros, L.F., Goldberg, K., Abel, M., Victorinetti, F., Mastella, L. & Castro, E. 2007. Advanced Acquisition and Management of Petrographic Information from Reservoir Rocks Using the PETROLEDGE® System: AAPG Annual Conference and Exhibition, Long Beach, CA, Extended Abstracts Volume, 6 pp.

Dias, J.L., Oliveira, J.Q. & Vieira, J.C. 1988. Sedimentological and Stratigraphyc analysis of the Lagoa Feia Formation, rift phase of Campos Basin, offshore, Brazil. Revista Brasileira de Geociências, 18(3): p. 252–260.

Dias, J.L., Scarton, J.L., Esteves, F.R., Carminatti, M. & Guardado, L.R. 1990. Aspectos da Evolução Tectono-Sedimentar e a Ocorrência de Hidrocarbonetos na Bacia de Campos, in: Raja Gabaglia G.P. & Milani E.J. (Eds.). Origem e Evolução de Bacias Sedimentares, Petrobras, Rio de Janeiro, p. 333-360.

Dickinson, W.R. 1985. Interpreting provenance relations from detrital modes of sandstones. In: Zuffa, G.G. (Ed.). Provenance of arenites. NATO-ASI Series C.D. Reidel Pub. Co., Dordrecht, Netherlands, p. 333-361.

Dyni, J.R. 1976. Trioctahedral smectite in the Green River Formation, Duchesne County. Utah. U.S.G.S. Professional Paper, 967: 14.

Eberl, D.D. Jones, B.F. & Khoury, H.N. 1982. Mixed-layer kerolite/stevensite from the Amargosa Desert, Nevada. Clays and Clay Minerals, 30(5): 321-326.

Figueiredo, A.M.F. & Mohriak, W.U. 1984. A tectônica salífera e as acumulações de petróleo da Bacia de Campos. in: XXXIII CONGRESSO BRASILEIRO DE GEOLOGIA, 1984, SBG, p.1380-1394.

Furquim, S.A.C., Graham, R.C., Barbiero, L., Queiroz Neto, J.P. & Vallès V. 2008. Mineralogy and genesis of smectites in an alkaline-saline environment of Pantanal Wetland, Brazil. Clays and Clay Minerals, 56(5): 579-595.

Galloway, W. E., 1974, Deposition and diagenetic alteration of sandstones in northwest Pacific arc-related basins: Implications for greywacke genesis: Geol. Soc. Am. Bull. 85, p. 379-390.

Galloway, W.E. 1979. Diagenetic controls of reservoir quality in arc-derived sandstones: Implications for petroleum exploration. In: Scholle, P.A. & Schluger, P.R. (Eds.). Aspects of diagenesis, SEPM Special Publication 26, p. 251-262.

Guardado, L.R., Gamboa, L.A.P. & Luchesi, C.F. 1989. Petroleum Geology of the Campos Basin, a Model for a Producing Atlantic-Type Basin. In: Edwards J.D. & P.A. Santogrossi, (Eds.). Divergent/Passive Margin Basins: AAPG Memoir 48, p.3-79.

Guardado, L.R., Gamboa, L.A.P. & Lucchesi, C.F., 1990. Petroleum geology of the Campos Basin, Brazil, a model for a producing Atlantic type basin. In: Edwards, J.D and Santogrissi, P.A., Divergent / Passive Margin Basins. Tulsa, Okla., AAPG Memoir 48, p. 3-79.

Guardado, L.R., Spadini, A.R., Brandão, J.S.L. and Melo, M.R. 2000. Petroleum System of the Campos Basin. In: Mello, M.R. & Katz, B. (Eds.). Petroleum System of South Atlantic Margins: AAPG Memoir 73, p. 317–324.

Hay, R.L. 1966. Zeolites and zeolitic reactions in sedimentary rocks: Geological Society of America Special Publication, 85: 130.

Hay R. L. and Sheppard, R.A., 2001, Occurrences of zeolites in sedimentary rocks. In: Bish, D.L., Ming, D.W. (Eds) Natural Zeolites: occurrence, properties; applications. Reviews in Mineralogy & Geochemistry, Vol. 45, Mineralogical Society of America, Washington, DC, p. 217-234.

Hover, V.C. & Ashley, G.M. 2003. Geochemical signatures of paleodepositional and diagenetic environments; a STEM/ AEM study of authigenic clay minerals from an arid rift basin, Olduvai Gorge, Tanzania. Clays and Clay Minerals, 51(3): 231-251.

Hover, V.C. Walter, L.M. Peacor, D.R. & Martini, A.M. 1999. Mg-smectite authigenesis in a marine evaporative environment, Salina Ometepec, Baja California. Clays and Clay Minerals, 47(3): 252-268.

Humphreys, B., Kemp, S.J., Lott, G.K., Bermanto, D.A., Dharmayanti, D.A. & Samsori, I. 1994. Origin of grain coating chlorite by smectite transformation: an example from Miocene sandstones, North Sumatra back-arc basin, Indonesia. Clay Minerals, 29(4): 681–692.

Iijima, A., 1980. Geology of zeolites and zeolitic rocks. Pure and Applied Chemistry, 52: 2115-2130.

Jeans, C.V., Wray, D.S., Merriman, R.J., & Fisher, M. J. 2000. Volcanogenic clays in Jurassic and Cretaceous strata of England and the North Sea Basin: Clay Minerals, 35: 25–55.

Jones, B.F. & Weir, A.H. 1983. Clay minerals of Lake Abert, an alkaline, saline lake. Clays and Clay Minerals, 31(3): 161-172.

Jones, B.F. 1986. Clay mineral diagenesis in lacustrine sediments. U.S. Geological Survey Bulletin, 1578: 291-300.

Kahn, J.S. 1956. The analysis and distribution of the properties of packing in sand-size sediments: 1. On the measurement of packing in sandstones: Journal of Geology, 64: 385-395.

Khoury, H.N., Eberl, D.D. & Jones, B.F. 1982. Origin of magnesium clays from the Amargosa Desert, Nevada. Clays and Clay Minerals, 30(5): 327-336.

Martin de Vidales, J.L., Pozo, M, Alia, J.M., Garcia-Navarro, F. & Rull, F. 1991. Kerolite-stevensite mixed-layers from the Madrid Basin, central Spain. Clay Minerals, 26(3): 329-342.

Mathisen, M.E. 1984. Diagenesis of Plio-Pleistocene nomarine sandstones, Cagayan Basin, Philippines. Early development of secondary porosity in volcanic sandstones. In: McDonald D.A. & Surdam, R.C., (Eds.). Clastic Diagenesis, AAPG Memoir 37, p. 177-193.

Mathisen, M.E. & McPherson, J.G. 1991. Volcaniclastic deposits: Implications for hydrocarbon exploration. In: R. V. Fisher & G.A. Smith. (Eds.). Sedimentation in volcanic settings. SEPM Special Publication 45. Society of Economic Paleontologists and Mineralogists, Tulsa, Okla. p. 27-36.

Mayayo, M.J., Bauluz, B. & Gonzalez-Lopez, J.M. 2000. Variations in the chemistry of smectites from the Calatayud Basin (NE Spain). Clay Minerals, 35(2): 365-374.

McKinley, J.M., Worden, R.H. & Ruffell, A.H. 2003. Smectite in sandstones: a review of the controls on occurrence and behaviour during diagenesis. In: Worden, R.H., Morad, S. (Eds.). Clay Cements in Sandstones. International Association of Sedimentologists - Blackwell Scientific Publications, Oxford, UK, p. 109-128.

Mizusaki, A.M.P., Thomaz Filho A. & Valença J.G. 1988. Volcano-sedimentary sequence of Neocomian age in Campos Basin (Brazil). Revista Brasileira de Geociências, 18: 247-251.

Mizusaki, A.M.P., Petrini, R., Bellieni, G., Comin-Chiaramonti, P., Dias, J., De Min, A. & Piccirillo, E.M. 1992. Basalt magmatism along the passive continental margin of SE Brazil (Campos basin). Contribution to Mineralogy and Petrology, 111: 143-160.

Morad, S., Ketzer, J.M. & De Ros, L.F. 2000. Spatial and temporal distribution of diagenetic alterations in siliciclastic rocks: implications for mass transfer in sedimentary basins. Sedimentology, 47: 95-120.

Morad, S., Ketzer, J.M. & De Ros, L.F. 2012. Linking diagenesis to sequence stratigraphy: an integrated tool for understanding and predicting reservoir quality distribution. In: Morad, S., Ketzer, J.M., & De Ros, L.F. (Eds.). Linking Diagenesis to Sequence Stratigraphy, IAS Special Publication 45. Chichester, UK, International Association of Sedimentologists. Wiley-Blackwell, p. 1-36.

Noh, J.H. & Boles, J.R. 1993. Origin of zeolite cements in the Miocene sandstones, North Tiejon oil fields, California. Journal of Sedimentary Petrology, 63: 248-260.

Ponte, E.C., & Asmus, H.E. 1978. Geological framework of the Brazilian continental margin, Geologische Rundschau, 68: 201-235.

Pozo, M. & Casas, J. 1999. Origin of kerolite and associated Mg clays in palustrine-lacustrine environments; the Esquivias Deposit (Neogene Madrid Basin, Spain). Clay Minerals, 34(3): 395-418.

Rangel, H.D., Martins, F.A., Esteves, F.R. & Feijó, F.J. 1994. Bacia de Campos. Boletim de Geociências da Petrobras, 8(1): 203-217.

Rangel, H.D. & Martins, C.C. 1998. Main exploratory compartments, Campos Basin. In: Searching for Oil and Gas in the Land of Giants. The Search. Schlumberger, Rio de Janeiro, p. 32-40.

Reed, J.K., Gipson, M.Jr. & Vass, D. 1993. Hydrocarbon potential of sandstone reservoirs in the Neogene east Slovakian Basin part 1: A petrographic examination of lithology, porosity. Journal of Petrol. Geol. 16: 89-108.

Rehim, H.A.A., Pimentel, A.M., Carvalho, M.D. & Monteiro, M. 1986. Talco e estevensita na Formação Lagoa Feia da bacia de Campos - possíveis implicações no ambiente deposicional. Anais XXXIV CONGRESSO BRASILEIRO DE GEOLOGIA, 1986, 1: 416-422.

Remy, R.R. 1994. Porosity reduction and major controls on diagenesis of Cretaceous- Paleocene volcaniclastic and arkosic sandstone, Middle Park Basin, Colorado. Journal of Sedimentary Petrology, 64: 797-806.

Schaller, H. 1973. Estratigrafia da Bacia de Campos. In: 27º CONGRESSO BRASILEIRO DE GEOLOGIA, Aracaju. Sociedade Brasileira de Geologia. Anais, v. 3, p. 247-258.

Schmidt, V. & Macdonald, D.A. 1979. The role of secondary porosity in the course of sandstone diagenesis. In: Aspects of Diagenesis. P.A. Scholle & P.R. Schluger (Eds.). SEPM Special Publication 29, Society of Economic Paleontologists and Mineralogists, Tulsa, Okla., 175–207.

Seemann, U. & Scherer, M. 1984. Volcaniclastics as potencial hydrocarbon reservoirs: Clay Minerals, 9: 457-470.

Stanley, K.O. & Benson, L.V. 1979. Early diagenesis of High Plains Tertiary vitric and arkosic sandstone, Wyoming and Nebraska. In: Scholle, P.A. & Schluger, P.R. (Eds.). Aspects of Diagenesis. SEPM Special Publication 26, Society of Economic Paleontologists and Mineralogists, Tulsa, Okla., p. 401-423.

Surdam, R.C. & Boles, J.R. 1979. Diagenesis of volcanic sandstones. In: Scholle, P. A. and Schluger, P.R., (Eds.). Aspects of Diagenesis. SEPM Special Publication 26, Society of Economic Paleontologists and Mineralogists, Tulsa, Okla., p. 227-242.

Tang, Z., Parnell, J. & Ruffell, A.H. 1994. Deposition and diagenesis of lacustrine-fluvial Cangfanggou Group (Uppermost Permian to Lower Triassic), southern Junggar Basin, NW China: A contribution from sequence stratigraphy: Journal of Paleolimnology, 1: 67-90.

Taylor, M.W. & Surdam, R.C. 1981. Zeolite reactions in the tuffaceous sediments at Teels Marsh, Nevada. Clays and Clay Minerals, 29: 341-352.

Tettenhorst, R. & Moore, G.E. Jr. 1978. Stevensite oolites from the Green River Formation of central Utah. Journal of Sedimentary Petrology, 48(2): 587-594.

Yuretich, R.F. & Ervin, C.R. 2002. Clay minerals as paleoenvironmental indicators in two large lakes of the African Rift Valleys: Lake Malawi and Lake Turkana. In: Sedimentation in Continental Rifts. SEPM Special Publication 73: 221-232.

Zuffa, G.G. 1985. Optical analysis of arenites: influence of methodology on compositional results. In: Zuffa, G. G., (Ed.). Provenance of Arenites, Volume 148: Dordrecht, The Netherlands, D. Reidel Pub. Co., p. 165-189.

Winter, W.R., Jahnert, R.J. & França, A.B. 2007. Bacia de Campos. Boletim de Geociências da Petrobras, 15(2): 511–529.

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# Deposition, Diagenesis and Reservoir Potential of Non-Carbonate Sedimentary Rocks from the Rift Section of the Campos Basin, Brazil

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## Abstract

A petrographic study was carried out on the rift section of the Lagoa Feia Group, Lower Cretaceous of the Campos Basin, eastern Brazilian margin, as part of a regional, integrated project. The main primary constituents of the analyzed rocks are siliciclastic and volcaniclastic grains, stevensitic ooids and peloids, and bivalve and ostracod bioclasts. This study focused in the clastic, stevensitic and hybrid rocks, as previous studies were limited to the bioclastic rudstones and grainstones that constitute the producing reservoirs. Most rift sedimentation was intrabasinal, with extrabasinal contribution concentrated at the proximity of rifted blocks along border faults. In the clastic sandstones and conglomerates, the mixture of rounded volcanic fragments with angular quartz, feldspars and plutonic fragments indicates the recycling of early rift epiclastic deposits, combined with first-cycle sediments eroded from uplifted granitic-gneissic basement blocks. Stevensitic ooids and peloids were formed in shallow, alkaline lacustrine environments, slightly agitated by waves or currents. They were mixed throughout the rift section with bivalve and ostracod bioclasts, and with clastic sediments. This gravitational re-deposition was promoted by intense and recurrent tectonic movements along the margins of rifted structural blocks. The main diagenetic processes in clastic sandstones and conglomerates and

hybrid arenites are cementation and grain replacement by smectite, zeolites, calcite and dolomite, limited compaction and dissolution of feldspars, volcanic fragments and bioclasts. Stevensitic arenites experienced early cementation and replacement of ooids and peloids by quartz, calcite and dolomite, or intense compaction of the ductile stevensitic grains in uncemented areas. Volcaniclastic sandstones and conglomerates, with intergranular porosity partially reduced by smectite rims and some grain dissolution, may constitute fair hydrocarbon reservoirs. Stevensitic and hybrid arenites with dissolution of stevensitic grains, bioclasts and calcite cement may also constitute reservoirs, with potential quality limited by the poor connection of such pore systems. The understanding of the space and time controls on the depositional and diagenetic evolution of the dominantly intrabasinal, gravitationally re-deposited rift carbonate and non-carbonate rocks will contribute to new exploration strategies for the Campos Basin.

## 1. Introduction

The Campos Basin lies at the eastern Brazilian margin, along the north coast of Rio de Janeiro State. It is separated from the Espírito Santo Basin to the North by the Vitória Arch, and from the Santos Basin to the South by the Cabo Frio Arch, covering an area of approximately 120,000 km$^2$ with only a small onshore area (5,800 km$^2$) (Figure 1). According to the National Petroleum Agency (ANP, 2013), the Campos Basin is the most prolific Brazilian basin, corresponding to 84% of oil and gas production in Brazil. Hydrocarbons are sourced in the rift (sub-salt) section of Lower Cretaceous Lagoa Feia Group (Guardado et al., 2000). Main producing reservoirs in this unit correspond to bioclastic, lacustrine limestones, called "coquinas" (Bertani and Carozzi 1985a; 1985b; Dias et al., 1988; Abrahão and Warme, 1990; Carvalho et al., 2000).

Due to the importance of oil production in the "coquinas", previous work on primary composition and diagenetic evolution of the Lagoa Feia deposits focused on carbonate rocks (Carvalho et al., 2000; Castro, 2006). This work aims at analyzing the genesis of non-carbonate sedimentary rocks in the Lagoa Feia Group, including volcaniclastic sandstones and conglomerates, hybrid and stevensitic arenites. This petrologic study was carried out as part of a larger project developed in partnership with the BG Group, which also involved sedimentologic, stratigraphic and seismic analyses intended to better understand the depositional setting, diagenetic processes and patterns, and potential reservoir quality of these rocks.

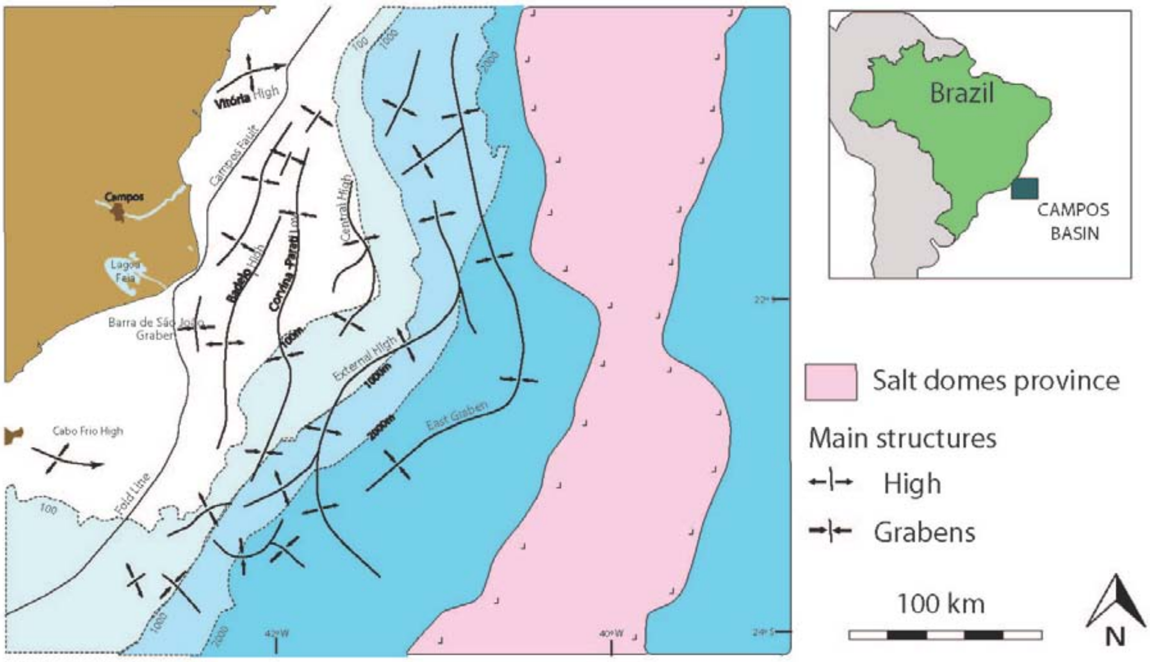


Figure 1. Localization map of the Campos Basin. Modified from Guardado et al. (2000).

## 2. Geological setting

The Campos Basin was formed in the Early Cretaceous as a consequence of Gondwana breakup, when the South American and African plates were separated by a rift system that propagated from south to north. The breakup led to the formation of several rifts with different timing and fillings, often with basaltic floods in the onset, culminating with crustal rupture and expansion of the ocean floor (Milani and Thomaz Filho, 2000).

The rift process reactivated regional, extensional NE/SW structures from the Proterozoic basement (formed during the Brazilian orogenic cycle), composed mainly of gneissic and granitic rocks from the Ribeira Province (Almeida and Carneiro, 1998). These weakening zones defined where reactivation took place during the opening of the Atlantic Ocean, controlling rift architecture (Heilbron et al., 2008).

In the initial stages, continental blocks were uplifted and normal-faulted, leading to an irregular geometry which alternated horsts and grabens filled with lacustrine and alluvial sediments, as well as basaltic floods. Sedimentation in the lower, Neocomian rift section (Pre-Alagoas) was dominantly lacustrine. During the Aptian (Alagoas local stage), the onset of the flexural stage in the basin led to the deposition of the post-rift units, culminating with marine ingression and the deposition of evaporites.

Some important tectonic structures developed in the basin during rifting include: 1) the Campos Fold line/fault, which separates areas with shallow (to the west) from deep basement (to the east); 2) The Badejo Regional High, a north-dipping horst that strongly controls sedimentary distribution; 3) the Pre-Alagoas

(Aptian) unconformity, marking the change from a rift to a post-rift tectonic pattern, characterized by faulting related to halokinesis (Dias et al., 1988). The disconformity was formed during a stage of regional uplifting and truncation.

The sedimentary infill of the Campos Basin was subdivided by Rangel et al., (1994) into 3 tectono-stratigraphic sequences, bound by erosive unconformities: the non-marine, transitional and marine sequences. The non-marine, rift sequence was formed in the initial stages of rifting. Part of the basin was filled with the subalkaline, subaerial and subaqueous, basaltic floods of the Cabiúnas Formation, unconformably overlying the Precambrian basement. This phase is characterized by synthetic and antithetic faulting, with the formation of several grabens and half-grabens with rotated blocks due to fast crustal subsidence. Rift architecture was strongly influenced by pre-existing Late Proterozoic structures inherited from the Brazilian Orogeny, which defined weakening zones that were ruptured during the opening process. The main facies associations are alluvial fan, fan delta and transitional deposits, lacustrine marls, shales and bivalve and ostracod carbonates that comprise the Lagoa Feia Group (Schaller, 1973; Dias et al., 1988; Winter et al., 2007). Sedimentary filling took place in an alkaline, lacustrine environment with strong tectonic control (Dias et al., 1988). The top of this sequence is relatively flat, indicating that the grabens of the rift systems had almost been filled up by the end of alluvial-lacustrine deposition (Abrahão and Warme, 1990).

Differential subsidence during the deposition of the Lagoa Feia Group resulted in areas with very thick alluvial/lacustrine deposits (Abrahão and Warme, 1990) (Figure 2). Along proximal margins in the rifts, alluvial fan deposits composed of conglomerates and sandstones rich in volcanic fragments dominated (Itabapoana

Formation). Fine-grained facies accumulated in lacustrine depocenters, where anoxic conditions allowed the deposition of organic-rich mudrocks. These are the main hydrocarbon source rocks in the Campos Basin (Mohriak et al.,1990; Mello et al.,1994; Guardado et al., 2000). They comprise the Atafona Formation, along with sandstones and siltstones with abundant stevensitic clay minerals, deposited in alkaline environments (Winter et al., 2007).

Accumulations of bivalve and ostracod bioclasts (“coquinas”), intercalated with mud deposits, occur along the flanks and crests of internal highs in the rifts, away from areas with terrigenous input (Bertani and Carozzi, 1985a; 1985b; Carvalho et al., 2000). These correspond to the Coqueiros Formation. The “coquinas” and fractured basalts (Mizusaki et al., 1988) are the presently producing reservoirs in the rift section of the Campos Basin.

The transitional sequence corresponds to the beginning of the drift stage, in a period of tectonic readjustment that marked the termination of erosion in uplifted and rotated blocks. The lower sequence is composed of conglomerates and sandstones that overlie the unconformity at the end of the rift stage, as well as stevensitic mudrocks and carbonate precipitates that constitute the pre-salt reservoirs (Carminatti et al., 2008; Wright, 2012). The upper sequence comprises evaporite deposits of the Retiro Formation (Neoaptian), mostly halite and anhydrite (Dias et al., 1988). Salt tectonics related to this unit affected the overlying rocks, forming a series of listric faults in evacuation zones, on slopes of basins surrounded and cut by salt domes. In the Campos Basin, most faults related to salt tectonics are synthetic. Salt movement in this stage molded the seafloor, creating preferential pathways for the sands deposited in the Upper Cretaceous (Winter et al., 2007). In the

Neocretaceous, high subsidence rates resulted in a starved depositional environment, with little sedimentary input in the basin, bypass or erosional events and halokinesis that resulted in faulting and depressions where sand was deposited (Dias et al., 1988). Throughout the sequence tuffs and dolerite sills occur, mostly in the south. These volcanic events are related to extension associated with E-W transform faults.

The drift stage started in the Albian, and the sedimentary deposits were strongly affected by salt tectonics (salt domes). The sequence starts with the gradational change from a transitional, evaporitic to marine environment, with the formation of the shallow, platform/ramp carbonate deposits of the Quissamã Formation, Eo-Albian Macaé Group, grading upwards into marls and shales of the Outeiro Formation (Neo-Albian). Concomitant with the formation of these deposits, the sandy turbidites of the Namorado Formation are formed, as well as polymictic conglomerates and sandstones of the Goitacás Formation in the proximal areas. The decreasing thermal anomaly formed during crustal stretching and increasing distance from the mid-Atlantic ridge resulted in mainly slight thermal subsidence towards offshore.

In the Middle Cretaceous a fully marine environment is established in the basin, leading to deposition of a thick succession of siliciclastic and carbonate sediments of the Campos Group, which includes the Ubatuba, Carapebus and Emboré Formations. The Ubatuba Formation comprises deep-water shales, marls, calcilutites and diamictites. The Carapebus Formation, interbedded with the Ubatuba Formation shales, is composed of fine-grained sandstones to conglomerates deposited by turbidity currents in the slope. The Emboré Formation comprises coarse-grained sediments, deposited by coastal fans, and carbonate platform deposits.

Tertiary sedimentation includes prograding sequences formed by siliciclastic and carbonate platform sediments of the Emboré Formation and slope, muddy deposits of the Ubatuba Formation, intercalated with turbiditic sandstones of the Carapebus Formation. This stage is characterized by high sedimentation and subsidence rates, which allowed the deposition of a thick siliciclastic package.

The main hydrocarbon reservoirs in the basin are the Upper Cretaceous to Lower Tertiary marine turbiditic sandstones of the Carapebus Formation (Guardado et al., 1990; Mello et al*.,* 1994). Other reservoirs correspond to fractured basalts of the Cabiúnas Formation, “coquinas” of the Coqueiros Formation and carbonate precipitates of the Macabu Formation (Lagoa Feia Group), carbonate rocks of the Quissamã Formation and turbiditic sandstones of the Namorado Formation (Macaé Group).

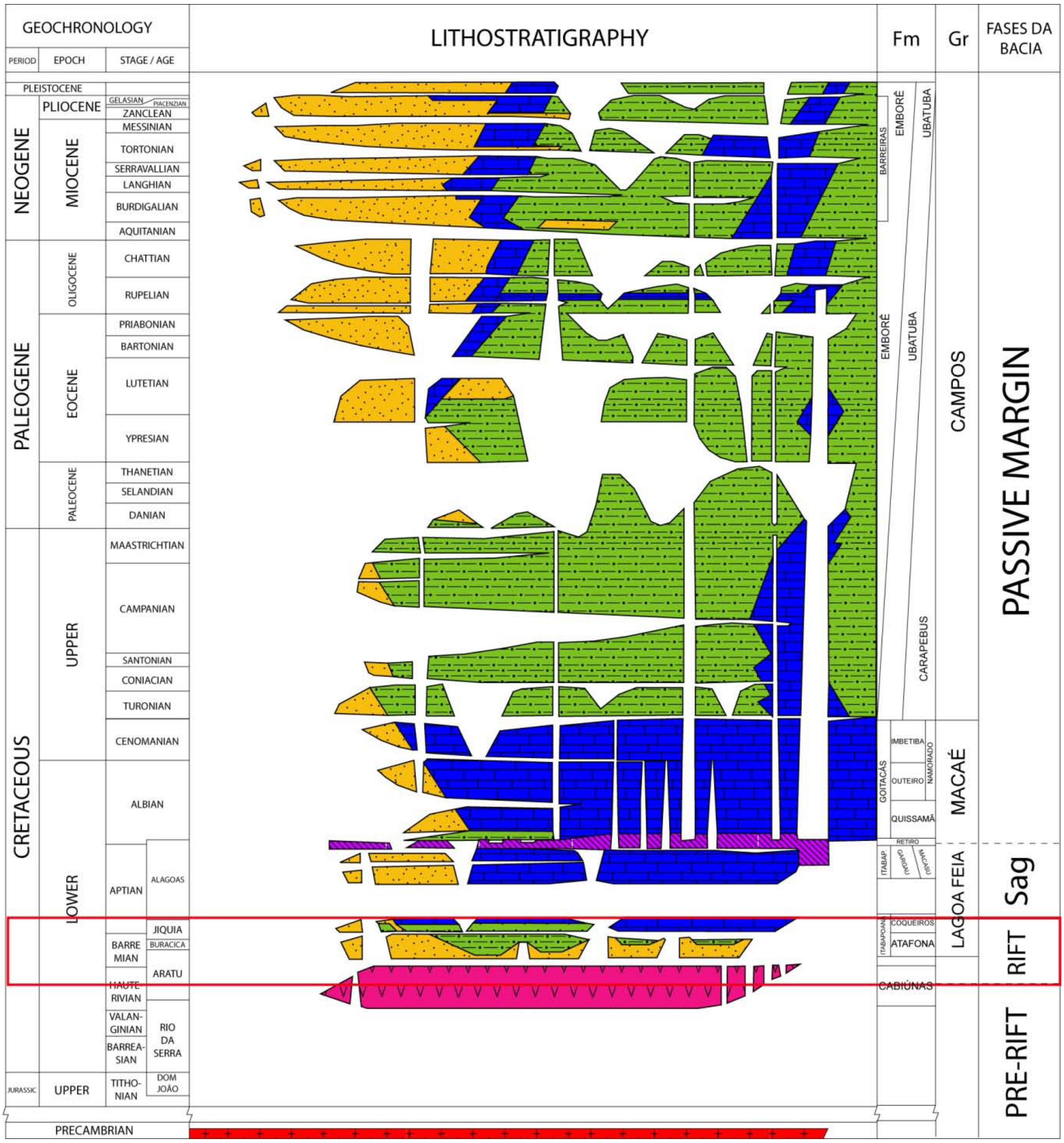


Figure 2. Stratigraphic chart of the Campos Basin (Winter et al., 2007), with the studied interval (red square).

## 3. Methods

In this study, 42 thin sections, prepared from samples taken from cores of 10 wells, were analyzed and quantified. The samples were impregnated with blue epoxy resin, and the thin sections were stained with a solution of alizarine red and potassium ferrocyanide (cf. Dickson, 1965) for identification of the carbonate minerals. Quantification was performed by counting 300 points in each thin section along transects perpendicular to the rock structure and fabric. Textural aspects of grain size, shape and fabric, as well as the composition, locations and the paragenetic relationships among primary and diagenetic constituents and pore types were recorded with use of the Petroledge software (De Ros et al., 2007). Scanning electron microscopy (SEM) analyses in the backscattered electrons (BSE) mode were performed for a better definition of the paragenetic relations among primary and diagenetic constituents on 7 selected thin sections using a JEOL JSM-6610LV microscope equipped with a Brucker energy dispersive spectrometer (EDS) for the identification of the elemental composition of the constituents.

## 4. Primary composition and texture

The primary composition of the analyzed samples includes siliciclastic, volcaniclastic, stevensitic and carbonate constituents. Siliciclastic grains (quartz, feldspars, micas, and heavy minerals grains, and plutonic rock fragments) are in general very angular (Figure 3A). Quartz grains are mostly monocrystalline, and microcline and plagioclase predominate among the feldspars. Micas are mostly represented by biotite, and amphiboles predominate among the heavy minerals. The most abundant rock fragments are basaltic, usually with hemicrystalline texture, and well rounded (Figure 3B). Siliciclastic mud is scarce in the analyzed cores. The presence of ooids or peloids of magnesian clay minerals identified in previous studies as stevensite, talc/stevensite mixed layers (Rehim et al., 1986), or kerolite (Bertani and Carozzi, 1985a; 1985b) is disseminated in most of the samples analyzed.

Stevensite is a smectitic clay mineral with strongly magnesian composition of general formula: $(Ca_{0.5}Na)_{0.33}(Mg,Fe^{++})_3Si_4O_{10}(OH)_2\bullet n(H_2O)$, or alternatively: $Na_{0.2}(Mg_{2.3}Al_{0.3}Fe_{0.1})Si_4O_{10}(OH)_2\bullet n(H_2O)$. Talc is a layer silicate with $Mg_3Si_4O_{10}(OH)_2$ formula, while kerolite is a hydrated version of talc that may contain some nickel, with formula: $Mg_3Si_4O_{10}(OH)_2\bullet nH_2O$, or: $(Mg,Ni)_3Si_4O_{10}(OH)_2\bullet nH_2O$. Stevensite is characteristically formed in alkaline and saline lakes, as well as in playas (Bradley and Fahey, 1962; Dyni, 1976; Tettenhorst and Moore, 1978; Eberl et al., 1982; Khoury et al., 1982; Jones and Weir, 1983; Jones, 1986; Martin de Vidales et al., 1991; Buch and Rose, 1996; Cerling, 1996; Chahi et al., 1997, 1999; Hover et al., 1999; Pozo and Casas, 1999; Mayayo et al., 2000; Yuretich and Ervin, 2002; Cuevas et al., 2003; Hover and Ashley, 2003).

In the analyzed rocks, stevensite occurs predominantly as ooids (Figure 3C), usually medium sand-sized and up to 2 cm in diameter, or as peloids with modal size of fine to very fine sand, which are more common in hybrid arenites. Irregular intraclasts made by agglomerates of these grains, or by stevensitic laminations, are much less common (Figure 3D).

The main carbonate constituents comprise bivalve and ostracod bioclasts. The bivalve bioclasts are almost always disarticulated and recrystallized, commonly broken (Figure 3E), and rarely abraded. The limited abrasion normally shown by the bivalve bioclasts indicates that the term “coquina” was incorrectly applied in the literature to these bioclastic deposits (e.g., Bertani and Carozzi, 1985a; 1985b; Abrahão et al., 1990; Carvalho et al., 2000). The ostracod bioclasts are commonly whole (Figure 3F), sometimes articulated, and rarely recrystallized. Other much less common carbonate grains include ooids (oncoids and oolites), intraclasts and peloids.

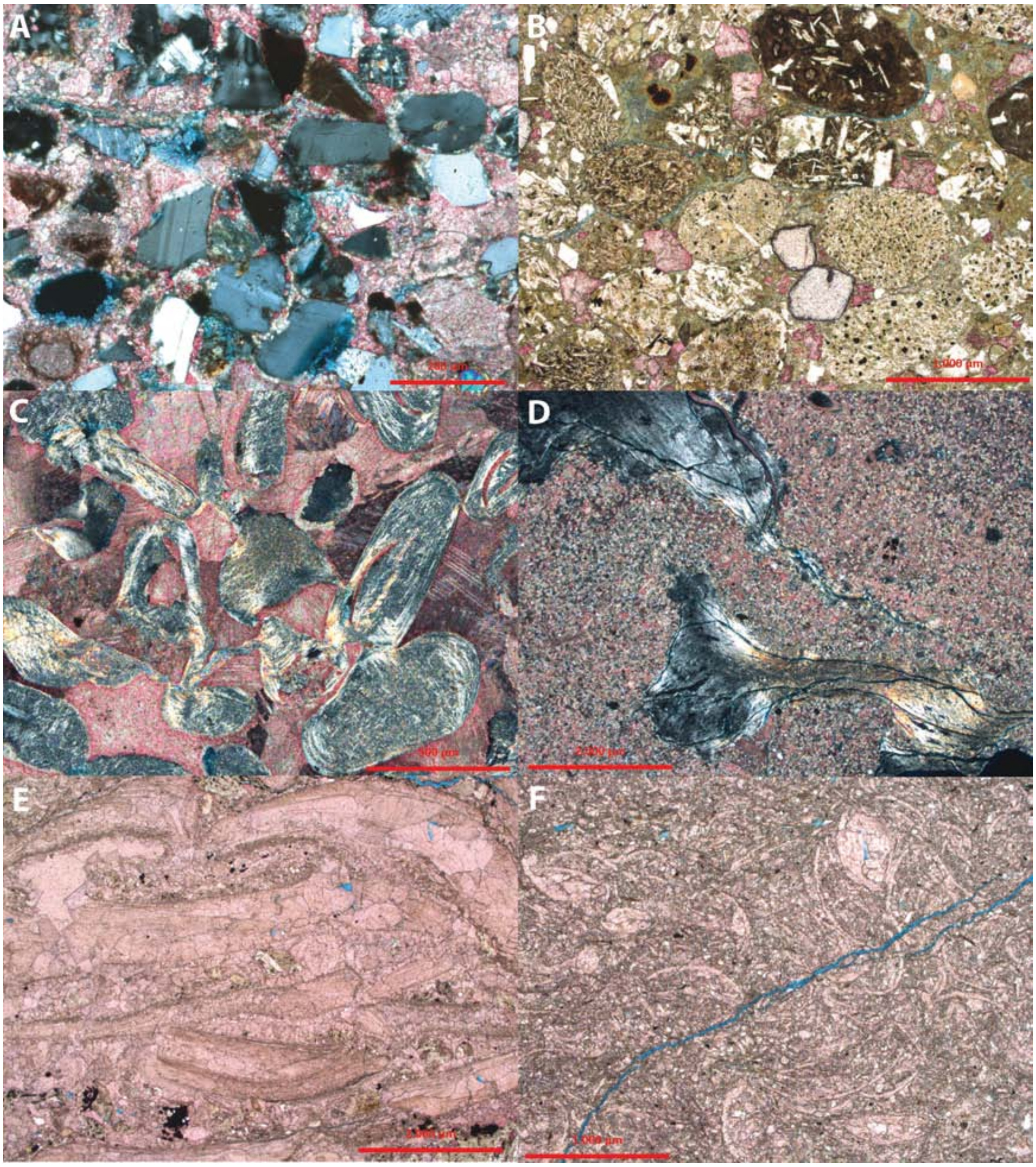


Figure 3. Photomicrographs of primary constituents. **A**) Angular feldspar and quartz grains. RJS114 2735,35. Crossed polarizers (XP). **B**) Rounded fragments of hemicrystalline and holocrystalline volcanic rocks and garnet grains, cemented by calcite (pink stained). RJS486A 2873,65. Plane polarizers (//P). **C**) Stevensite ooids partially replaced and cemented by calcite (pink stained). RJS514 4165,75. XP. **D**) Stevensite intraclasts cemented and replaced by calcite and dolomite. RJS514 4217,00. XP. **E**) Bivalve bioclasts cemented by drusiform calcite in rudstone. RJS379D 3100,25. //P. **F**) Ostracode bioclasts, some articulated, in grainstone. Pink-stained calcite. RJS386A 5240,75. //P.

Considering the importance of the stevensitic and carbonate intrabasinal grains and their pervasive mixture with clastic constituents throughout the analyzed wells, we decided to define the types of sedimentary rocks constituting the studied section in a CL-S-C diagram (Clastics-Stevensitics-Carbonatics). The main lithologic types comprise **clastic rocks**, which primary constituents correspond to more than 2/3 of siliciclastic and volcaniclastic grains, **carbonate rocks**, with more than 2/3 of carbonate allochem grains, **stevensitic rocks**, with more than 95% of stevensitic grains, **carbonate-stevensitic rocks**, made up by more than 2/3 of carbonate and stevensitic grains, but with less than 95% of the latter, and **hybrid rocks**, with a proportion between 1/3 and 2/3 of clastic grains to the sum of carbonate and stevensitic grains. The results of the petrographic quantification of the analyzed samples are displayed in the CL-S-C diagram (Figure 4).

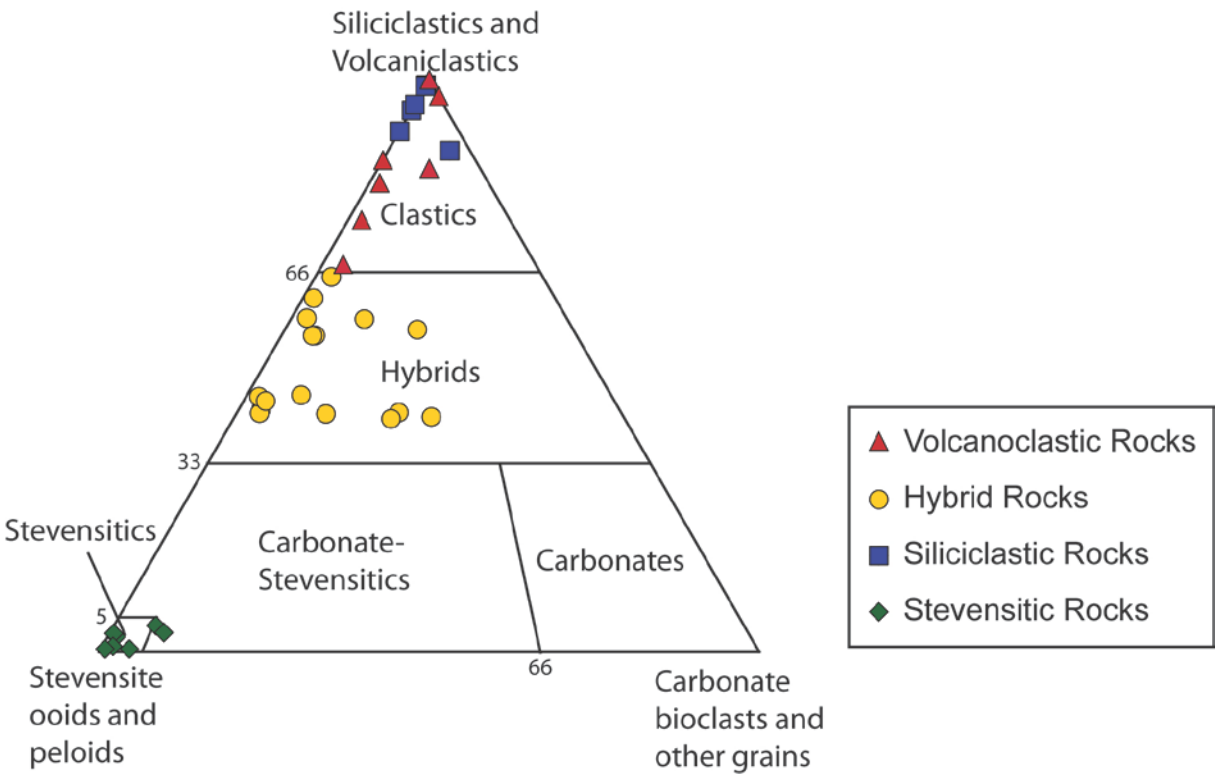


Figure 4. Compositional diagram with the main sedimentary rock types analyzed.

The **clastic rocks** occurring in the studied section correspond to siliciclastic sandstones, volcaniclastic conglomerates and sandstones, siliciclastic mudstones and shales, with limited contribution by intrabasinal carbonate or stevensitic grains. Siliciclastic sandstones are dominantly medium-grained (Figure 5A), and moderately sorted. Their occurrence is restricted, and commonly associated to volcaniclastic sandstones and conglomerates, or to siliciclastic mudrocks.

Sandstones and conglomerates in which most primary constituents correspond to volcaniclastic grains show a much wider distribution, occurring in most of the analyzed wells. The volcaniclastic sandstones range from medium- to very coarse-grained, and from moderately to very poorly sorted (Figure 5B). The volcaniclastic conglomerates display a sandy fraction equivalent to the associated sandstones, and a gravel fraction of rounded basalt and diabase fragments. Quartz grains are absent from the volcaniclastic conglomerates and coarse sandstones, occurring subordinately, only in the medium- and fine-grained sandstones. Quartz and feldspar grains are mostly angular. Volcaniclastic sandstones are massive, while some sandstones show irregular stratification.

Siliciclastic mudstones and shales are relatively scarce in the studied wells, comprising irregularly laminated, silty-sandy types with grains of feldspars, quartz, micas and heavy minerals, and more argillaceous types, parallel to irregularly-laminated and richer in organic matter (Figure 5C-D), which correspond to the main source rocks of the rift section. These fine-grained rocks show stylolites, carbonate concretions, and rare bioturbation, as well as ostracod and phosphate bioclasts.

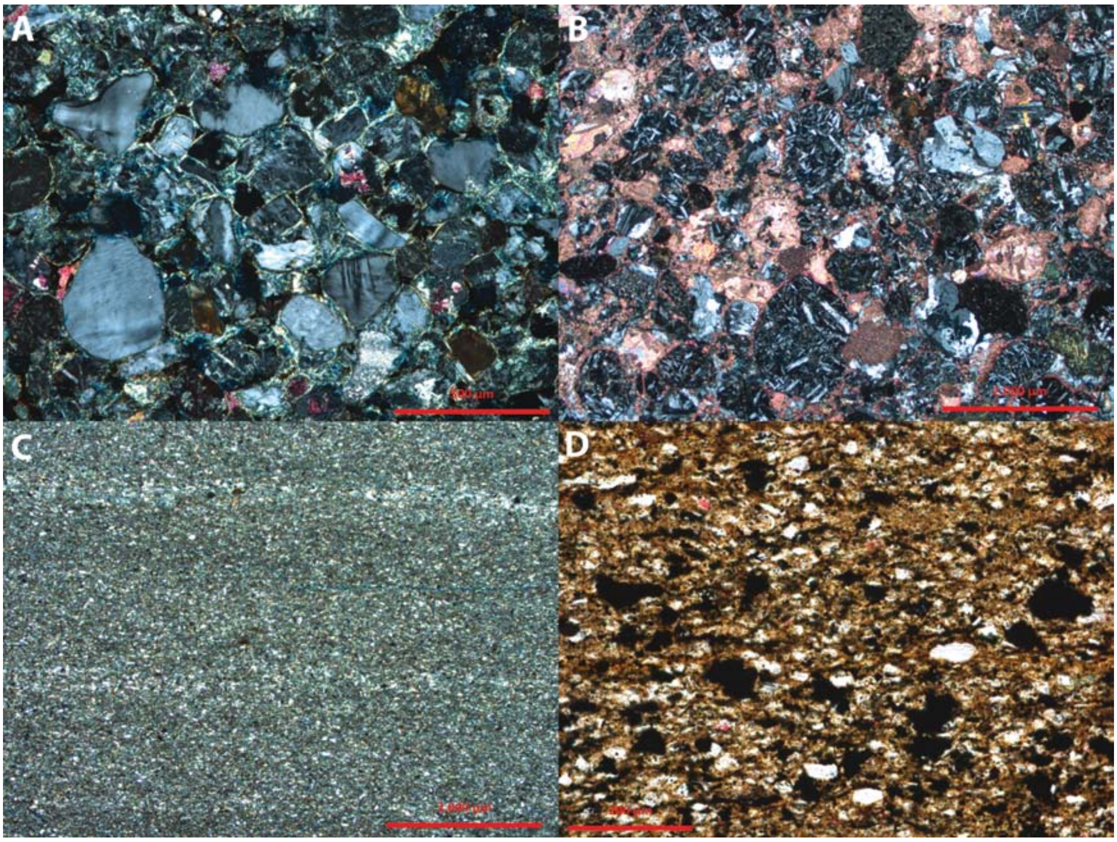


Figure 5. **A**) Medium-grained siliciclastic sandstone, poorly sorted, rich in feldspar. RJS113 2918,00. Crossed polarizers (XP). **B**) Coarse-grained volcaniclastic sandstone composed of volcanic fragments with hemicrystalline and holocrystalline textures and garnet grains. RJS486A 2873,65. Plane polarizers (//P). **C**) Micaceous sandy mudrock with irregular lamination. RJS114 2719,65. XP. **D**) Micaceous sandy mudrock with organic matter and replacive pyrite. RJS514 4102,50. PP.

Stevensitic grains are present in most of the analyzed samples, apparently as a product of their re-deposition in mixtures with clastic and carbonate constituents that are not compatible with their formation. Therefore, it was established that the truly **stevensitic rocks** should correspond essentially (> 95%) to stevensitic particles, in this case formed *in situ*. This category comprises arenites and conglomerates of stevensite ooids, and finer-grained arenites of stevensite peloids. Among the studied well, a significant presence of stevensitic arenites and conglomerates is restricted to RJS-514 well. Argillaceous stevensitic laminations are much less common than these types.

Ooidal stevensitic arenites are constituted by predominantly coarse sand-sized ooids (Figure 6A). Stevensitic conglomerates show ooids with up to 2 cm of diameter (pisoids), and irregular intraclasts of stevensite laminations or reworked, agglomerated ooids (Figure 6B). Stevensite peloids, ostracod and less often bivalve bioclasts occur as nuclei of ooids and pisoids. Quartz, feldspar and mica grains are rather rare in these rocks.

Pure stevensite laminites, constituted by syngenetic, wavy stevensite laminations, are almost absent from the analyzed section, but intercalations of stevensite laminations with stevensite, carbonate-stevensite and hybrid arenites, or with siliciclastic mudstones occur locally. Commonly, these laminations are deformed, folded or fragmented.

Stevensitic sandstones show massive, folded, patchy or fractured structure. Their packing ranges from loose to tight, often heterogeneously between cemented, less compacted, and non-cemented, compacted areas (Figure 6A, 6C). Cementation and replacement of the stevensite grains and laminations by dolomite, calcite, quartz and chalcedony is extensive in these rocks (Figure 6D).

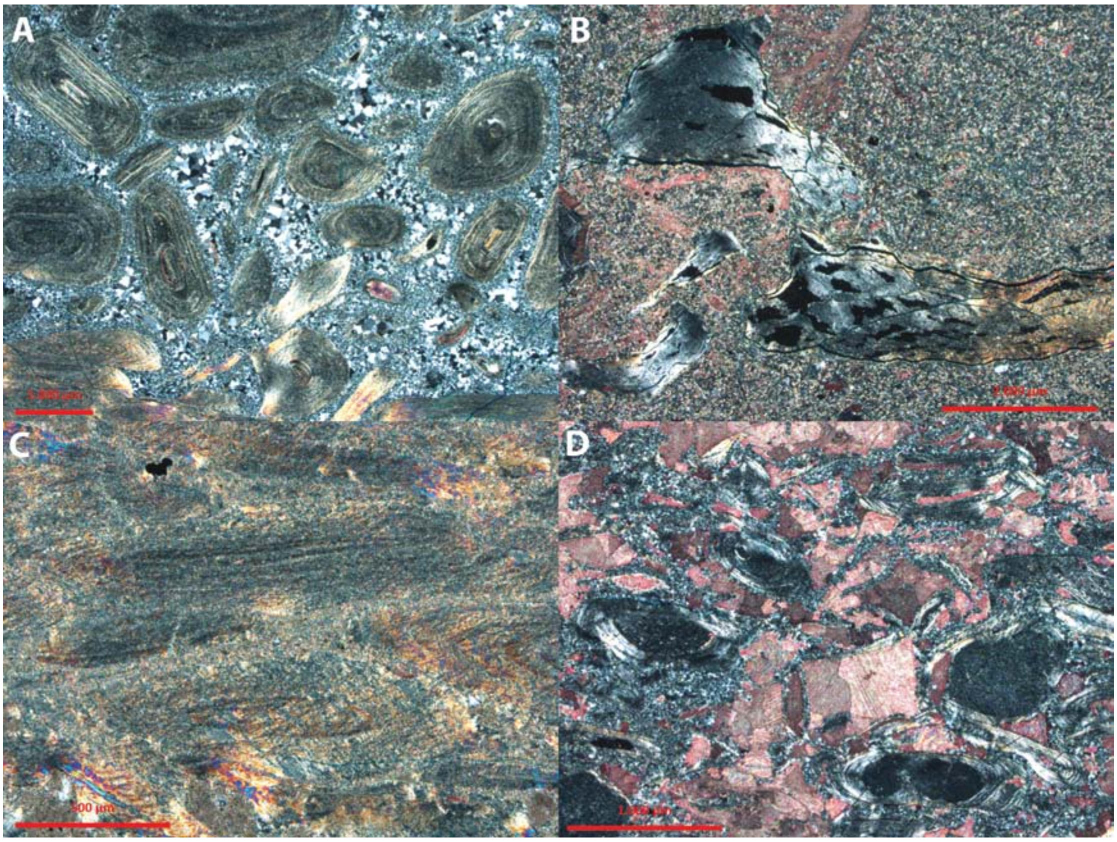

Figure 6. **A**) Sandy conglomerate with stevensite ooids cemented by drusiform silica. RJS514 4163,7. Crossed polarizers (XP). **B**) Sandy conglomerate with laminated stevensite intraclasts and reworked ooids. RJS514 4218,65. XP. **C**) Deform stevensite ooids by mechanincal compaction. RJS514 4163,7. XP. **D**) Stevensite ooids cemented and partially replaced by microcrystalline quartz and macrocrystalline calcite. RJS514 4220,75. XP.

The hybrid rocks, constituted by 1/3 to 2/3 of clastic grains, include bioclastic-clastic conglomerates and rudstones, and peloidal-ooidal arenites. The hybrid, bioclastic-clastic conglomerates/rudstones are made of similar proportions of bivalve bioclasts and siliciclastic-volcaniclastic fragments, also containing commonly stevensite ooids and peloids (Figure 7A-B). Hybrid arenites are constituted by stevensite peloids and ooids, and commonly also by ostracod bioclasts, in a proportion approximately equivalent to that of the siliciclastic and volcanic grains (Figure 7C-D).

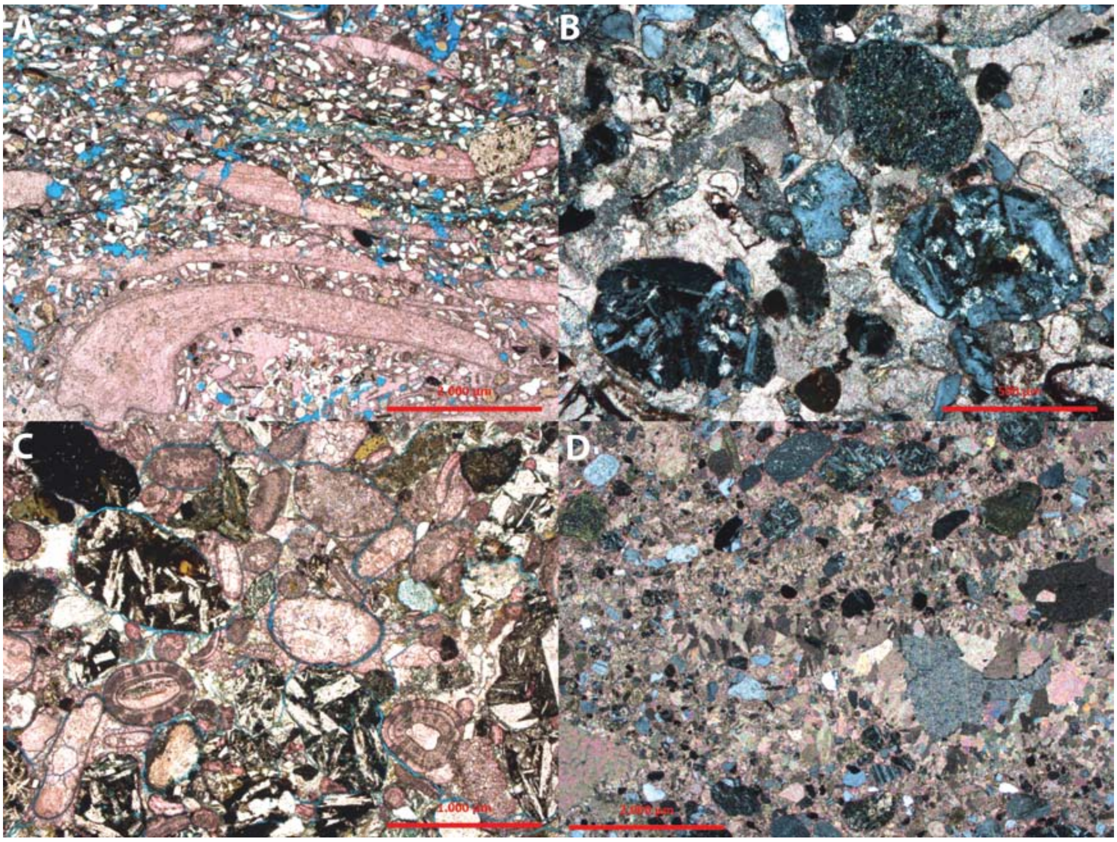


Figure 7. **A**) Rudstone/hybrid bioclastic-siliciclastic sandy conglomerate. RJS114 2735,35. Plane polarizers (//P). **B**) Bioclastic-volcaniclastic sandy conglomerate. RJS165 2425,10. Crossed polarizers (XP). **C**) Coarse-grained hybrid volcaniclastic-stevensitic sandstone with carbonate oolites, bivalve and ostracode bioclasts. RJS36 2965,60. //P. **D**) Hybrid stevensitic-siliciclastic-carbonate sandstone rich in recrystallized bivalve biclasts and volcanic fragments. RJS108 4531,75. XP.

## 5. Diagenetic constituents

Diagenetic constituents occurring in the studied rocks include calcite, smectite, zeolites, silica, dolomite, pyrite, titanium and iron oxides, gypsum, barite, albite, K-feldspar and kaolin. The percentages of these constituents in their diverse locations reported in this section refer to their averages (avg.) and maxima (max.) in bulk rock volume. Average percentages less than 1% are not reported.

## 5.1. Calcite

Calcite is the most abundant diagenetic constituent in the three defined lithologic types, occurring dominantly with a macrocrystalline, and subordinately with a microcrystalline, habit. Macrocrystalline calcite fills part of the intergranular pores in volcanic sandstones and conglomerates (avg.= 7.1%; maximum = 10%), and partially replaces intergranular authigenic smectite (avg.= 1.3%; max.= 4%) and zeolite (max.= 3.3%), volcanic fragments (avg.= 1.4%; max.= 5,3%), feldspars (avg.= 1%; max.= 2%), stevensite ooids and peloids (avg.= 3.7% max.= 9%), and undetermined grains (avg.= 2.2%; max.= 4%). Microcrystalline calcite replaces intergranular authigenic smectite (max.= 4%), stevensite peloids and ooids (avg.= 1.5%; max.= 4.3%). Macrocrystalline calcite also fills fractures, intraparticle and moldic pores formed by dissolution of bivalve bioclasts. In the siliciclastic mudstones and very fine-grained, muddy sandstones, macrocrystalline calcite (avg.= 15.7%; max.= 57.3%) and microcrystalline calcite (avg.= 29.6%) replace the mud matrix, and form rounded and elongated concretions that replace matrix, stevensite peloids (avg.= 1.2%; max.= 4%), feldspars (max.= 1.3%) and undetermined grains (avg.= 2.3%; max.= 5%).

In hybrid rocks, macrocrystalline calcite grain replacement and cementation is rather intense, dominantly filling intergranular pores (avg.= 8.7%; max.= 29.3%), replacing intergranular smectite (avg.= 1.1%; max.= 8%) and zeolites (max.= 1%). The most commonly replaced grains are stevensite peloids and ooids (avg.= 9.2%; max.= 29.6%). Intraparticle pores generated by dissolution of bivalve bioclasts are filled by drusiform to macroscopic calcite (avg.= 1.7%; max.= 9.3%). Microcrystalline calcite fills intergranular pores (avg.= 2.6%; max.= 22.3%), and replaces stevensitic grains (avg.= 1.8%; max.= 11.3%).

Macrocrystalline calcite cementation is the main process of reduction of primary interparticle porosity in the stevensitic arenites (avg.= 8.1%; max.= 27%), also filling pores from the shrinkage and dissolution of stevensitic grains. Calcite replaces stevensite ooids and peloids (avg.= 14.2%; max.= 29%), as well as interparticle microcrystalline quartz (avg.= 1.2%; max.= 9.7%).

### 5.2. Dolomite

Dolomite occurs only in the stevensitic and hybrid arenites, and in the siliciclastic mudstones. In the stevensitic rocks, small dolomite rhombohedra fill interparticle pores (avg.= 6.1%; max.= 19%), replace ooids and peloids (avg.= 15.3%; max.= 30%), and intraclasts (max.= 5.3%), as well as feldspars (avg. <1%; max.= 3%) and other grains. Microcrystalline dolomite occurs locally.

In hybrid rocks, small dolomite rhombohedra occurs only in two of the 17 analyzed samples, filling intergranular pores (max.5%), replacing stevensitic grains (max.= 6.7%) and laminations (max.= 8.3%), as well as bioclasts and other grains. In the siliciclastic mudstones, small dolomite rhombohedra replace the mud matrix (max.= 43.3%) and grains (max.= 2.3%).

### 5.3. Smectite

Smectite is the main authigenic clay mineral present in the analyzed rocks, occurring in the volcaniclastic sandstones and conglomerates and in hybrid rocks. In the sandstones and conglomerates, smectite occurs mostly as rims (avg.= 9.9%; max.= 18%) and coatings, covering continuous- to discontinuously the grains. In some samples, thick rims totally fill the intergranular pores. The coatings are commonly covered by the rims. Chaotic microcrystalline or sheaf aggregates fill partial- to totally intergranular pores (avg.= 1.6%; max.= 4.6%). Volcanic rock fragments are intensely replaced by microcrystalline smectite (avg.= 9.7%; max.= 16.7%). The replacement of feldspars, mostly of plagioclase in plutonic fragments (avg.= 2.4%; max.= 5.7%), of amphibole, biotite (with expansion) and other grains, is common.

Smectite habits and replacement processes observed in the hybrid arenites are similar to those in the volcaniclastic sandstones. Volcanic fragments are also the main constituents replaced by microcrystalline smectite (avg.= 2.2%; max.= 5.6%), being the feldspars, amphiboles, biotite and other grains much less replaced. Coatings and rims cover continuous- to discontinuously the primary constituents (avg.= 2.4%; max.= 11.6%). Microcrystalline and sheaf aggregates fill partially to totally the intergranular pores (avg.= 1.7 %; max.= 10%), as well as part of the pores generated by dissolution of feldspars, bivalve bioclasts (max.= 2.3%) and volcanic fragments (max.= 2.7 %).

## 5.4. Zeolites

In the volcaniclastic sandstones and conglomerates, zeolites occur as intergranular cement with prismatic (avg.= 4.2%; max.= 16.3%), and rarely radiated habit. Zeolites also replace the volcanic fragments (avg.= 2.6%; max.= 8.7%), feldspars (max.= 3.3%) and other grains (avg.= 1.7%; max.= 7%). Discrete intergranular crystals of analcime and prismatic zeolites are less common, being often replaced by calcite.

In the hybrid arenites, zeolites occur with prismatic and lamellar habits, filling intergranular pores (max.= 10%), and rarely intraparticle pores in bivalve bioclasts. Zeolites replace volcanic fragments (max.= 2.7%), stevensitic grains (max.= 1,7%), and feldspars (max.= 2.3%), as well as smectite that rimmed intergranular pores and replaced volcanic grains and feldspars. In muddy siliciclastic rocks, zeolites occur in only one sample, replacing feldspar grains.

## 5.5. Other diagenetic constituents

In the volcaniclastic sandstones and conglomerates, microcrystalline **pyrite** commonly replaces the volcanic fragments (avg.= 1.4%; max.= 4%), and rarely other grains. Siliciclastic mudrocks and fine-grained sandstones commonly show microcrystalline pyrite replacing the mud matrix (avg.= 5.1%; max.= 22%), stevensite peloids, biotite and other grains (avg.= 2.3%; max.= 8.3%). In the hybrid rocks, microcrystalline pyrite replaces volcanic fragments (max.= 3.3%) and stevensite peloids (max.= 6,3%), besides bivalve bioclasts, feldspars, biotite and mud matrix. Coarse, blocky pyrite replace grains in some samples (max.= 3.7%). In stevensitic rocks, microcrystalline pyrite replaces ooids and peloids (max.= 2.7%), and seldom stevensite laminations.

Diagenetic **iron and titanium oxides** occur in small amounts in the analyzed rocks, always with microcrystalline habit. Part of the volcanic rock fragments replaced by hematite (max.= 2.7%) may have come already oxidized from the source areas.

Some volcaniclastic sandstones display thin hematite coatings (max.= 1.3%), which are covered by pore-lining smectites. Microcrystalline aggregates and discrete crystals of titanium oxides replace volcanic fragments (max.= 1.7%) and heavy mineral grains in the volcaniclastic sandstones and conglomerates. Titanium oxides are very scarce in the fine-grained siliciclastic deposits, where they occur mostly replacing biotite. In hybrid arenites, microcrystalline hematite replaces volcanic fragments (max.= 5.3%) and other grains (max.= 2%). Titanium oxides are scarce in these rocks, replacing volcanic fragments.

Authigenic **silica** occurs significantly in the stevensitic rocks only. In the stevensitic arenites, quartz fills interparticle pores with drusiform (avg.= 2.8%; max.= 12.7 %), microcrystalline (avg.= 1.9%; max.= 9.6%), and coarse mosaic habit (7.7 %). Stevensite ooids and peloids are replaced by microcrystalline (avg.= 4.1%; max.= 18.7%) and mosaic habit (max.= 2,6%). Large stevensitic intraclasts are replaced by microcrystalline quartz (avg.= 1%; max.= 4%). Intraclasts shrinkage pores are filled by microcrystalline and mosaic quartz (avg.= 1.2%; max.= 5.6%). In the hybrid arenites, radiated chalcedony locally replaces bivalve and ostracod bioclasts, and is replaced by calcite or dolomite. Quartz overgrowths and discrete crystals are scarce both in hybrid arenites and in volcaniclastic sandstones.

**Potassium feldspar** overgrowths cover microcline grains discontinuously, and discrete prismatic crystals partially fill intergranular and intragranular pores in volcaniclastic sandstones and hybrid arenites (max.= 2.6%). Less commonly, authigenic K-feldspar replaces grain-replacive zeolites in microcline and stevensitic grains in hybrid arenites. The **albitization** of detrital feldspars is limited to some of the hybrid arenites and volcaniclastic sandstones only, affecting dominantly plagioclase grains (max.= 1.7%).

Lamellar **kaolinite** occurs scarcely in hybrid rocks, replacing and expanding biotite and muscovite (max.= 1.7%). Booklet aggregates fill intergranular pores (max.= 2.6%) and rarely replace feldspars. In siliciclastic mudstones, lamellar kaolinite replaces and expands muscovite (med.= 1.3%; max.= 8%). Calcite and dolomite replace kaolinite in both rock types. **Sulfates** are rare in the studied rocks. Fibrous gypsum fills fractures in fine-grained rocks. Macrocrystalline barite replaces carbonate and phosphate bioclasts, and fills intergranular pores. These sulfates are commonly replaced by calcite, dolomite and pyrite. Present only in hybrid arenites, **pseudomatrix** was generated by the compaction of mud intraclasts, and stevensite peloids and ooids.

## 6. Magmatic and hydrothermal processes and products

Basic magmatic rocks occur interbedded with or intruded at the base of the rift section of the Lagoa Feia Group. These rocks include basalts, diabases (dolerites), hydrothermal breccias and hydromagmatic types. The basalts are dominantly hemicrystalline with intersertal texture and amygdaloidal structure. Their alteration is commonly intense, with glass, pyroxenes and plagioclases replaced by smectites or by celadonite, which also fill the amygdales (Figure 8A), together with Fe, Ti and Mn oxides, marcasite, calcite, zeolites and dolomite. The textural and compositional aspects of the basaltic rocks interbedded at the base of the Lagoa Feia Group are similar to those of the underlying Cabiúnas Formation (Winter et al., 2007), which correspond to subalkaline to tholeiitic types with affinity to the Paraná Basin floods

(Mizusaki et al., 1988; 1992). Diabases (dolerites) with composition equivalent to the basalts occur intruded locally in the sedimentary rocks of the Lagoa Feia Group. Their intrusive character is indicated by the scarcity of glass (including holocrystalline types) and of amygdales, and by the pattern of plagioclase-pyroxene intergrowth (Figure 8B).

The occurrence of hydromagmatic types records the interaction between magmatic extrusive and shallow intrusive activity and the rift sediments. The interaction between magma and sediments is expressed as chaotic mixtures of hybrid sediments and very angular, vitreous volcanic fragments (Figure 8C), promoted by the explosive vaporization of the phreatic fluids by the hot magma (pepperites).

Hydrothermal volcanic breccias are common in the analyzed section. These rocks correspond to strongly fractured basalts and diabases, with filling of the fractures by calcite, silica, zeolites, celadonite and scapolite, and strong alteration of the rocks, mostly to smectites, calcite, celadonite and zeolites (Figure 8D).

Alteration of the sedimentary rocks by hydrothermal fluids circulating through fractures was also observed in several cases. Some samples contain spherulitic or fascicular calcite aggregates characteristic of abiotic travertines (Chafetz and Guidry, 1999), such as those extensively observed in the sag “pre-salt” reservoirs of the Santos, Campos e Espírito Santo basins (Wright, 2012; Figure 8E). The main observed process of hydrothermal alteration of sedimentary materials is, however, the intense silicification observed in some samples in the form of pervasive replacement by microcrystalline to macrocrystalline quartz (Figure 8F).

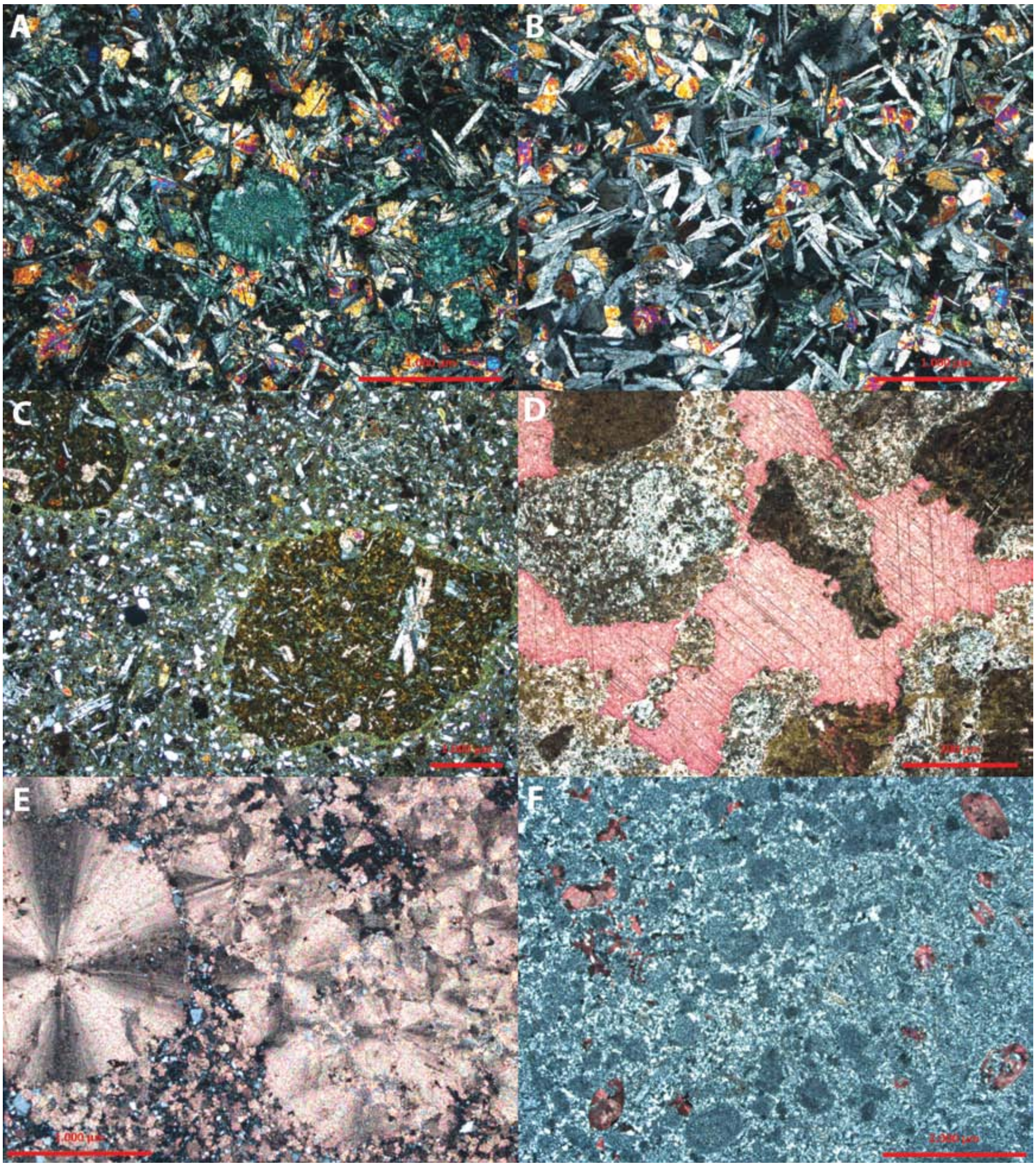

Figure 8. **A**) Celadonite-filled vesicles in basalt. RJS36 3475,6. Crossed polarizers (XP). **B**) Plagioclase and pyroxene intergrowth in diabase. RJS36 3473,1. XP. **C**) Large, irregular-shaped vitreous volcanic fragments, chaotically mixed with hybrid sediments (probable peperite). RJS485 3096,8. XP. **D**) Intense alteration of volcanic fragments in a calcite-cemented breccia. RJS379 3113,05. Plane polarizers (//P). **E**) Spherulitic aggregates replacing hydrothermally-altered sediment. RJS165 2544,65. **F**) Intense silicification of stevensitic particles and ostracode bioclasts, probably related to the percolation of hydrothermal fluids. PM04RJS 2932,95. XP.

# 7. Discussion

## 7.1. Aspects of provenance and depositional environments

The pioneer work of Schaller (1973) interpreted the depositional setting of the Lagoa Feia Group (then termed Formation) as alluvial-lacustrine and subdivided the succession in three intervals. The lower interval would be constituted by sandstones and conglomerates intercalated with mudrock lenses, and separated from the crystalline basement by basaltic rocks. The intermediate interval would be formed by intercalations of carbonates, shales, anhydrite and halite, and the upper interval would correspond to shales, calcilutites, siltstones and sandstones intercalated with anhydrite.

Bertani and Carozzi (1985a; 1985b) developed a specific study of the succession, interpreting the depositional environments of the Lagoa Feia Group as an alternation between alkaline and saline lacustrine (playa lakes), with deposits rich in ostracods formed during arid periods, and fresh water lacustrine (pluvial lakes), with sediments rich in bivalves deposited during wetter periods. They defined four main sequences: terrigenous, ostracod, bivalve and volcaniclastic, recognizing 14 microfacies within these sequences. According to them, the volcanic activity that took place during the early infill of the basin would have influenced the chemical composition of the lakes, promoting the formation of kerolite ooids, commonly nucleated on basaltic glass particles.

In the work by Dias et al*.,* (1988), the Lagoa Feia Group is divided into 4 depositional sequences: clastic, stevensitic, carbonatic ("coquinas") and clastic-evaporitic, formed in lacustrine environments. The clastic sequence would be composed at the base by volcaniclastic sandstones and conglomerates, and by alluvial-fan sandstones with carbonate bioclasts and stevensite ooids at the top. The talc-stevensitic sequence is characterized by a marginal lacustrine facies of siltstones and arenites constituted by talc-stevensite ooids and peloids. The “coquinas” sequence would correspond to bioclastic deposits of bivalves and ostracods. They interpreted most of these deposits as bioaccumulations, what is incongruent with the definition of coquina as an extensively reworked and abraded bioclastic deposit (Scholle and Ulmer-Scholle, 2003). The clastic-evaporitic sequence is constituted by mostly alluvial fan, volcaniclastic sandstones and conglomerates at the base, and by evaporites at the top, deposited after a basin-wide marine incursion. Stable isotopic analyses of the organic matter contained in shales intercalated with this unit likewise indicate an increasing salinity during its deposition.

Abrahão and Warme (1990) reinterpreted the deposition of the Lagoa Feia Group as occurred in several environments and sub-environments in diverse lakes within a basin in active rifting process. The beginning of sedimentation would correspond to volcaniclastic sands and gravels deposited by alluvial fans along the edges of uplifted basement blocks covered by the basaltic rocks of the Cabiúnas Formation. Such deposits occur both at the beginning of the basin infill and at the end of the rift and beginning of the sag phase, when they are covered by the evaporites. The presence of bivalve bioclasts in the conglomerates would indicate subaqueous fan-deltaic deposition. Fine-grained siliciclastic facies would correspond to mud plain, lake margin deposits. The rare occurrences of mud cracks, nodular anhydrite and molds of halite and other evaporitic minerals would indicate periods of more severe aridity during this phase. Conversely, dark, organic-rich lacustrine shales and turbiditic sandstones would have been deposited with the deepening and interconnection of the lakes. Banks of bivalve bioclasts would have been deposited during periods of shallowing of the lakes, along the edges of the faulted and tilted structural blocks. According to Abrahão and Warme (1990), the stevensite deposits would have been formed as result of the volcanic activity close and/or contemporaneous to the sedimentation.

The present integrated study, which incorporated seismic analysis, detailed core description and systematic petrographyduring the development of a larger project, revealed, however, that the conditions of deposition for most of the rift section of the Lagoa Feia Group deviate substantially from the depositional models proposed in the above summarized previous works.

Most of the described volcaniclastic sandstones and conglomerates are massive, or with rare, irregular stratification. The primary composition of the conglomerates corresponds to volcanic litharenites in Folk (1968) classification (Figure 9A), and to an undissected arc provenance, according to Dickinson (1985; Figure 9B). The volcaniclastic sandstones are lithic arkoses and feldspathic litharenites *sensu* Folk (1968, Figure 9A), with a transitional magmatic arc provenance according to Dickinson (1985; Figure 9B). Although volcanic fragments occur disseminated throughout most of the different types of deposits analyzed, the volcaniclastic conglomerates and sandstones are strongly concentrated in the lows along the half graben border faults, and their massive structure suggest deposition by hyperconcentrated, gravitational flows. The presence of bivalves is rather scarce in these lithotypes, being limited to rare, broken bioclasts in the volcaniclastic sandstones. Conversely, stevensite ooids and peloids occur commonly in volcaniclastic conglomerates and sandstones.

The usual mixture of rounded volcanic fragments with angular quartz and feldspars grains and plutonic fragments in these conglomerates and sandstones is indicative of the recycling of epiclastic deposits of the Cabiúnas Formation, mixed with first-cycle sediments eroded from the faulted edges of uplifted basement blocks. This in line with the stratigraphic interpretation of Rangel and Carminatti (2000), and Winter et al., (2007), of a regional unconformity between the Cabiúnas Formation, which would correspond to an earliest phase of infill of shallow and wide proto-rift depressions, and the syn-rift Lagoa Feia Group.

On the other hand, the presence of pepperites containing particles typical of the Lagoa Feia lacustrine deposits, such as bivalve bioclasts and stevensite ooids, indicates that the magmatic activity persisted locally during the active rift phase, what was not reported in previous works on the basin.

Ooids, peloids and laminations of magnesian phyllosilicates (stevensite, talc-stevensite or kerolite; Bertani and Carozzi, 1985a; 1985b; Rehim et al., 1986) occur conspicuously in the rift and the sag sections of the Lagoa Feia Group. Conditions for the syngenetic formation of these layer silicates are considered to involve high activity of Mg and Si and high pH (above 9). According to Tettenhorst and Moore (1978), and Noack (1989), such conditions are likely to be promoted by hydrothermal activity associated to magmatism in lacustrine settings. Likewise, Jones and Weir (1983), Darragi and Tardy (1987), and Cerling (1996) identified the formation of stevensite in African alkaline lakes with small supply of detrital mud as a product of very high pH and Mg activity related to hydrothermal and/or magmatic activity.

Pozo and Casas (1999) interpreted the formation of kerolite and stevensite in a lacustrine deposit from the Tertiary of Spain as a product of mixing between the lake water with groundwater with high Si activity. This would promote an increase in Mg/H ratio and the formation of a Si and Mg gel, which would later crystallize as kerolite or stevensite. Likewise, Rehim et al., (1986) interpreted that the generation of stevensite and mixed-layer talc-stevensite in the Lagoa Feia Group would be a product of colloidal precipitation of a Mg and Si gel in alkaline lacustrine environments. According to them, brackish waters rich in Si and Mg would precipitate stevensite under pH ranging between 8 and 9, and talc under pH higher than 9.

Jones (1996) explored experimentally the control of pH, Mg and Si activities over the mineralogy of syngenetic magnesian phyllosilicates. He suggested that more dilute lacustrine or palustrine environments would favor the precipitation of sepiolite. The increase of pH and Mg/Si ratio would favor the precipitation of kerolite. Further increase of alkalinity and of salinity during arid periods would favor the formation of stevensite.

Tosca and Masterson (in press) performed experiments to constrain the precipitation conditions of magnesian phyllosilicates in relation to pH, Mg/Si ratio and salinity. According to them, kerolite is precipitated in environments with low salinity, high pH and high Mg/Si ratio. Sepiolite is precipitated in environments with lower pH and low Mg/Si ratio, while stevensite is precipitated in environments with high pH, high salinity, and high Mg/Si ratio. pH seems to exert the main control on the mineralogy of magnesian phyllosilicates, with 2:1 structures (stevensite or kerolite) favored in pH above 9, and sepiolite or similar structures (palygorskite, atapulgite) being precipitated at pH below 9 together with amorphous silica.

The examination of the occurrence of stevensite and talc-stevensite or kerolite in the Lagoa Feia Group (Bertani and Carozzi, 1985a; 1985b; Rehim et al., 1986) at the light of several other occurrences of 2:1 magnesian phyllosilicates (Bradley and Fahey, 1962; Dyni, 1976; Tettenhorst and Moore, 1978; Eberl et al., 1982; Khoury et al., 1982; Jones and Weir, 1983; Jones, 1986; Darragi and Tardy, 1987; Martin de Vidales et al., 1991; Buch and Rose, 1996; Cerling, 1996; Chahi et al., 1997, 1999; Hover et al., 1999; Pozo and Casas, 1999; Mayayo et al., 2000; Yuretich and Ervin, 2002; Cuevas et al., 2003; Hover and Ashley, 2003; Furquin et al., 2008) indicates that these minerals are essentially formed by syngenetic precipitation in alkaline lacustrine and palustrine environments. Our interpretation is that the morphology of the precipitates is controlled by the energy of the depositional environment, as observed for glauconite, berthierine and other syngenetic clay minerals (Van Houten and Purucker, 1984; Odin, 1988). As for those clays, the stevensitic ooids are likely to be formed in environments more agitated by waves or currents, while the laminations are certainly deposited in low-energy environment. The peloids are probably formed in environments of energy level between that for the formation of the ooids and that for the laminations. The common deformation of the ooids, visible even in areas with very early cementation by silica or carbonates, indicates the very soft original aspect of these grains, probably formed by the colloidal precipitation of a Mg-Si gel (Pozo and Casas, 1999). The stevensitic conglomerates are formed either by intraclasts eroded from laminated or from ooidal-peloidal deposits, or by pisoids, probably formed in higher energy depositional conditions equivalent to those of the ooids. These environmental interpretations apply for the stevensitic arenites and conglomerates, which were formed by the *in situ* precipitation of magnesian phyllosilicates.

However, the widespread distribution of stevensitic ooids and peloids in most of the lithotypes described in the Lagoa Feia Group, including the volcaniclastic sandstones and the bioclastic rudstones and grainstones, indicates the occurrence of substantial reworking of stevensitic deposits and mixing with other sediment types within the rift grabens. The mixing of stevensitic ooids and peloids and bivalve bioclasts in practically every sample described of the rudstones that constitute the main Lagoa Feia reservoirs is particularly indicative of this extensive re-sedimentation. As stevensite is characteristically precipitated at pH higher than 9 and bivalves cannot effectively multiply at pH higher than 8 (Locke, 2008) the mixing of these incompatible sediments must have occurred after their primary generation in different shallow lacustrine environments. The frequently massive structure of the rudstones, as well as of the volcaniclastic sandstones and conglomerates and of the hybrid arenites, all containing stevensitic peloids and ooids, suggest that this mixing was promoted by gravitational flows, which re-deposited different shallow lacustrine and alluvial sediments into deeper lacustrine settings. The hybrid arenites, constituted by mixtures in variable proportions of siliciclastic grains, volcanic fragments, bivalve and ostracod bioclasts, and stevensitic ooids and peloids, are the ultimate product of this re-sedimentation (Figure 4).

The complex distribution of volcaniclastic, siliciclastic, stevensitic and carbonate sediments within the rift section of the Lagoa Feia Group is probably related to the dynamic faulting and tilting of the multiple structural blocks that constituted, in different periods, the basement and the sedimentary source areas. The intense tectonism occurring during rifting was the most probable mechanism for the redistribution of sediments from shallow areas located along the flexural margins of faulted blocks and relay ramps between grabens to adjacent deeper settings, or even from one half graben to another.

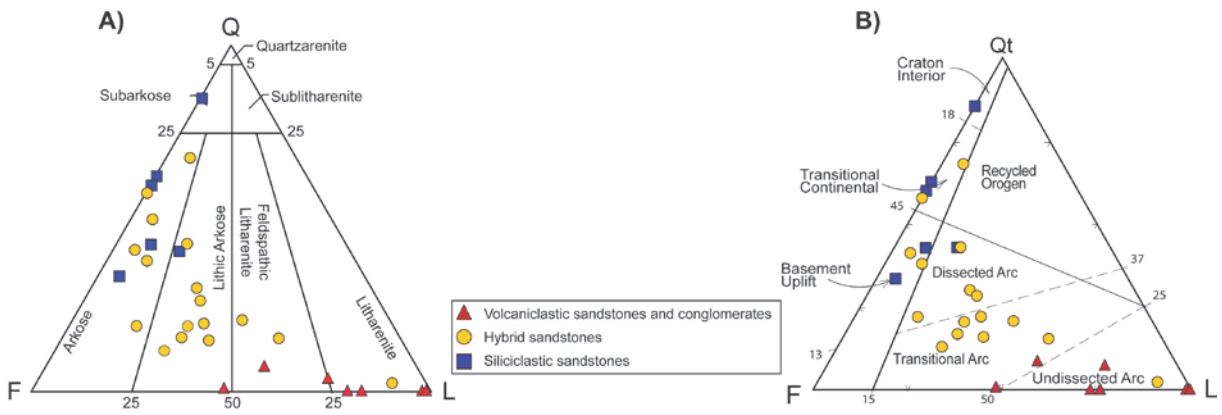


Figure 9. A) Folk (1968) classification diagram, and B) Dickinson (1985) provenance diagram with the analyzed samples.

## 7.2. Diagenetic patterns

### 7.2.1. Clastic rocks

The very immature and unstable primary composition of clastic rocks, rich in feldspars, micas and heavy minerals grains and in volcanic fragments, caused intense diagenetic alterations. Most of these promoted significant porosity reduction. However, some samples show fair remaining intergranular porosity, besides secondary intragranular porosity, which are commonly filled by oil or residual bitumen. The synthetic sequence of the main diagenetic processes occurring in the analyzed clastic rocks can be summarized as follows (the main processes or constituents in bold).

1) **Dissolution** of volcanic fragments, particularly of glass in hemicrystalline fragments, and of unstable heavy minerals, such as pyroxenes; less intensely of feldspars (Figure 10A).
2) Kaolinite pseudomorphically replacing and expanding muscovite (almost only in fine-grained and muddy facies); rare replacement of feldspars by microcrystalline kaolinite.
3) Heterogeneous, shallow tectonic fracturing.
4) Fibrous gypsum filling rock fractures.
5) Replacement of volcanic fragments, biotite and siliciclastic matrix by microcrystalline pyrite (Figure 10B).
6) Replacement of volcanic fragments, biotite and heavy minerals by iron and titanium oxides; hematite coatings on grains (Figure 10C).
7) **Smectite** as pore-lining coatings, rims, locally microcrystalline pore-filling aggregates (Figure 10D); replacing volcanic rock fragments, unstable heavy minerals, feldspars and biotite (Figure 11A).
8) Prismatic and blocky **zeolites** encircling and filling intergranular pores (Figure 11B-C); locally replacing smectite, volcanic fragments and feldspars (Figure 11D).
9) Discontinuous **quartz and K-feldspar overgrowths.** Prismatic K-feldspar replacing feldspar grains and filling intergranular pores (Figure 12A-C).
10) Replacement of matrix by radial chalcedony in mudrocks.

**11)** Macrocrystalline and locally poikilotopic **calcite** filling intergranular pores and replacing feldspars and heavy mineral grains, volcanic fragments, smectite and zeolites (Figure 12D).

**12)** Blocky to macrocrystalline **dolomite** filling intergranular pores and replacing siliciclastic mud matrix, feldspars, micas and heavy mineral grains and volcanic fragments.

**13)** Local dissolution of intragranular and intergranular calcite and zeolites, and of fracture-filling gypsum.

**14)** Deformation of ductile smectite-altered grains, micas and rare mud intraclasts. Limited quartz and feldspars grains alteration.

**15)** Macrocrystalline, grain-replacive barite.

**16)** Coarse mosaic calcite replacing chalcedony and barite.

**17)** Blocky pyrite replacing barite and volcanic fragments.

**18)** Limited feldspars albitization (Figure 12F).

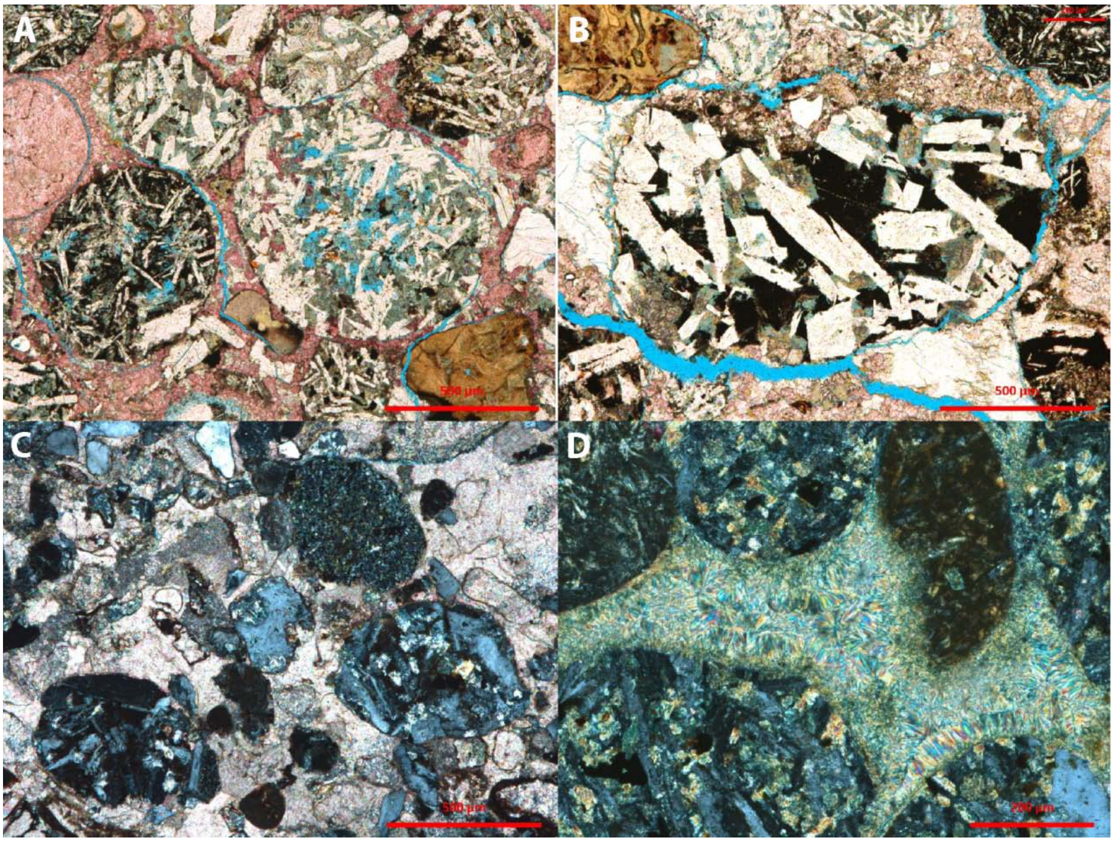


Figure 10. **A**) Intragranular porosity in hemicrystalline volcanic fragments. RJS36 2973,55. Plane polarizers (//P). **B**) Microcrystalline pyrite replacing volcanic fragments cemented by calcite and zeolite. RJS36 2973,55. //P. **C**) Iron oxide coatings surrounding continuously the grains of a clastic sandstone. RJS165 2425,10. Crossed polarizers (XP). **D**) Smectite as thick, pore-filling rims and microcrystalline aggregates replacing lithic grains. RJS486A 2873,65. XP.

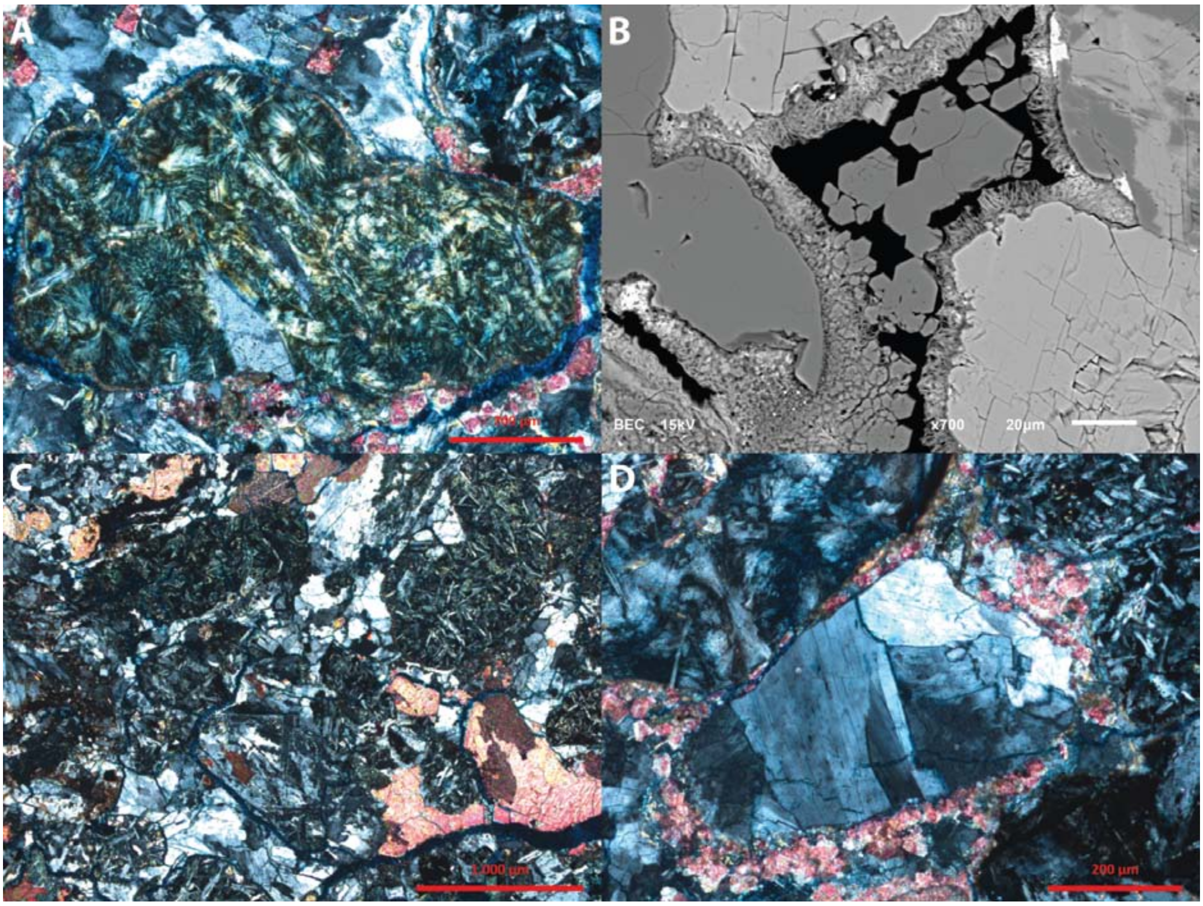


Figure 11. **A**) Volcanic fragments replaced by radial smectite aggregates and cemented by zeolite and calcite. RJS36 2973,5. Crossed polarizers (XP). **B**) Discrete zeolite crystals in primary intergranular pores. Smectite rims covering the grains. RJS108 3567,9. BSE. **C**). Coarse zeolite and calcite cementing and replacing volcanic fragments. RJS36 2969,2. XP. **D**) Grain replaced by coarse prismatic zeolite. Macrocrystalline calcite replacing intergranular zeolite. RJS36 2973,55. XP.

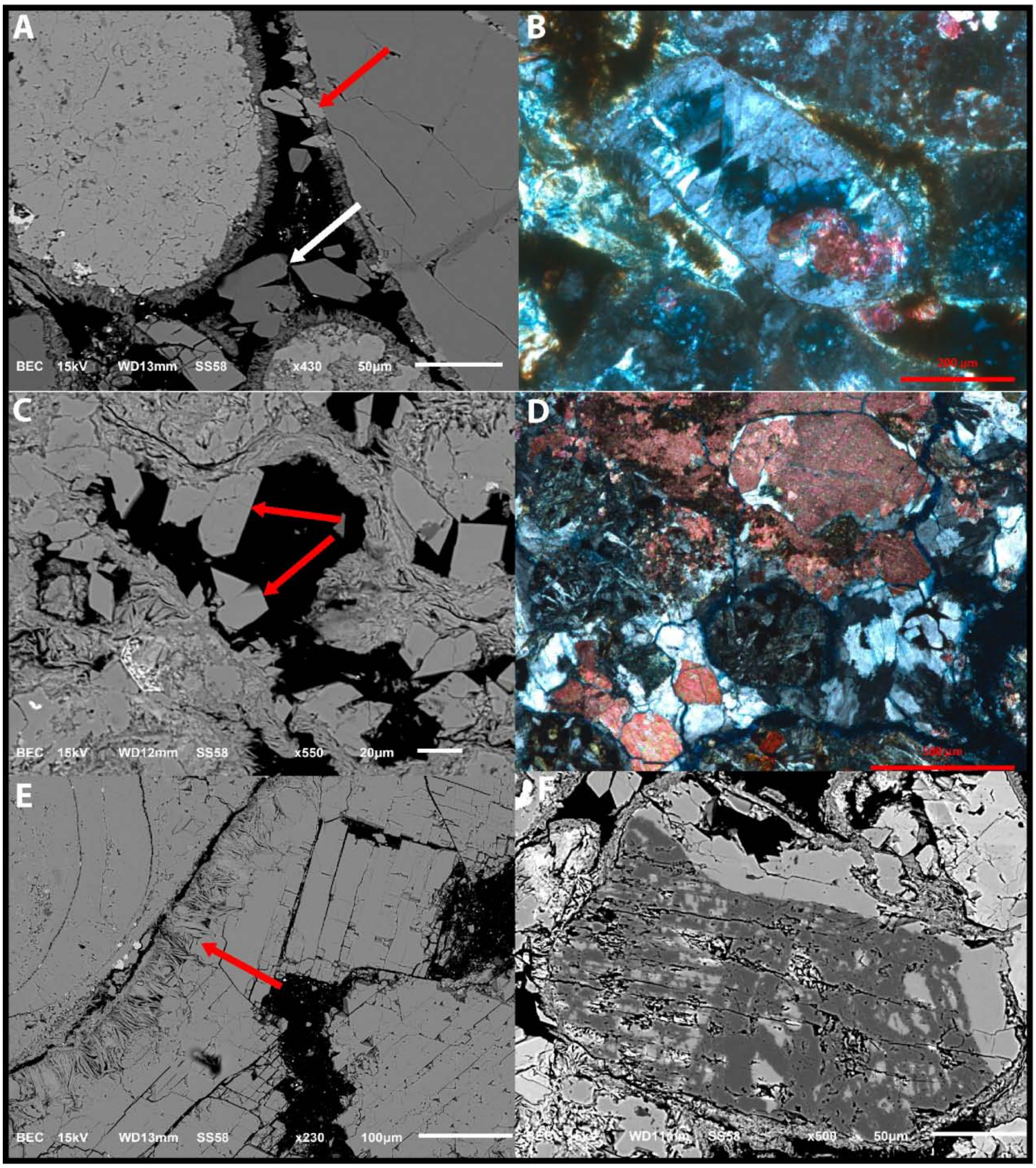


Figure 12. **A**) Discrete zeolite crystals (white arrow) and feldspar. K-feldspar overgrowths engolfing smectite rims (red arrow). RJS108 3567,9. BSE. **B**) Partially dissolved feldspar grain with K-feldspar ingrowths and outgrowths. RJS113 2918,00. Crossed polarizers (XP). **C**) Discrete K-feldspar crystals (arrows) in moldic pore with smectite rims. RJS113 2918,00. BSE. **D** ) Macrocrystalline calcite replacing grains and cementing intergranular pores with prismatic zeolite. RJS36 2969,2. XP. **E**) Calcite replacing smectite rims (arrow). RJS36 2965,6. BSE. **F**) Albitized plagioclase grains. RJS113 3567,9. BSE.

A representation of the diagenetic sequence characteristic of the volcaniclastic rocks is showed in Figure 13.

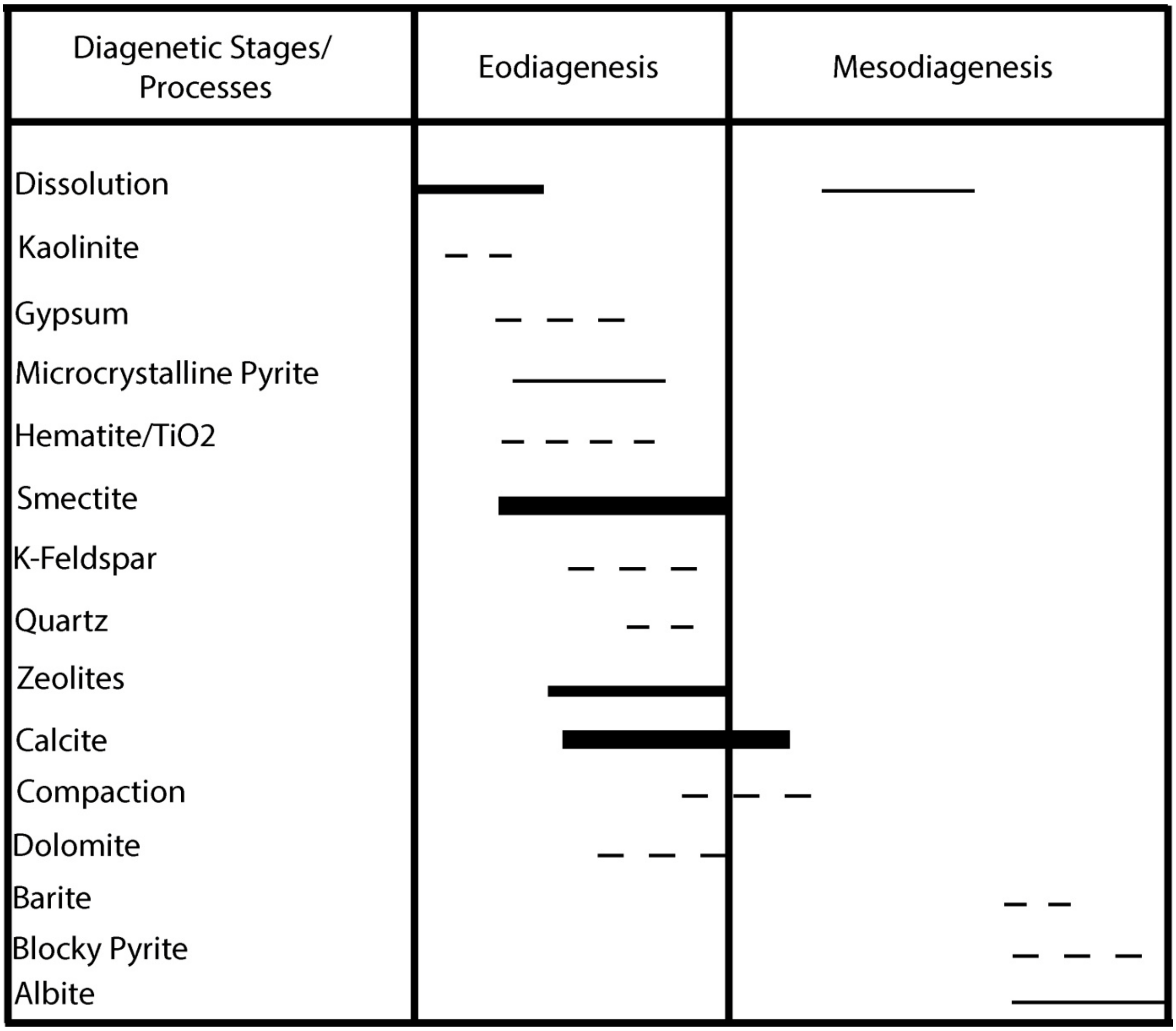


Figure 13. Diagenetic sequence for volcaniclastic sandstones/conglomerates.

### 7.2.2. Stevensitic rocks

The diagenesis of stevensitic rocks is conditioned by the physical and chemical instability of the stevensite ooids and peloids. Where not cemented during early diagenesis, these particles were strongly compacted, resulting in total porosity loss. In areas cemented by eodiagenetic silica or carbonates, the stevensitic grains were also partially replaced by these minerals, and by zeolites. The only porosity present in these rocks is secondary, generated by the dissolution of calcite cement, and mostly of the stevensitic grains. The synthetic sequence of the main diagenetic processes occurring in the analyzed stevensitic rocks can be summarized as follows (the main processes or constituents in bold).

**1)** Microcrystalline **quartz** replacing stevensite ooids and peloids; drusiform quartz filling interparticle pores in some samples (Figure 14A).

**2)** Microcrystalline and dominantly macrocrystalline **calcite** filling interparticle pores, replacing ooids, peloids and silica cement (Figure 14B).

**3)** Blocky to macrocrystalline dolomite, rarely microcrystalline or as rims, filling or lining interparticle pores, or more commonly replacing stevensitic grains and laminations, and silica cement (Figure 14C).

**4)** Strong **compaction** and deformation of ooids and peloids in areas devoid of early cementation (Figure 14D).

**5)** Dissolution of ooids and peloids promoting the formation of intraparticle microporosity and macroporosity and moldic porosity (Figure 14E-F).

**6)** Dissolution of calcite cement.

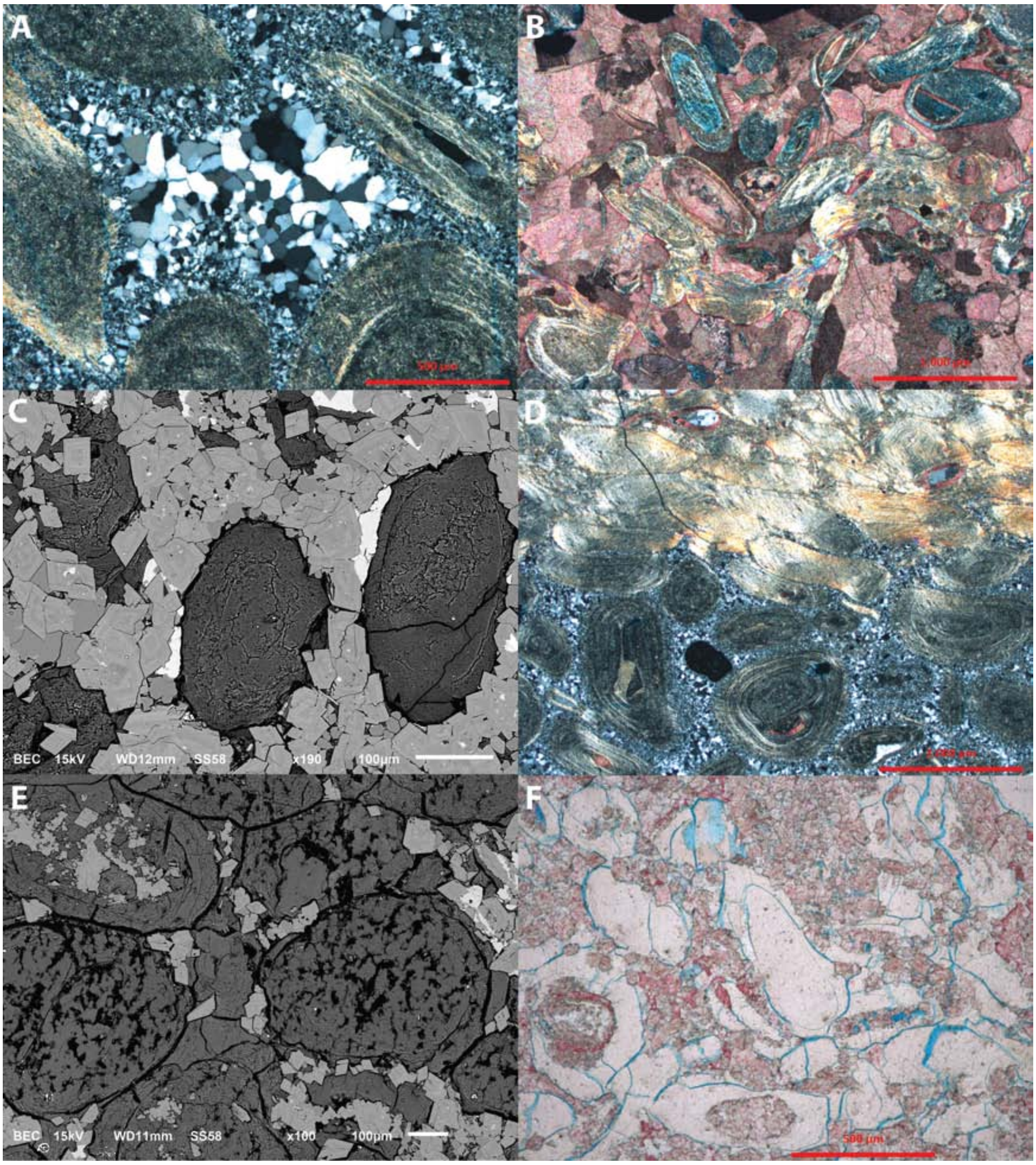


Figure 14. **A**) Drusiform quartz cementing stevensite ooids. RJS514 4163,70. Crossed polarizers (XP). **B**) Partially-deformed stevensite ooids replaced by macrocrystalline calcite. RJS514 4165,75 XP. **C**) Stevensite ooids partially replaced and cemented by blocky dolomite. RJS514 4213,95. BSE. **D**) Contrast between an area where stevensite ooids have been cemented by early silica and not compacted, and an area without cementation, with intense compaction. RJS514 4163,70. XP. **E)**. Partially dissolved stevensite ooids. RJS514 4213,95. BSE. **F)** Dissolved, contracted and fractured stevensite ooids, partially replaced by calcite. RJS514 4013,95. Plane polarizers (//P).

A representation of the diagenetic sequence characteristic of the stevensitic rocks is showed in Figure 15.

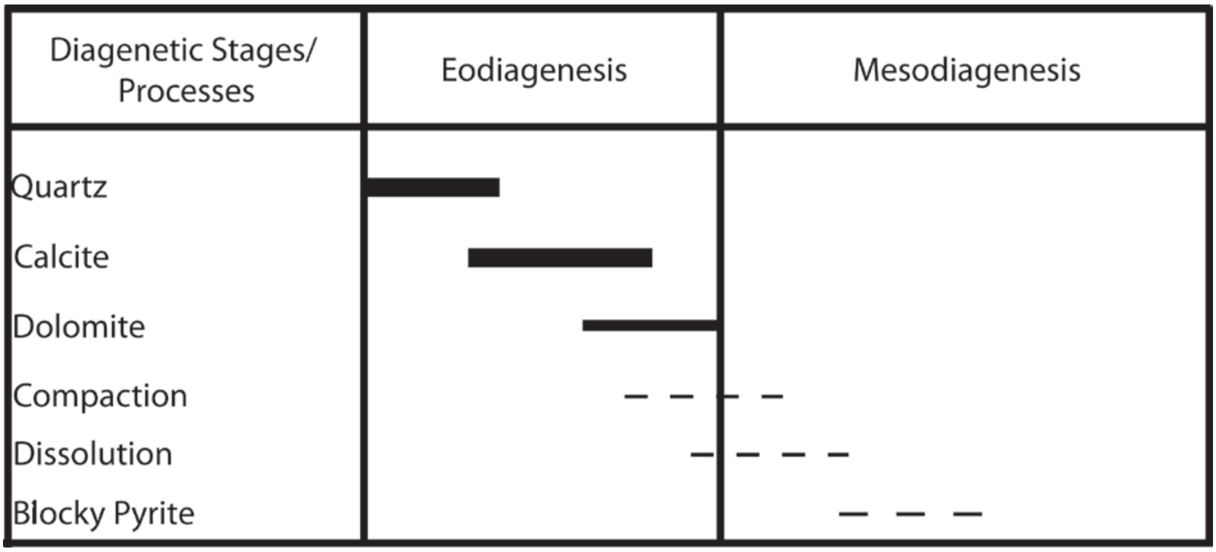


Figure 15. Diagenetic sequence for stevensitic arenites.

#### 7.2.3. **Hybrid rocks**

The synthetic sequence of the main diagenetic processes occurring in the analyzed hybrid rocks can be summarized as follows (the main processes or constituents in bold).

1) **Dissolution** of volcanic fragments, particularly of glass in hemicrystalline fragments, and of unstable heavy minerals, such as pyroxenes; less intensely of feldspars and bioclasts.
2) Microcrystalline kaolinite replacing feldspars, muscovite and stevensitic grains.
3) Thin hematite and smectite coatings non-selectively covering grains (Figure 16A).
4) **Smectite** replacing volcanic rock fragments, unstable heavy minerals, feldspars and biotite (Figure 16B), filling intergranular pores, and locally as rims covering the coatings.
5) **K-feldspar** overgrowths on microcline grains. Discrete prismatic K-feldspar and quartz in intergranular pores (Figure 16C).

6) Prismatic or microcrystalline zeolites replacing volcanic fragments, feldspars, stevensitic ooids and peloids, and smectite, and locally filling intergranular pores (Figure 16D-E).

7) Microcrystalline pyrite replacing volcanic fragments (Figure 16F).

8) Commonly abundant intergranular pore-filling macrocrystalline or blocky calcite. Heterogeneous replacement of stevensite ooids and peloids by microcrystalline or macrocrystalline calcite (Figure 17A-B).

9) Recrystallization of bivalve bioclasts to coarse calcite mosaic (Figure 17C).

10) Fibrous gypsum and macrocrystalline calcite filling sub-horizontal fractures.

11) Small dolomite rhombohedra rimming grains. Blocky to macrocrystalline dolomite filling intergranular pores and replacing stevensitic grains and feldspars.

12) Microcrystalline quartz and chalcedony replacing stevensitic and carbonate grains. Drusiform quartz filling intergranular pores. Rare chalcedony rims.

13) Mechanical compaction through deformation of stevensitic grains, and fracturing of bioclasts and feldspars (Figure 17D).

14) Chemical compaction through pressure dissolution along intergranular contacts and stylolites.

15) Dissolution of feldspars and stevensitic grains, generating intragranular and moldic pores, and of intergranular calcite (Figure 17E).

16) Macrocrystalline barite replacing grains and filling intergranular pores.

17) Blocky, coarse pyrite replacing grains (Figure 17F).

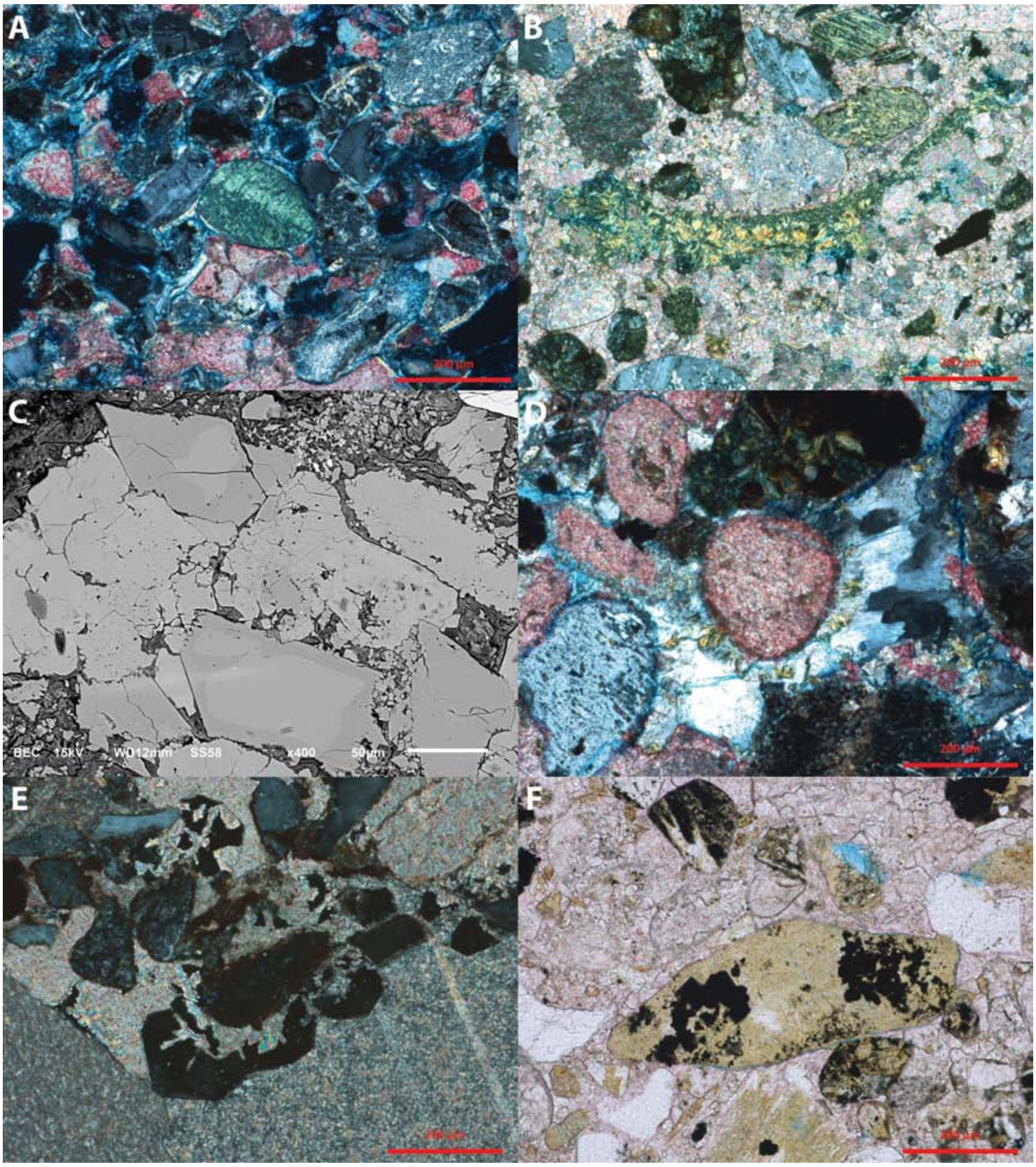


Figure 16. **A)** Discontinuous grain-covering smectite coatings in hybrid arenite. Celadonite grain and coarse calcite cementation and replacement. RJS165 2424,30. Crossed polarizers (XP). **B)** Sheath smectite filling moldic pore in bivalve and microcrystalline smectite replacing volcanic fragments. RJS108 4531,75. XP. **C**) Continuous K-feldspar overgrowths in feldspar grains. RJS486A 2877,5. BSE. **D)** Relicts of smectite rims replaced by prismatic zeolite cement in hybrid arenite with calcitized stevensite peloids/ooids. RJS36 2965,6. XP. **E)** Intergranular blocky analcime partially replaceed by calcite. RJS165 2425,10. XP. **F**) Microcrystalline pyrite replacing volcanic fragment replaced by smectite. Coarse calcite mosaic cementing and replacing grains. RJS108 4521,25. Plane polarizers (//P).

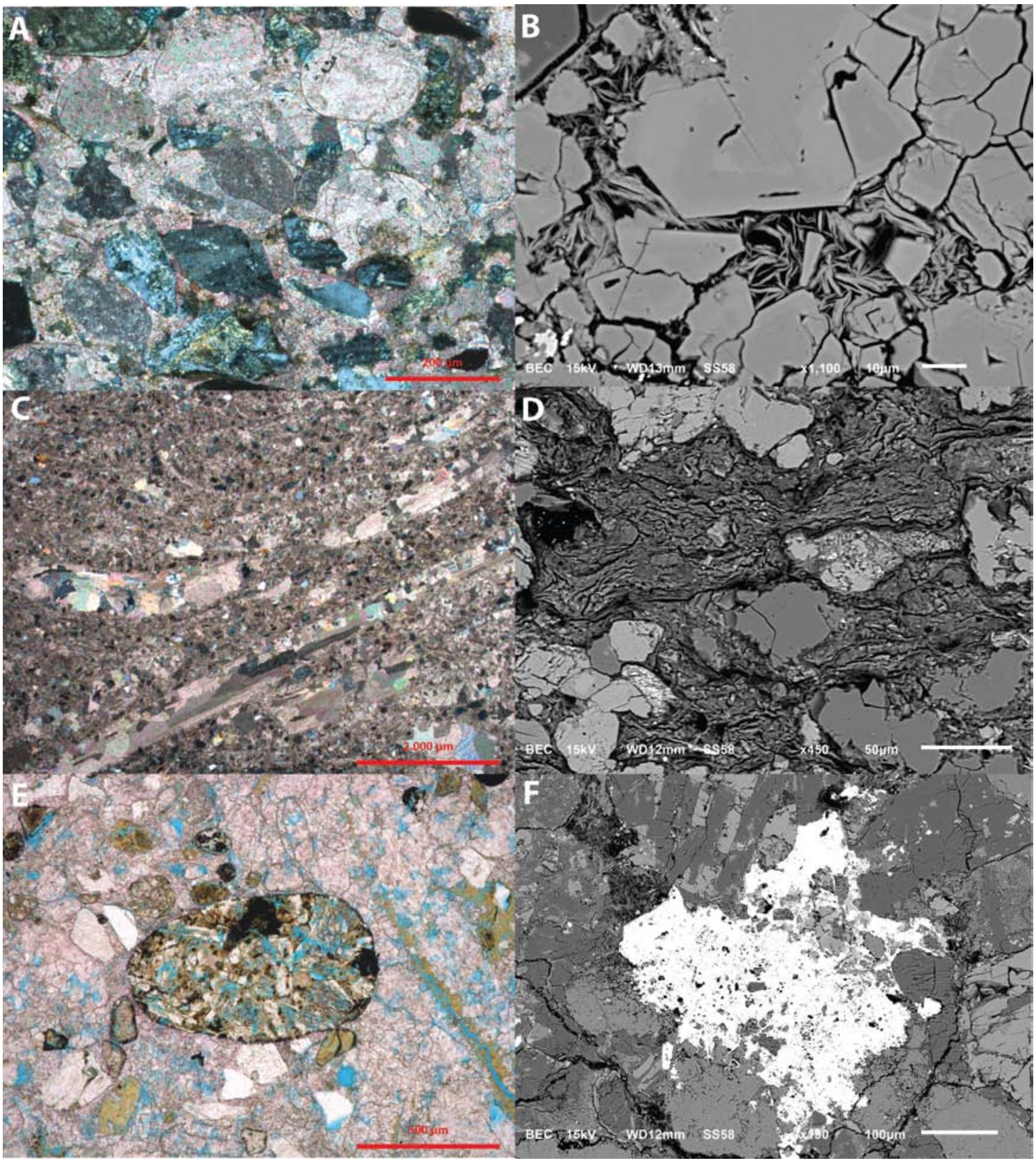


Figure 17. **A**) Macrocrystalline calcite replacing stevensite ooids. RJS108 4521,25. Crossed polarizers (XP). **B**) Macrocrystalline calcite replacing intergranular smectite. RJS36 2965,6. BSE. **C**) Recrystallized bivalve bioclasts in hybrid arenite. RJS485 2821,40. XP. **D**) Stevensite ooids and peloids, deformed by mechanical compaction. RJS486A 2877,5. BSE. **E)** Secondary intergranular porosity due to dissolution of calcite cement, and intragranular porosity in volcanic fragment and stevensitic grains. RJS108 4531,75. Plane polarizers (//P). **F)** Blocky, corrosive pyrite replacing unidentified grains. RJS36 2965,6. BSE.

A representation of the diagenetic sequence characteristic of the hybrid rocks is showed in Figure 18.

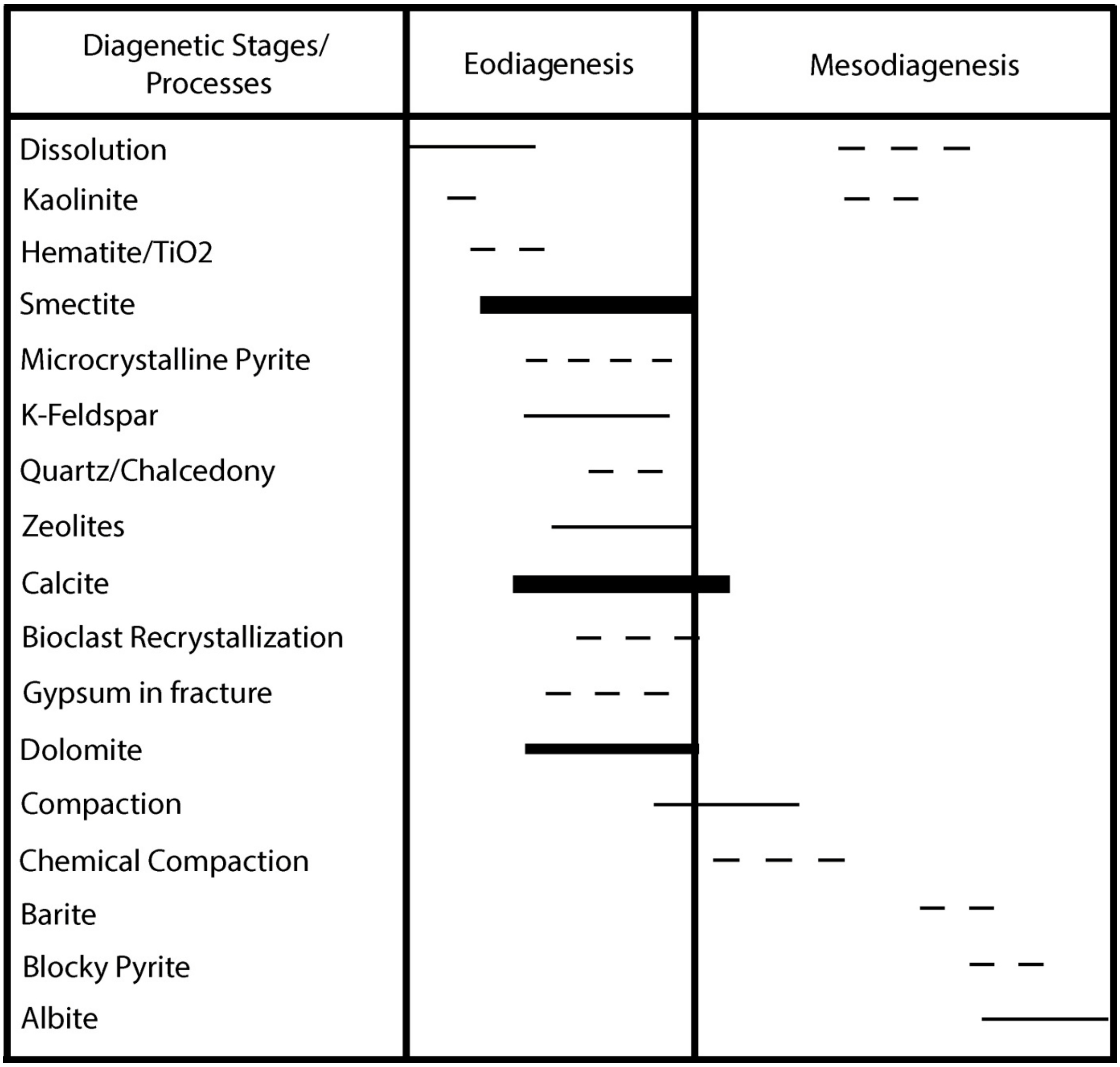


Figure 18. Diagenetic sequence for hybrid arenites.

## 7.3. Diagenetic environments

The diagenesis of the volcaniclastic sandstones and conglomerates, as of the hybrid arenites containing volcanic rock fragments, and of volcaniclastic sediments of different ages and basins (e.g. Davies et al., 1979; Mathisen, 1984; Hawlader, 1990; Tang et al., 1994; De Ros et al., 1997; Wolela and Gierlowski-Kordesch, 2007), is conditioned by the dissolution and alteration of these highly unstable constituents.

These reactions are very important, because they release diverse ions and increase the pH of interstitial fluids, promoting the precipitation of characteristic authigenic minerals (Surdam and Boles, 1979; Mathisen, 1984). The main early diagenetic product in the Lagoa Feia sediments was the precipitation of trioctahedral smectites. This process was favored by the alkaline composition of the interstitial fluids and the availability of Mg, Fe and Ca from the dissolution and alteration of basic volcanic fragments and other unstable grains, as pyroxenes, biotite and amphiboles (Davies et al., 1979; Mathisen, 1984; Hawlader, 1990; McKinley et al., 2003). The authigenesis of smectites was early, as indicated by the continuity of the rims, but still after the formation of hematite coatings in several samples.

The very early hematite precipitation, as well as the scarcity of microcrystalline pyrite, indicates that oxidizing conditions prevailed throughout significant areas of the bottom of the alkaline lakes during and just after deposition (Mathisen, 1984; Tang et al., 1994; Stonecipher, 2000). Reducing conditions, responsible for the preservation of organic matter in fine-grained sediments, should prevail only in deep depocenters where the rift source rocks were accumulated. Another byproduct of the early alteration of volcanic fragments, biotite and ferromagnesian heavy minerals was the precipitation of microcrystalline titanium oxides.

The scarce microcrystalline pyrite that replaced volcanic fragments, biotite and other ferromagnesian grains, and carbonate bioclasts in the sandstones, conglomerates and arenites is a product of dissolved sulfate reduction by bacteria (Berner, 1984). Conversely, coarse blocky pyrite that non-selectively replaced grains and diagenetic constituents is probably derived from thermal sulfate reduction during burial (Machel, 2001). A limited amount of $SO_4^{2-}$ dissolved in the rift lakes waters is inferred from the scarcity of sulfate minerals in the section. Most of the described occurrences consist of fibrous gypsum filling early tectonic or horizontal hydropressure fractures. Small amounts of macrocrystalline barite that cement and replace grains in volcaniclastic and hybrid rocks were precipitated during burial, with $SO_4^{2-}$ probably coming from the dissolution of eodiagenetic gypsum, and $Ba^{2+}$ from the dissolution and replacement of feldspars (Gluyas et al., 1997).

The diagenetic alteration of eodiagenetic smectites provided ions for the formation of later authigenic phases, as zeolites, calcite and K-feldspar (Surdam and Boles, 1979; McKinley, et al., 2003). Authigenic zeolites are common in the diagenesis of volcaniclastic sandstones, as a product of interactions between alkaline interstitial fluids and volcanic fragments or other unstable detrital constituents, as plagioclases and pyroxenes (Surdam and Boles, 1979; De Ros et al., 1997). High reaction rates are favored by the increase of Si and Al solubility in environments with pH above 9 (Taylor and Surdam, 1981). $Na^+$, $K^+$, $Ca^{2+}$ ions for the precipitation of zeolites may be sourced by a wide variety of materials, including volcanic glass and rock fragments, feldspars, smectites and other clay minerals. Zeolites commonly replace pore-lining and intragranular smectites in the analyzed samples. The formation of the different species of authigenic zeolites is controlled by the ratio between these ions, by $Al^{3+}$ and $Si^{4+}$ activities, pH, salinity and temperature (Mumpton, 1981; Hay and Sheppard, 2001). EDS analyses revealed that the zeolites of Lagoa Feia volcaniclastic sandstones and conglomerates are rich in Na, while the zeolites of hybrid arenites are commonly rich in Ca. Paradoxically, analcime, a sodic zeolite typical of the early diagenesis of volcaniclastic sandstones, was detected, in small amounts, only in the hybrid arenites.

The precipitation of overgrowths and discrete prismatic crystals of K-feldspar occurred after the authigenesis of smectites and early zeolites, as K-feldspar replaces these phases both in volcaniclastic and hybrid rocks. The same can be deduced for the rarely observed overgrowths and outgrowths of quartz. The precipitation of quartz overgrowths is very limited in the volcaniclastic sandstones and conglomerates, apparently due to small amounts of detrital quartz in these rocks, and the thick and continuous smectite rims. Early quartz cementation is relevant only in stevensitic rocks, where it fills interparticle pores as drusiform or mosaic aggregates, and replaces stevensitic grains. Voluminous silicification is, however, observed in rocks affected by the circulation of hydrothermal fluids.

The precipitation of macrocrystalline calcite took place in siliciclastic sandstones, volcaniclastic sandstones and conglomerates, and hybrid arenites after the authigenesis of smectite, zeolites and quartz, which were extensively replaced by the calcite. Calcite cementation was more intense in the hybrid arenites, probably owing to the presence of bivalve and ostracod bioclasts, and of carbonate ooids, which acted as source and nuclei for the precipitation. EDS analyses revealed that early calcite cements are devoid of Fe, and with low Mg and Mn contents. This composition is in line with the dominance of oxidizing conditions during eodiagenesis, and with the internal source provided by the carbonate bioclasts. In the stevensitic rocks, calcite cementation and replacement of ooids and peloids, and of early quartz cement was intense. Most of the replacement of stevensitic grains occurred during eodiagenesis, as result of the very alkaline composition of interstitial fluids, but some may have also taken place during burial, and related to the decarboxylation of organic matter in associated mudrocks (Morad et al., 2000).

Dolomite precipitation was much more common in the stevensitic arenites than in the other lithotypes. This is probably related to a combination of the same very high pH and Mg activity conditions responsible for the syngenetic precipitation of the stevensitic ooids and peloids, with dissolved bicarbonate supplied by organic reactions. Part of dolomite precipitation occurred during burial diagenesis, replacing eodiagenetic smectite, silica and calcite, and can thus be ascribed to decarboxylation. Stevensite ooids and peloids remained a source of Mg for dolomite, and were practically the only sites of dolomite precipitation in the hybrid arenites and in siliciclastic mudrocks.

The albitization of plagioclase grains and crystals within volcanic fragments is common in volcaniclastic sandstones and conglomerates, and in hybrid arenites. This process occurred during mesodiagenesis, probably favored by the enrichment of interstitial fluids in $Na^+$ due to the partial transformation and replacement of smectites and zeolites. The transformation of smectites in mixed-layer clays during progressive burial is classically recognized as a source for plagioclase albitization (Boles, 1982; Morad et al., 1990).

## 7.4. Implications for the quality of potential reservoirs

The petrographic and petrologic aspects observed in the non-carbonate rocks from the rift section of Campos Basin allow the development of the following implications for the quality of their quality as potential reservoirs:

The main reservoirs of the rift section correspond to bioclastic rudstones and grainstones (Bertani and Carozzi, 1985a; 1985b; Dias et al., 1988; Abrahão and

Warme, 1990; Carvalho et al., 2000; Castro, 2006). Fractured basalts constitute locally reservoirs of small expression. The sedimentary non-carbonate rocks with better potential as hydrocarbon reservoirs correspond to clastic sandstones and conglomerates with remaining intergranular porosity together with secondary porosity from the dissolution of feldspars and volcanic fragments. Volcaniclastic sandstones are commonly strongly affected by rapid diagenetic processes, which substantially reduce their porosity and permeability, but may constitute hydrocarbon reservoirs where some primary porosity is preserved, or where secondary porosity is generated by dissolution of primary or early diagenetic materials (Mathisen, 1984; Seeman and Scherer, 1984; Hawlader, 1990; Mathisen and McPherson, 1991; Reed et al., 1993; De Ros et al., 1997; Tang et al., 1997; Ryu and Niem, 1999).

Volcaniclastic Lagoa Feia sandstones and conglomerates with remaining primary intergranular porosity characteristically display smectite coatings and rims, and limited compaction and filling of intergranular pores by calcite or zeolites. Intergranular secondary pores generated by partial dissolution of calcite and zeolites, and by smectite shrinkage may locally contribute to the enhancement of porosity in these rocks. However, their eodiagenetic cementation by smectite, zeolites and calcite was commonly intense, preventing collapse of the framework due to mechanical compaction, but filling most of the primary porosity. The precipitation of smectite coatings and rims reduced substantially the permeability of these rocks, but locally prevented the total destruction of intergranular porosity, by inhibiting cementation by later phases, as zeolites and calcite. Smectite coatings and rims may preserve porosity in volcaniclastic sandstones (Mathisen and McPherson, 1991; Tang et al., 1997; McKinley et al., 2003), as also may chlorite rims, developed from smectite transformation (Hawlader, 1990; Humphreys et al., 1994; Anjos et al., 2003).

Other effects of authigenic smectites detrimental to exploration and production activities include the intense decrease of resistivity, even in oil-saturated rocks, owing to the large irreducible water saturation contained in the microporosity of smectite aggregates (Almon and Schultz, 1979; Swanson, 1995; Worthington, 2003; Reed et al., 1993), and the potential for formation damage by swelling of the smectites in contact with drilling, stimulation of injection fluids (Almon and Davies, 1978; Reed et al., 1993).

Intragranular pores within volcanic fragments and feldspar grains are common in these rocks, with subordinate contribution from moldic and fracture pores. Their present average petrographic macroporosity is of 5.2%, reaching up to 9.7%. Unfortunately, part of this porosity is filled by heavy oil or bitumen, probably a product of degradation promoted during the uplift responsible for the post-rift unconformity, what may have not occurred in deeper areas and more distant from the grabens border faults (Figure 19A). Rock fracture porosity is scarce in the analyzed samples (up to 2.3%), and obstructed partially to totally by gypsum, calcite or dolomite, but may be important in the proximity of major faults.

Stevensitic and hybrid arenites also present potential for constituting reservoirs, owing to the generation of moldic and intraparticle porosity, through dissolution of stevensitic grains or carbonate bioclasts, or the shrinkage of stevensitic grains. The macroporosity of stevensitic arenites averages 5%, reaching up to 9%. Analyzed hybrid arenites contain in average 3.2% and up to 14.7% of petrographic macroporosity. These rocks show moldic pores, intraparticle pores from dissolution of stevensitic grains or of bivalve bioclasts, from shrinkage of stevensitic grains, intragranular pores from dissolution of feldspars and volcanic fragments, besides rock fracture pores. However, their pore system is little effective in relation to permeability owing to the limited connection of these secondary pores (Figure 19B). Such rocks may constitute reservoirs where calcite cement porosity is added to the porosity from stevensitic grains dissolution and shrinkage (Figure 19C-D).

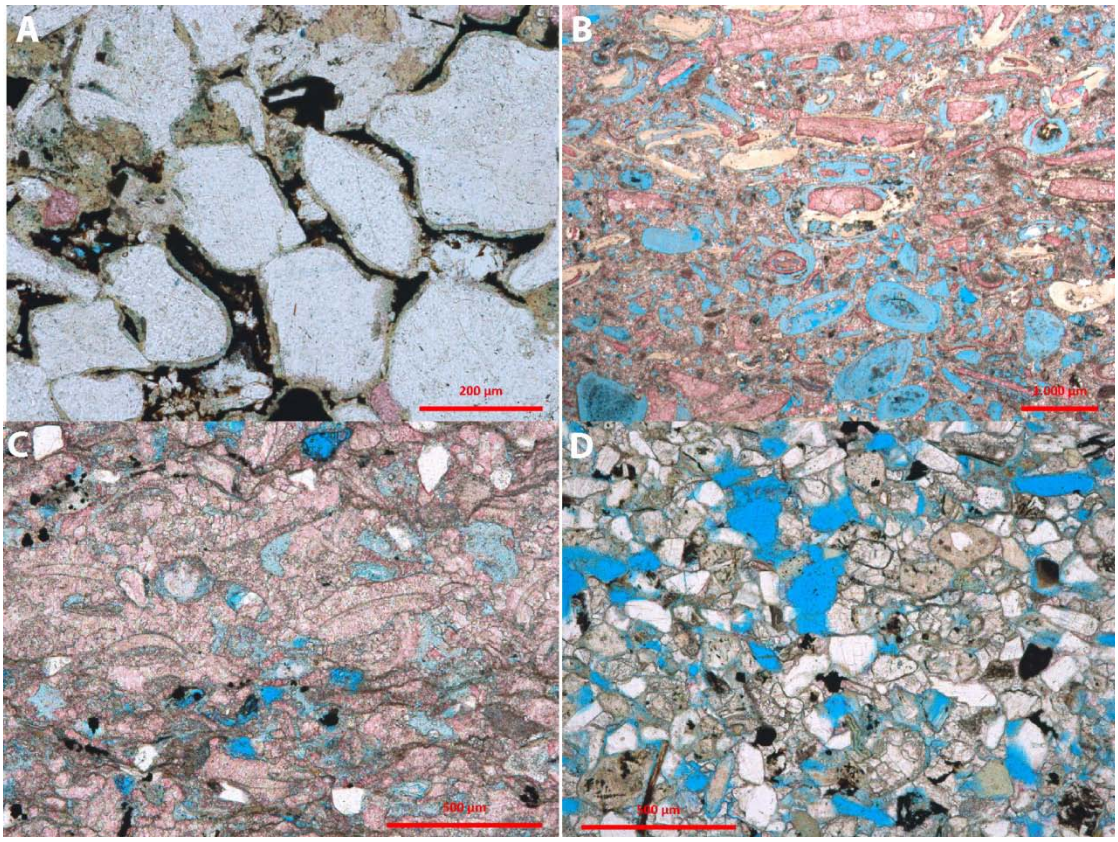


Figure 19. **A**) Bitumen-filled intergranular porosity in siliciclastic sandstone with thin smectita rims. RJS-108 3567,90. Plane polarizers (//P). **B**) Moldic and intraparticle pores derived from the dissolution of stevensitic particles. RJS514 4162,00. //P. **C**) Partial dissolution of carbonate cement in hybrid arenite, and intraparticle and moldic pores in stevensite peloids and ostracod bioclasts. PM4 2929,7. //P. **D**) Moldic pores due to dissolution of stevensitic particles in hybrid arenite. RJS165 2424,30. //P.

## 8. Conclusions

**1-** Sediments of the rift section in the Lagoa Feia Group, Lower Cretaceous of the Campos Basin, eastern Brazil, were deposited on basaltic rocks of the Cabiúnas Formation, extruded over Pre-Cambrian crystalline rocks during the initial phase of South America and Africa separation. Initial deposition took place in diverse half-graben lakes, initially isolated from each other, which later merged to form larger lakes.

**2-** The main classes of primary constituents of the analyzed rocks are clastic (siliciclastic and volcaniclastic grains, siliciclastic mud), carbonatic (bioclasts of ostracods, bivalves and rarely gastropods, intraclasts, ooids and peloids), and stevensitic (ooids, peloids, intraclasts and syngenetic laminations).

**3-** The analyzed rocks were classified into a diagram with these three classes as end members, and compositional categories defined as clastics, stevensitics and hybrids, which were the objective of this study, and also carbonatics and carbonatic-stevensitics.

**4-** Most rift sedimentation was intrabasinal, with extrabasinal contribution restricted to the proximity of the blocks border faults.

**5-** Tectonic activity, as well as magmatic and hydrothermal activity, was active during sedimentation of the rift section. Magmatic rocks comprise basalts, diabases, hydrothermal breccias, and hydromagmatic types (pepperites).

**6-** The mixture of rounded volcanic rock fragments with angular quartz, feldspars and plutonic fragments indicates the recycling of older epiclastic deposits, formed during the initial rifting stage, combined with first cycle sediments eroded from the edges of faulted and uplifted granitic-gneissic basement blocks.

**7-** The interaction between hydrothermal and lacustrine fluids with basaltic rocks and fragments represented a source of Mg and Ca for the precipitation of stevensite and calcite.

**8-** Stevensitic ooids and peloids were formed in shallow lacustrine environments of high pH, slightly agitated by waves or currents. However, the occurrence of stevensite is common in the entire rift section, and not restricted to specific intervals or areas.

**9-** The strongly alkaline environment required for the formation of stevensite would be totally intolerable for the life of bivalves. Therefore, the mixture of bivalve bioclasts and stevensite grains, which is widespread in the entire rift section, indicates a systematic gravitational re-deposition of shallow lacustrine sediments formed in different environments and times.

**10-** Intense tectonic activity during the sedimentation of the rift section would be the major mechanism for the extensive and recurrent gravitational re-deposition of shallow-lacustrine and alluvial sediments to deeper lacustrine settings.

**11-** The main diagenetic processes occurring in the clastic sandstones and conglomerates are the cementation and grain replacement by smectite, zeolites, calcite and dolomite, with relatively limited compaction and dissolution of feldspars and volcanic fragments.

**12-** The major diagenetic processes in the hybrid arenites were the same of the clastic rocks, added by dissolution and calcite replacement of stevensitic grains, and dissolution of the bioclasts.

**13-** Stevensitic arenites experienced early cementation and replacement of ooids and peloids by quartz, calcite and dolomite. Areas devoid of early cementation were intensely compacted though mechanical deformation of the ductile stevensitic grains.

**14-** The main reservoirs producing from the rift section of the Campos Basin are rudstones of bivalve bioclasts. However, volcaniclastic sandstones and conglomerates with intergranular porosity partially reduced by smectite rims together with secondary porosity from dissolution of feldspars and volcanic fragments show potential for constituting hydrocarbon reservoirs. In several samples, such porosity is filled by heavy oil or bitumen, probable product of degradation during the uplift responsible for the post-rift unconformity, what may have not occurred in areas deeper and away from the grabens marginal faults.

**15-** Stevensitic and hybrid arenites may also constitute reservoirs where secondary moldic and intraparticle porosity was generated by dissolution of bioclasts and stevensitic grains, and of calcite cement, and by shrinkage of stevensitic grains. However, the potential quality of such reservoirs is likely to be limited, owing to the poorly-connected and ineffective character of such pore systems.

**16-** The integrated petrographic-seismic-stratigraphic-sedimentologic study has revealed the dominantly intrabasinal, gravitationally re-deposited nature of the Campos Basin rift section. The understanding of the space and time controls on the depositional and diagenetic evolution of the rift carbonate and non-carbonate rocks will contribute to the planning of new exploration strategies for the basin.

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**References**


Abrahão, D. and Warme, J. E., 1990, Lacustrine and associated deposits in a rifted continental margin – lower Cretaceous Lagoa Feia Formation, Campos Basin, offshore Brazil. Lacustrine Basin Exploration: Case Studies and Modern Analogs, AAPG Memoir 50, p. 287-305.

Almeida, F. F. M. and Carneiro, C. D. R., 1998. Origem e evolução da Serra do Mar. Revista Brasileira de Geociências, v. 28, p. 135-150.

Almon, W. R., and Schultz, A. L., 1979, Electric log detection of diagenetically altered reservoirs and diagenetic traps. Transactions of the Gulf Coast Association of Geological Societies, v. 29, p. 1-10.

Anjos, S. M. C., L. F. De Ros, and C. M. A. Silva, 2003, Chlorite authigenesis and porosity preservation in the Upper Cretaceous marine sandstones of the Santos Basin, offshore eastern Brazil, in R. H. Worden, and S. Morad, eds., Clay Cements in Sandstones: IAS Special Publication, v. 34: Oxford, UK, International Association of Sedimentologists - Blackwell Scientific Publications, p. 291-316.

Berner, R. A., 1984, Sedimentary pyrite formation: An update. Geochimica et Cosmochimica Acta, v. 48, p. 605-615.

Bertani ,R. T. and Carozzi A. V., 1985a, Lagoa Feia Formation (Lower Cretaceous) Campos Basin, offshore Brazil: Rift valley type lacustrine carbonate reservoirs – I. Journal of Petroleum Geology, v. 8, n. 1, p. 37-58.

Bertani ,R. T. and Carozzi A. V., 1985b, Lagoa Feia Formation (Lower Cretaceous) Campos Basin, offshore Brazil: Rift valley type lacustrine carbonate reservoirs – II. Journal of Petroleum Geology, v. 8, n. 2, p. 199-220.

Boles, J. R., 1982, Active albitization of plagioclase, Gulf Coast Tertiary: American Journal of Science, v. 282, p. 165-180.

Bradley, W. H. and Fahey, J. J., 1962, Occurrence of stevensite in the Green River Formation of Wyoming. American Mineralogist, v. 47, p. 996-998.

Buch M. W. and Rose D., 1996, Mineralogy and geochemistry of the sediments of the Etosha Pan Region in northern Namibia: a reconstruction of the depositional environment. Journal of African Earth Sciences, v. 22, n. 3, p. 355-378.

Carminatti, M., Wolff, B. and Gamboa, L., 2008, New exploratory frontiers in Brazil. Proceedings of the 19th World Petroleum Congress. Madrid.

Carvalho, M. D., Praça, U. M., Silva-Telles, A. C., Jahnert, R. J. e Dias, J. L. 2000. Bioclastic Carbonate Lacustrine Facies Models in the Campos Basin (Lower Cretaceous), Brazil. Em E.H, Gierlowski-Kordesch and K. R. Kelts, eds., Lake Basins Through Space and Time: AAPG Studies in Geology 46, p. 245-256.

Castro, J. C., 2006, Evolução dos conhecimentos sobre as coquinas-reservatório da Formação Lagoa Feia no trend Badejo-Linguado-Pampo, Bacia de Campos. Geociências, UNESP, v. 25, n. 2, p. 175–186.

Cerling, T. E., 1996, Pore water chemistry of an alkaline lake; Lake Turkana, Kenya. In: The limnology, climatology and paleoclimatology of the East African lakes. Johnson, T.C. and Odada, E.O. (editors). Gordon and Breach Publishers, Australia. p. 225-240.

Chafetz, H.S., and Guidry, S.A., 1999, Bacterial shrubs, crystal shrubs, and ray-crystal crusts: bacterially induced vs. abiotic mineral precipitation. Sedimentary Geology, v. 126, p. 57–74.

Chahi, A., Duringer, P., Ais, M., Bouabdelli, M., Gauthier, L. F. and Fritz, B., 1999, Diagenetic transformation of dolomite into stevensite in lacustrine sediments from Jbel Rhassoul, Morocco. Journal of Sedimentary Research, v. 69, n. 5, p. 1123-1135.

Chahi, A., Fritz, B., DuPlay, J., Weber, F. and Lucas, J., 1997, Textural transition and genetic relationship between precursor stevensite and sepiolite in lacustrine sediments (Jbel Rhassoul, Morocco). Clays and Clay Minerals, v. 45, n. 3, p. 378-389.

Cuevas, J., de la Villa, R., Ramirez, S., Petit, S., Meunier, A. and Leguey, S., 2003, Chemistry of Mg smectites in lacustrine sediments from the Vicalvaro sepiolite deposit, Madrid Neogene Basin (Spain). Clays and Clay Minerals, v. 51; n. 4, p. 457-472.

Darragi, F. and Tardy, Y., 1987, Authigenic trioctahedral smectites controlling pH, alkalinity, silica and magnesium concentrations in alkaline lakes. Chemical Geology, v. 63, p. 59-72.

Davies, D. K., Almon, W. R., Bonis, S. B. and Hunter, B. E., 1979, Deposition and diagenesis of Tertiary-Holocene volcaniclastics, Guatemala, in P. A. Scholle, and P. R. Schluger, eds., Aspects of Diagenesis: SEPM Special Publication, v. 26: Tulsa, OK, Society of Economic Paleontologists and Mineralogists, p. 281-306.

De Ros, L. F., Morad, S. and Al-Aasm, I. S., 1997, Diagenesis of siliciclastic and volcaniclastic sediments in the Cretaceous and Miocene sequences of NW African margin (DSDP Leg 47A, Site 397): Sedimentary Geology, v. 112, p. 137-156.

De Ros, L. F., Goldberg, K., Abel, M., Victorinetti, F., Mastella, L. and Castro, E., 2007, Advanced Acquisition and Management of Petrographic Information from Reservoir Rocks Using the PETROLEDGE® System: AAPG Annual Conference and Exhibition, Long Beach, CA, Extended Abstracts Vol., 6 pp.

Dias, J. L., Oliveira, J. Q. and Vieira, J. C. 1988. Sedimentological and Stratigraphyc analysis of the Lagoa Feia Formation, rift phase of Campos Basin, offshore, Brazil. Revista Brasileira de Geociências, v. 18, n. 3, p. 252–260.

Dickinson, W. R., 1985. Interpreting provenance relations from detrital modes of sandstones. In: Zuffa, G. G. (Ed.). Provenance of arenites. NATO-ASI Series C. D. Reidel Pub. Co., Dordrecht, Netherlands, p. 333-361.

Dyni, J. R., 1976, Trioctahedral smectite in the Green River Formation, Duchesne County. Utah. U.S. Geol. Surv. Prof. Pap. 967, 14 pp.

Eberl, D. D., Jones, B.F. and Khoury, H.N., 1982, Mixed-layer kerolite/ stevensite from the Amargosa Desert, Nevada. Clays and Clay Minerals, v. 30; n. 5, p. 321-326.

Folk, R. L., 1968, Petrology of sedimentary rocks. Austin: Hemphill's. 182p.

Furquim, S. A. C., Graham, R. C., Barbiero, L., Queiroz Neto, J. P. and Vallès V., 2008, Mineralogy and genesis of smectites in an alkaline-saline environment of Pantanal Wetland, Brazil. Clays and Clay Minerals; v. 56; no. 5; p. 579-595.

Gluyas, J., Jolley, L. and Primmer, T.J., 1997, Element mobility during diagenesis; sulfate cementation of Rotliegend sandstones, southern North Sea. Marine Petroleum Geology, v. 14, p. 1001-1011.

Guardado, L.R., Gamboa, L.A.P. and Lucchesi, C.F., 1990. Petroleum geology of the Campos Basin, Brazil, a model for a producing Atlantic type basin. In: Edwards, J.D and Santogrissi, P.A., Divergent / Passive Margin Basins. Tulsa, Okla., AAPG Memoir 48, p. 3-79.

Guardado, L. R., Spadini, A. R., Brandão, J. S. L. and Melo, M. R., 2000, Petroleum System of the Campos Basin.In: Mello, M.R. and Katz, B. eds., Petroleum System of South Atlantic Margins: AAPG Memoir 73, p. 317–324.

Hawlader, H. M., 1990, Diagenesis and reservoir potential of volcanogenic sandstones - Cretaceous of the Surat Basin, Australia: Sedimentary Geology, v. 66, p. 181-195.

Hay R. L. and Sheppard, R.A., 2001, Occurrences of zeolites in sedimentary rocks. In: Bish, D.L., Ming, D.W. (eds) Natural Zeolites: occurrence, properties; applications. Reviews in Mineralogy and Geochemistry, Vol. 45, Mineralogical Society of America, Washington, DC, pp. 217-234.

Heilbron, M., Valeriano C. M., Tassinari, C. C. G., Almeida J. C. H., Tupinambá M., Siga Jr. O. and Trouw, R. J. A., 2008, Correlation of Neoproterozoic terranes between the Ribeira Belt, SE Brazil and its African counterpart: comparative tectonic evolution and open questions. In: Pankhurst, R.J.; Trouw, R.A.J, Brito Neves; B.B., De Wit M.J. (Eds). West Gondwana Pre-Cenozoic Correlations Across the South Atlantic Region. The Geological Society of London. London 294, p.211-237.

Hover, V. C. and Ashley, Gail, M., 2003, Geochemical signatures of paleodepositional and diagenetic environments; a STEM/ AEM study of authigenic clay minerals from an arid rift basin, Olduvai Gorge, Tanzania. Clays and Clay Minerals, v. 51; n. 3, p. 231-251.

Hover, V. C. Walter, L. M. Peacor, D. R. and Martini, A. M., 1999, Mg-smectite authigenesis in a marine evaporative environment, Salina Ometepec, Baja California. Clays and Clay Minerals, v. 47, n. 3, p. 252-268.

Humphreys, B., Kemp, S. J., Lott, G. K., Bermanto, D., Dharmayanti, A. and Samsori, I., 1994, Origin of grain-coating chlorite by smectite transformation: an example from Miocene sandstones, North Sumatra back-arc basin, Indonesia: Clay Minerals, v. 29, p. 681-692.

Jones, B. F. and Weir, A. H., 1983, Clay minerals of Lake Abert, an alkaline, saline lake. Clays and Clay Minerals, v. 31; n. 3, p. 161-172.

Jones, B. F., 1986, Clay mineral diagenesis in lacustrine sediments. U. S. Geological Survey Bulletin v. 1578, p. 291-300.

Khoury, H. N., Eberl, D. D. and Jones, B. F., 1982, Origin of magnesium clays from the Amargosa Desert, Nevada. Clays and Clay Minerals, v. 30; n. 5, p. 327-336.

Locke, A., 2008, Tabulated observations of the pH tolerance of marine and estuarine biota. Fisheries and Oceans Canada, Moncton, NB, 28 pp.

Machel, H. G., 2001, Bacterial and thermochemical sulfate reduction in diagenetic settings - old and new insights. Sedimentary Geology 140, 143-175.

Martin de Vidales, J. L., Pozo, M, Alia, J. M., Garcia-Navarro, F. and Rull, F., 1991, Kerolite-stevensite mixed-layers from the Madrid Basin, central Spain. Clay Minerals, v. 26; n. 3, p. 329-342.

Mathisen, M. E., 1984, Diagenesis of Plio-Pleistocen Non-marine sandstones, Cagayan Basin, Philipines: early development of secondary porosity in volcanic sandstones, in: Surdam, R., McDonald, D.A. (Eds.), Clastic Diagenesis. American Association of Petroleum Geologists, Tulsa, OK, pp. 177-193.

Mathisen, M. E., and McPherson, J. G. ,1991, Volcaniclastic deposits: implications for hydrocarbon exploration, in R. V. Fisher, and G. A. Smith, eds., Sedimentation in Volcanic Settings: SEPM Special Publication, v. 45: Tulsa, OK, Society of Economic Paleontologists and Mineralogists, p. 27-36.

Mayayo, M. J., Bauluz, B. and Gonzalez-Lopez, J. M., 2000, Variations in the chemistry of smectites from the Calatayud Basin (NE Spain). Clay Minerals, v. 35; n. 2, p. 365-374.

McKinley, J. M., Worden, R. and Ruffell, A. H., 2003, Smectite in sandstones: a review of the controls on occurrence and behavior during diagenesis, in: Worden, R.H., Morad, S. (Eds.), Clay Cements in Sandstones. International Association of Sedimentologists - Blackwell Scientific Publications, Oxford, UK, pp. 109-128.

Mello, M. R, Mohriak, W. U., Koutsoukos, E. A. M. and Bacoccoli, G., 1994, Selected petroleum systems in Brazil. In: Magoon, L.B. and Dow., W.G. (Eds.), The Petroleum System – from source to trap: AAPG Memoir 60, p. 499-512.

Milani, E.J. and Thomaz Filho, A., 2000. Sedimentary basins of South America. In: Cordani, U.G.; Milani, E.J.; Thomaz Filho, A.; and Campos, D.A. (Eds.), Tectonic evolution of South America. Rio de Janeiro, 31st International Geological Congress, pp. 389-452.

Mizusaki A. M. P., Thomaz Filho A. and Valença J. G., 1988, Volcano-sedimentary sequence of Neocomian age in Campos Basin (Brazil). Revista Brasileira de Geociências, v.18, p. 247-251.

Mizusaki, A.M.P., Petrini, R., Bellieni, G., Comin-Chiaramonti, P., Dias, J., De Min, A. and Piccirillo, E.M. 1992. Basalt magmatism along the passive continental margin of SE Brazil (Campos basin). Contribution to Mineralogy and Petrology, v.111: p.143-160.

Mohriak, W. U., Mello, M. R., Dewey, J. F. and Maxwell, J. R., 1990, Petroleum geology of the Campos Basin, offshore Brazil. In: Brooks, J. (Ed.), Clastic Petroleum Provinces: Geological Society of London Special Publication 50, p. 119-142

Morad, S., Bergan, M. Knarud, R. and Nystuen, J. P., 1990, Albitization of detrital plagioclase in Triassic reservoir sandstones from the Snorre Field, Norwegian North Sea: Journal of Sedimentary Petrology, v. 60, p. 411-425.

Morad, S., Ketzer, J. M., De Ros, L. F., 2000, Spatial and temporal distribution of diagenetic alterations in siliciclastic rocks: implications for mass transfer in sedimentary basins. Sedimentology, v. 47, p. 95-120.

Mumpton, F. A., 1981, Mineralogy and Geology of Natural Zeolites. Reviews in Mineralogy, Volume 4: Washington, Mineralogical Society of America, 233 pp.

Noack, Y., Decarreau, A., Boudzoumou, F. and Trompette, R. (1989) Low-temperature oolitic talc in upper Proterozoic rocks, Congo. Journal of Sedimentary Research, v. 59, p. 717-723.

Odin, G. S., 1988, Green Marine Clays. Developments in Sedimentology 45. Elsevier, Amsterdam, 445 pp.

Pozo, M. and Casas, J., 1999, Origin of kerolite and associated Mg clays in palustrine-lacustrine environments; the Esquivias Deposit (Neogene Madrid Basin, Spain). Clay Minerals, v. 34; n. 3, p. 395-418.

Rangel, H. D., Martins, F. A. L. Esteves, F. R. and Feijó, F. J., 1994, Bacia de Campos, Boletim de Geociências da Petrobras v. 8 (1), p. 203-218.

Rangel, H. D. and Carminatti, M., 2000, Rift lake stratigraphy of the Lagoa Feia Formation, Campos Basin, Brazil. In: Gierlowski-Kordesch, E. H. and Kelts, K. R., eds., Lake Basins Through Space and Time: AAPG Studies in Geology 46, p. 225-244.

Reed, J. K., Gipson, M., Jr., Neese, D. G., 1993, Hydrocarbon potential of sandstone reservoirs in the Neogene East Slovakian Basin, Part 1: a petrographic examination of lithology, porosity, and diagenesis. Journal of Petroleum Geology 16, 89-108.

Rehim, H. A. A. A., Pimentel, A. M., Carvalho, M. D. and Monteiro, M., 1986, Talco e estevensita na Formação Lagoa Feia da bacia de Campos - possíveis implicações no ambiente deposicional. Anais XXXIV Congresso Brasileiro de Geologia, v. 1, p. 416-422.

Ryu, I.-C., and Niem, A. R., 1999, Sandstone diagenesis, reservoir potential, and sequence stratigraphy of the Eocene Tyee Basin, Oregon: Journal of Sedimentary Research, v. 69, n. 2, p. 384-393.

Schaller, H., 1973, Estratigrafia da Bacia de Campos. In: 27º Congresso Brasileiro de Geologia,. Aracaju. Sociedade Brasileira de Geologia, Anais v. 3, p. 247-258.

Scholle, P. A. and Ulmer-Scholle, D. S., 2003, A Color Guide to the Petrography of Carbonate Rocks: Grains, textures, porosity, diagenesis. AAPG Memoir 77, The American Association of Petroleum Geologists, Tulsa Okla., 474 pp.

Seemann, U. and Scherer, M., 1984, Volcaniclastics as potencial hydrocarbon reservoirs. Clay Minerals, v. 9, p. 457-470.

Stonecipher, S. A., 2000, Applied Sandstone Diagensis–Practical Petrographic Solutions for a Variety of Common Exploration, Development, and Production Problems. SEPM Short Course Notes no. 50. Tulsa, Okla. 143 pp.

Surdam, R.C. and Boles, J. R., 1979, Diagenesis of volcanic sandstones. In Scholle, P. A. and Schluger, P.R., (Eds.) Aspects of Diagenesis. SEPM Special Publication 26, Society of Economic Paleontologists and Mineralogists, Tulsa, Okla. p. 227-242.

Swanson, B. F., 1995, Microporosity in reservoir rocks; its measurement and influence on electrical resistivity: SPE Reprint Series, v. 39, p. 94-104.

Tang, Z., Parnell, J. and Ruffell, A. H., 1994, Deposition and diagenesis of the lacustrine-fluvial Cangfanggou Group (uppermost Permian to Lower Triassic), southern Junggar Basin, NW China: a contribution from sequence stratigraphy: Journal of Paleolimnology, v. 11, p. 67-90.

Tang, Z., Parnell, J. and Longstaffe, F.J., 1997, Diagenesis and reservoir potential of Permian–Triassic fluvial/lacustrine sandstones in the Southern Junggar Basin, Northwestern China. AAPG Bulletin, v. 81, p. 1843-1865.

Taylor, M.W. and Surdam, R.C., 1981, Zeolite reactions in the tuffaceous sediments at Teels Marsh, Nevada. Clays and Clay Minerals, v. 29, p. 341-352.

Tettenhorst, R. and Moore, G. E. Jr., 1978, Stevensite oolites from the Green River Formation of central Utah. Journal of Sedimentary Petrology, v. 48; n. 2, p. 587-594.

Van Houten, F. B. and Purucker, M. E., 1984, Glauconite peloids and chamosite ooids--favourable factors, constraints and problems. Earth Sci. Rev., v. 20, p. 211-243.

Yuretich, R. F. and Ervin, C. R., 2002, Clay minerals as paleoenvironmental indicators in two large lakes of the African Rift Valleys: Lake Malawi and Lake Turkana. In: Sedimentation in Continental Rifts. SEPM Special Publication No 73, p. 221-232.

Winter, W. R., Jahnert, R. J. and França, A. B., 2007, Bacia de Campos. Boletim de Geociências da Petrobras, v. 15, n. 2, p. 511–529.

Wolela, A. M., and Gierlowski-Kordesch, E. H., 2007, Diagenetic history of fluvial and lacustrine sandstones of the Hartford Basin (Triassic-Jurassic), Newark Supergroup, USA: Sedimentary Geology, v. 197, p. 99-126..

Worthington, P. F., 2003, Effect of clay content upon some physical properties of sandstone reservoirs, in R. H. Worden, and S. Morad, eds., Clay Cements in Sandstones: IAS Special Publication, v. 34: Oxford, UK, International Association of Sedimentologists - Blackwell Scientific Publications, p. 191-211.

Wright, V. P., 2012, Lacustrine carbonates in rift settings: the interaction of volcanic and microbial processes on carbonate deposition. In: Garland, J. Neilson, J. E. Laubach, S. E. and Whidden K. J. (eds.) Advances in Carbonate Exploration and Reservoir Analysis. Geological Society, London. Special Publication 370, p. 39-47.

# 11. ANEXOS

## Anexo 1 – Tabela de resultados da petrografia quantitativa

## Anexo 2 – Descrições petrográficas

## Anexo 3 – Fotomicrografias

## Anexo 4 – Analises de BSE – EDS

| ANEXO I |
| --- |
| Título da Dissertação: **Deposição, Diagênese e Qualidade dos Potenciais Reservatórios Sedimentares Não-Carbonáticos da Seção Rifte da Bacia de Campos** |
| Área de Concentração: **Estratigrafia** |
| Autor: **Garibaldi Armelenti** |
| Orientador: **Prof. Dr. Luiz Fernando De Ros** |
| Examinador: **Profa. Dra. Ana Maria Góes** |
| Data: **10 de Junho de 2014** |
| Conceito: **A** |
| PARECER: |
| Excelente produto final referente à dissertação do estudante Garibaldi Armelenti intitulado "DEPOSIÇÃO, DIAGÊNESE E POTENCIAL DE RESERVATÓRIO DAS ROCHAS SEDIMENTARES NÃO-CARBONÁTICAS DA SEÇÃO RIFTE DA BACIA DE CAMPOS" do Programa de Pós-Graduação em Geociências da Universidade Federal do Rio Grande do Sul. |
| Considero que o estudante Garibaldi cumpriu os objetivos previstos em um mestrado em Geociências. Pela análise do documento, nota-se o envolvimento do estudante na busca da caracterização das rochas não carbonáticas do Grupo Lagoa Feia da Bacia de Campos. Para isso, utilizou fortemente as técnicas da petrologia sedimentar com auxilio do microscópio petrográfico e o eletrônico de varredura. |
| Os resultados produzidos são robustos e de alto nível. Demonstrou capacidade de organização e sua redação, de modo geral, tanto em português quanto em inglês, é clara, concisa. Suas ilustrações são abundantes e no geral de ótima qualidade. O artigo submetido, bem organizado, redigido de forma clara e bem ilustrado, apresenta relevante contribuição sobre as rochas terrígenas e vulcanoclásticas que compõe o Grupo Lagoa Feia, descrição e interpretação com vistas ao potencial de reservatório. |
| Pelo exposto anteriormente, considero que sua dissertação está aprovada. No sentido de contribuir com o aperfeiçoamento deste trabalho, tenho uma lista de sugestões e indagações, que são as seguintes: |
| GERAL: Considero necessário que o documento apresente tanto na sua parte inicial como no artigo, as localizações da área de estudo e dos poços selecionados. |

Além disso, são necessárias as seções colunares dos poços com a identificação das litofácies, unidades estratigráficas e localização das lâminas estudadas.

PRIMEIRA PARTE INTRODUÇÃO

Colocar claramente os objetivos desta pesquisa, que é a caracterização das rochas terrígenas, vulcanoclásticas e estevensíticas do Grupo Lagoa Feia da Bacia de Campos para verificação do potencial petrolífero em termos de reservatório. O item métodos está bem desenvolvido principalmente na parte das rochas estevensíticas e petrografia, diagênese e potencial petrolífero das rochas vulcanoclásticas. Senti falta da apresentação do diagrama de Folk 1968 e de um maior detalhamento a respeito da identificação das rochas estevensíticas, critérios químicos e petrográficos.

No item principais resultados e interpretações, a frase: "A estevensita é abundante em toda seção rifte, não estando restrita a intervalos deposicionais determinados" deveria ser demonstrada nos perfis dos poços.

Outros: primeiro parágrafo p. 9 erro de concordância e ortografia:

Winter *et al.* (2007) separaram a evolução tectônica e estratigráfica da Bacia de Campos em três supersequências: Superssequência Rifte, : Superssequência Pós-Rifte e : Superssequência Drifte.

p.13 item 4.2 acrescentar referencias bibliográficas

Faltaram as adições no item Referências Bibliográficas de: Chang et al. 1996, McKinley et al. 2001, Jones (1996) e Tosca & Masterson (in press). Blatt 1979, Guardado et al. 1989 são citadas no item Ref. Bibl, porém não no corpo do texto. Melhorar as figuras 2e 3 com o aumento tamanho de letra para facilitar leitura.

SEGUNDA PARTE; ARTIGO

Faltaram as adições no item Referências Bibliográficas: Almon and Davies 1978 ANP, 2003; Dickson 1965; Jones 1996 Tosca and Mastersn in press

Padronizar a sexta referencia da p. 103. No item geological setting caberia mais citações bibliográficas como, por exemplo, nas p.46 segundo e terceiro parágrafos e p.47 segundo parágrafo.

Assinatura: Ana Maria Góes Data: 10/06/2014

Ciente do Orientador:

Ciente do Aluno:

## ANEXO I

| Título da Dissertação/Tese: |
|---|
| "DEPOSIÇÃO, DIAGÊNESE E POTENCIAL DE RESERVATÓRIO DAS ROCHAS SEDIMENTARES NÃO-CARBONÁTICAS DA SEÇÃO RIFTE DA BACIA DE CAMPOS" |
| Área de Concentração: ESTRATIGRAFIA |
| Autor: **GARIBALDI ARMELENTI** |
| Orientador: Prof. Dr. Luiz Fernando de Ros |
| Examinador: Profa. Dra. Ana Maria Pimentel Mizusaki |
| Data: 23/05/2014 |
| Conceito: A (excelente) |

PARECER:

A dissertação de mestrado sobre "Deposição, diagênese e potencial de reservatório das rochas sedimentares não-carbonáticas da seção rifte da Bacia de Campos" apresenta-se muito bem estruturada, organizada, texto escrito de maneira clara e de leitura extremamente didática. Os objetivos são claros e foram integralmente cumpridos.

A dissertação foi apresentada sob forma de artigo, já submetido à periódico internacional. Verifica-se assim o interesse e o avanço científico obtido pelo mestrando.

O texto, em geral, está bem redigido, com raros erros ortográficos e os termos técnicos utilizados de forma precisa.

Assim, em termos gerais, o texto traduz toda a extensão da pesquisa e o excelente trabalho realizado. Pequenas sugestões podem ser colocadas mas não comprometem a pesquisa realizada:

- figuras do capítulo introdutório como, por exemplo, figuras 2 e 3 podem ser melhoradas pois há dificuldade para leitura;
- na descrição da "metodologia", o autor comenta que foi realizada a "descrição petrográfica quantitativa de 42 lâminas delgadas". No entanto, não apresenta os critérios que foram utilizados

para a seleção das amostras visando a confecção das lâminas petrográficas. Sente-se falta também da real área de estudo pois o texto comenta sobre a Bacia de Campos. No artigo apresentado, já comenta-se que foram analisados testemunhos de 10 poços. Onde estes poços localizam-se? Esta dúvida surge também com a análise das figuras apresentadas no artigo técnico pois há referência de poços e prováveis profundidades para coleta das amostras. Como exemplo, Figura 5 (página 54): A) ..... RJS 113 2918,00 (profundidade?).

- na metodologia também é interessante citar onde as análises foram realizadas.
- as conclusões apresentadas no capítulo introdutório reproduzem as conclusões do artigo tornando-se repetitivas.

Agradeço a participação na avaliação do trabalho e parabenizo o mestrando e orientador pela excelência da pesquisa.

Assinatura: Ana Maria Mizusaki Data: 23/05/2014

Ciente do Orientador:

Ciente do Aluno:

**ANEXO I**

Título da Dissertação/Tese:

**"DEPOSIÇÃO, DIAGÊNESE E POTENCIAL DE RESERVATÓRIO DAS ROCHAS SEDIMENTARES NÃO-CARBONÁTICAS DA SEÇÃO RIFTE DA BACIA DE CAMPOS"**

Área de Concentração: ESTRATIGRAFIA

Autor: **GARIBALDI ARMELENTI**

Orientador: Prof. Dr. Luiz Fernando de Ros

Examinador: Prof. Dr. Leonardo Fonseca Borghi de Almeida

Data: 27/8/2014

Conceito: A - Excelente

PARECER:

A dissertação está bem acima da expectativa. Apresenta-se bem formatada e ilustrada, bem redigida de leitura fácil. Apesar do formato não tradicional da parte pré-textual, contém objetivos claros, para os quais os materiais e métodos coadunam-se e estão adequados. Os resultados, sob forma de artigos, são abundantes, bem expostos e relevantes para o conhecimento petrogenético da seção Rifte, em particular da bacia de Campos. Trata-se de uma contribuição inovadora não só para o conhecimento da margem continental brasileira, mas pelos aspectos de abordagem de classificação petrográfica de rochas híbridas e da diagênese a que se submeteram. Sem dúvida, uma contribuição à Geologia Sedimentar do País. Assim sou de parecer favorável à aprovação sem modificações. Parabéns ao mestrando e ao orientador.

Rio de Janeiro, 27 de agosto de 2014

Leonardo Borghi

Prof. Leonardo Fonseca Borghi de Alm[illegible]
Instituto de Geociências da UFRJ
Matr. 0114073

Assinatura: Leonardo Borghi Data: 27/8/2014

Ciente do Orientador:

Ciente do Aluno: