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EVOLUÇÃO SISMOESTRATIGRÁFICA DA PORÇÃO SUBMERSA DO DELTA DO RIO SÃO FRANCISCO

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SALVADOR

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Dissertação de mestrado apresentada ao Programa de Pós-Graduação em Geologia do Instituto de Geociências da Universidade Federal da Bahia como requisito parcial à obtenção do Título de Mestre em Geologia, Área de Concentração: Geologia Marinha, Costeira e Sedimentar.

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Aos meus pais,

que sempre lutaram pelo meu crescimento

moral e intelectual.

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RESUMO

O presente estudo utilizou registros de sísmica rasa de alta resolução para caracterizar a arquitetura e evolução estratigráfica da última Sequência Deposicional do Delta do São Francisco (SFDS), depositada durante o final do Pleistoceno/Holoceno. A SFDS acumulou-se em uma depressão batimétrica da plataforma continental associada à cabeceira do cânion do São Francisco. A existência dessa depressão resultou em um espaço de acomodação adicional de várias dezenas de metros, fato que permitiu o acúmulo de um registro estratigráfico contínuo da subida do nível eustático do mar desde o Último Máximo Glacial (UMG) em uma das plataformas continentais mais estreitas e rasas do mundo. Seis grandes unidades sismoestratigráficas foram individualizadas. As unidades inferiores (Su1, Su2 e Su3) se acumularam em torno da cabeceira do cânion, limitadas pelas paredes da depressão batimétrica. A Su1 foi possivelmente afogada pelo MWP1A (*Melt Water Pulse 1A*). As unidades Su2 e Su3 acumularam-se durante o período subsequente de reduzidas taxas de subida do nível eustático do mar entre o final do MWP1A e o Younger Dryas (YD). O topo do Su3 passa lateralmente para um terraço de abrasão esculpido nas paredes da depressão batimétrica. Este terraço está localizado entre 55 e 60 m abaixo do nível do mar atual, e provavelmente marca a posição da linha costa durante o YD. A Su3 foi afogada pelo MWP1B (*Melt Water Pulse 1B*), como indicado por cunhas de sedimentos (Su6) que enterraram o topo desta unidade nas porções laterais da depressão batimétrica. A Su4 foi depositada ainda dentro de um embaiamento costeiro, durante um período de reduzidas taxas de subida do nível do mar após o MWP1B. A Su4 foi aparentemente afogada pelo MWP1C por volta de 8,4 ka BP, marcando a formação da superfície de inundação máxima. A Su5, que corresponde à unidade mais recente do delta, foi depositada somente após a estabilização do nível eustático do mar que iniciou em 8-7,5 ka AP. Este estudo demonstrou como as variações nas taxas de subida do nível eustático do mar combinadas com a morfologia antecedente favoreceram a criação de um registro sedimentar contínuo da transgressão do Holoceno.

Palavras-chave: Deltas dominados por ondas; Variações do Nível do Mar; Sismoestratigrafia.

ABSTRACT

The present study used shallow high-resolution seismic surveys to characterize the architecture and stratigraphic evolution of the last depositional sequence of the São Francisco Delta (SFDS), deposited during the late Pleistocene/Holocene. The SFDS accumulated in a bathymetric depression of the continental shelf associated with the São Francisco canyon head. The existence of this depression resulted in additional accommodation space measuring several tens of meters, which allowed the accumulation of a continuous stratigraphic record of eustatic sea-level rise since the Last Glacial Maximum on one of the shallowest and narrowest continental shelves in the world. Six major seismostratigraphic units were individualized. The lower units (Su1, Su2 and Su3) accumulated around the head of the canyon, limited by the walls of the bathymetric depression. Su1 was possibly drowned by MWP1A. Units Su2 and Su3 accumulated over the ensuing period of reduced rates of eustatic sea-level rise between the end of MWP1A and the Younger Dryas (YD). The top of Su3 transitions laterally to a wave-cut terrace engraved in the walls of the bathymetric depression. This terrace is located between 55 and 60 m below present sea level, and probably marks shoreline position during the YD. Su3 was drowned by MWP1B, as indicated by sediment wedges (Su6) that buried the top of this unit in the lateral portions of the bathymetric depression. Su4 was deposited still within a coastal embayment, during a period of low sea-level rise rates after MWP1B. Su4 was apparently drowned by MWP1C around 8.4 ka BP, marking the formation of the maximum flooding surface. The deposit of Su5, which corresponds to the most recent unit of the delta, only took place after the stabilization of eustatic sea level that began in 8-7.5 ka BP. This study demonstrated how variations in eustatic sea-level rise rates combined with antecedent local morphology favored the creation of a continuous sedimentary record of Holocene transgression.

Keywords: wave-dominated delta; sea-level changes; seismic stratigraphy.

I do not know what I may appear to the world,

but to myself I seem to have been only like a boy playing on the seashore,

and diverting myself in now and then finding a smoother pebble or

a prettier shell than ordinary,

whilst the great ocean of truth lay all undiscovered before me.

Isaac Newton

SUMÁRIO

CAPÍTULO 1
INTRODUÇÃO GERAL

O termo delta foi inicialmente proposto por Heródoto, no século IV A.C., ao reconhecer que as planícies aluviais entre os distributários do rio Nilo apresentavam o formato da letra grega "Δ" (BHATTACHARYA, 2006). Somente na primeira metade do século XIX, o termo foi introduzido na literatura geológica por Lyell, que o definiu como terras aluviais, de qualquer formato, depositadas por um rio em sua desembocadura (LE BLANC, 1975).

Deltas têm sido atualmente definidos como ambientes transicionais associados a desembocaduras de rios, onde os aportes sedimentares fluviais resultam em progradação da linha de costa, podendo incluir ainda os sedimentos retrabalhados pela ação de ondas e marés. Por isso, constituem feições extremamente dinâmicas, sujeitos ainda à subsidência, inundações e processos erosivos (COLEMAN; WRIGHT, 1975; GALLOWAY, 1975; DOMINGUEZ; 1996). Por se tratar de uma acumulação sedimentar construída ao nível do mar, as planícies deltaicas são extremamente sensíveis, mesmo às pequenas variações deste nível. Assim, os deltas só se formam, quando o suprimento de sedimento que chega até a bacia receptora excede a capacidade de retrabalhamento dos processos bacinais (e.g. ondas e marés) e as taxas de subida do nível relativo do mar. Estes fatores exercem, portanto, importantes controles na evolução destas feições (DOMINGUEZ; 1990; 1996; BHATTACHARYA, 2006).

Apesar da extrema relevância dos ambientes deltaicos para o desenvolvimento de importantes civilizações desde a Antiguidade, como a do Egito, por oferecer diversas facilidades para as populações que neles se estabeleceram, como a proximidade a cursos fluviais e à zona costeira, bem como a presença de terras férteis, os primeiros trabalhos científicos sobre deltas foram publicados apenas no final do século XIX por Credner em 1878 e Gilbert em 1885 e 1890 (LE BLANC, 1975). Contudo, a intensificação do estudo dos ambientes deltaicos ocorreu apenas em meados do século XX, com o objetivo de reconhecer possíveis ambientes geradores e armazenadores de hidrocarbonetos (LE BLANC, 1975).

Atualmente é possível encontrar um grande número de trabalhos publicados na literatura que se propuseram a estudar os diversos deltas ao redor do mundo. A leitura desses trabalhos mostra que embora cada delta possua suas particularidades, existem fatores em comum envolvidos na sua formação. Os fatores frequentemente mencionados que exercem maior influência são as variações do nível do mar, correntes longitudinais e mudanças climáticas, principalmente nos deltas dominados por ondas.

As variações do nível do mar, alvo deste trabalho, não foram consideradas importantes para a evolução deltaica nos trabalhos inicialmente propostos por Coleman e Wright (1975) e Galloway (1975). A influência das variações do nível do mar como principal fator controlador dos ambientes deltaicos foi investigado no final do século passado por diversos autores (SUGUIO et al., 1985; DOMINGUEZ et al., 1987; DOMINGUEZ, 1990; DOMINGUEZ et al., 1992; MARTIN et al., 1996; DOMINGUEZ, 1996).

A ideia dominante em trabalhos publicados sobre os deltas brasileiros (BITTENCOURT et al., 1982; DOMINGUEZ et al., 1992; DOMINGUEZ, 1996; MARTIN et al., 1996) era a de que o abaixamento do nível relativo do mar desde 5,7 ka AP havia favorecido a progradação da linha de costa, com os rios desempenhando um papel secundário. Entretanto, em seu trabalho sobre a evolução do delta do São Francisco, Guimarães (2010) mostrou que o início da progradação do delta antecedeu em pelo menos 2100 anos o máximo alcançado pelo nível do mar. Os resultados encontrados por Guimarães (2010) apresentam uma melhor concordância com o que tem-se reconhecido em outras regiões do mundo, nas quais o aporte sedimentar fluvial é o principal fator na construção dos sistemas deltaicos, visto que a superação do espaço de acomodação pelo aporte sedimentar permite que o delta possa avançar em direção ao oceano, sem que haja, necessariamente, um abaixamento do nível relativo do mar (STANLEY; WARNE, 1994; HORI; SAITO, 2002; TA et al., 2002; LIU et al., 2004; TÖRNQVIST et al., 2004; TANABE et al., 2006; TAMURA et al., 2009; GUIMARÃES, 2010; HANEBUTH et al., 2011; MILLI et al., 2013).

Estudos recentes sobre o Quaternário tardio têm mostrado que períodos de aceleração e desaceleração do nível do mar, durante o último ciclo completo de queda e posterior subida do nível de base, foram o fator determinante no controle da geometria externa e das características internas dos ambientes deltaicos (POREBSKI; STEEL, 2006). Após o período conhecido como Último Máximo Glacial (UMG), quando a plataforma continental esteve inteiramente exposta e o nível do mar situava-se cerca de 120 m abaixo do nível do mar atual (HORI; SAITO, 2007; TAMURA et al., 2009; HANEBUTH et al., 2011; MILLI et al., 2013), a subida do nível do mar, desde então, exibindo taxas variáveis, permitiu a formação de diferentes tipos de deltas à medida que a plataforma continental era inundada.

Durante o estágio de subida do nível do mar é necessário um grande fluxo de sedimento para manter a progradação de um delta, devido à necessidade de preenchimento de um espaço de acomodação cada vez maior. A falta de um aporte sedimentar adequado durante a subida do nível relativo do mar resulta em um inevitável recuo da linha de costa, mecanismo conhecido como *auto-retreat* (HOLZ, 2012). Por outro lado, uma desaceleração das taxas de subida do nível relativo do mar, interrompe a criação do espaço de acomodação, permitindo que o delta avance por toda a largura da plataforma (regressão normal).

Um ciclo completo de queda e posterior subida do nível do mar pode favorecer a formação de uma família de deltas que transita ao longo da plataforma, desde a plataforma mais interna até a plataforma mais externa, intitulados deltas de cabeceira de baía (*bayhead deltas*), deltas de plataforma interna (*inner-shelf deltas*), deltas de plataforma média (*mid-shelf deltas*) e deltas de quebra de plataforma (*margin-shelf deltas*) (POREBSKI; STEEL, 2003; 2006).

Os deltas de cabeceira de baía (*bayhead deltas*) são deltas confinados às reentrâncias costeiras (estuários), dominados principalmente por processos fluviais, podendo sofrer influência ainda das marés. São representados por uma clinoforma ainda pouco desenvolvida, com poucos metros de espessura, formada durante o estágio inicial de subida do nível do mar (trato de sistemas de nível baixo). À medida que o nível do mar sobe até alcançar o nível de mar alto (trato de sistemas de nível alto), o aumento do espaço de acomodação permite a formação de depósitos de cordões litorâneos com algumas dezenas de metros de espessura, dispostos paralelamente à costa, denominados deltas de plataforma interna (*inner-shelf deltas*). Os deltas de plataforma interna são responsáveis pela formação das planícies deltaicas, originando uma clinoforma sigmoidal extensa de baixa amplitude, tendendo a um formato em “rabo de cavalo”, característica distintiva desse tipo de delta (POREBSKI; STEEL, 2003; 2006).

Os deltas de plataforma média (*mid-shelf deltas*) são formados durante o estágio inicial de queda do nível do mar (trato de sistemas de regressão forçada), quando a diminuição do espaço de acomodação é muito acentuada, a ponto de não existir espaço suficiente para a deposição de sedimentos, predominando a erosão. Isso permite a formação de uma clinoforma com baixo ângulo de mergulho, que tende a avançar sobre a frente deltaica (POREBSKI; STEEL, 2003; 2006).

Por sua vez, os deltas de quebra de plataforma (*margin-shelf deltas*) são formados durante o estágio final de queda do nível do mar (trato de sistemas de regressão forçada) e o estágio de nível de mar baixo (trato de sistemas de nível baixo), quando o nível do mar está localizado em níveis mais baixos, próximo ou abaixo da quebra da plataforma (POREBSKI; STEEL, 2006). Nessa situação a alta declividade do talude favorece a formação de depósitos de leques submarinos, influenciados por processos gravitacionais, como correntes de turbidez, deslizamentos e fluxos de detritos (SUMMERHAYES et al., 1978), gerando uma clinoforma com grande espessura e comprimento, podendo apresentar uma espessa sucessão de turbiditos arenosos na região da frente deltaica (POREBSKI; STEEL, 2006).

Porebski e Steel (2006) sugerem que a classificação e o reconhecimento dessa família de deltas de plataforma são importantes porque enfatiza o sistema de energia mista ao invés das categorias convencionais de energia dos membros finais, colocando os deltas dentro de um contexto estratigráfico mais dinâmico. Contudo, essa classificação não deve ser tomada como uma alternativa à classificação convencional de Galloway (1975), baseada inteiramente sob o ponto de vista dos processos intrabacinais. Deve haver uma integração entre as duas classificações de modo a permitir uma melhor compreensão sobre os processos que regem a formação e manutenção dos sistemas deltaicos.

Durante o Quaternário as variações do nível do mar em conjunto com o aporte sedimentar dos rios desempenharam um importante papel no desenvolvimento de ambientes fluviais, estuarinos e deltaicos nos vales dos grandes rios do mundo. Como já mencionado a construção dos sistemas deltaicos é controlada por fenômenos tais como tectônica, isostasia, clima, aporte de sedimentos, características da bacia de drenagem, geometria da bacia receptora, marés e correntes costeiras, além de outros fatores que variam consideravelmente de região para região

(COLEMAN; WRIGHT, 1975). Entretanto, todos estes fatores combinados não poderiam ter provocado uma iniciação quase que simultânea da maioria dos deltas do mundo durante o Holoceno. Deste modo, tem-se reconhecido as alterações nas taxas de subida do nível do mar como o principal fator no desenvolvimento precoce dos deltas holocênicos em nível mundial, particularmente no estudo de sequências estratigráficas baseado em registros sísmicos ao longo das margens continentais (STANLEY; WARNE, 1994; LIU et al., 2004; HANEBUTH et al., 2011; MILLI et al., 2013).

Com base em uma compilação de datações pelo método do radiocarbono em sedimentos deltaicos de diferentes regiões do mundo, Stanley e Warne (1994) mostraram que a construção da maioria dos deltas ao redor do mundo iniciou por volta de 8,5 - 6,5 ka AP, como resultado da desaceleração nas taxas de subida do nível do mar que ocorreu por volta deste período, o que aliado ao grande aporte de sedimentos pelos grandes rios do mundo, permitiu que o espaço de acomodação criado pela subida do nível do mar fosse rapidamente preenchido e ultrapassado. Estudos recentes, suportados por testemunhagens, registros sísmicos e datações pelo método do radiocarbono (HORI; SAITO, 2002; TA et al., 2002; LIU et al., 2004; TÖRNQVIST et al., 2004; TANABE et al., 2006; TAMURA et al., 2009; HANEBUTH et al., 2011; MILLI et al., 2013), confirmaram que a construção dos grandes deltas asiáticos teve início segundo o modelo de evolução deltaica inicialmente proposto por Stanley e Warne (1994).

As variações do nível do mar durante o Quaternário tardio foram essencialmente controladas pela dinâmica das grandes camadas de gelo do Hemisfério Norte. Diversos autores têm mostrado que a subida do nível do mar desde o UMG não foi uniforme, mas caracterizada por intervalos de subida rápida (pulsos de degelo), intercalados por períodos com taxas de subida mais reduzidas (FAIRBANKS, 1989; 1990; LIU et al., 2004; BARD et al., 2010; HANEBUTH et al., 2011; DESCHAMPS et al., 2012; ABDUL et al., 2016).

De maneira geral, as curvas de variação do nível do mar desde o UMG, construídas para diferentes regiões do mundo, mostram dois períodos principais de subida rápida do nível do mar, conhecidos como Pulsos de Degelo MWP-1A e MWP-1B, por volta de aproximadamente 14,6 e 11,4 ka AP, respectivamente (FAIRBANKS, 1989; 1990; LIU et al., 2004; BARD et al., 2010; HANEBUTH et al., 2011; DESCHAMPS et al., 2012; ABDUL et al., 2016). O período e a magnitude desses pulsos ainda são objeto de controvérsias, devido a existência de hiatos estratigráficos que cobrem, principalmente, o intervalo de tempo entre 14 e 9 ka AP (BARD et al., 2010).

É consensual entre estes autores que a subida do nível do mar iniciou de maneira muito lenta a partir de aproximadamente 18 ka AP, como resposta ao derretimento das primeiras camadas de gelo do Hemisfério Norte, não tendo alcançado mais do que 5 metros de amplitude no decorrer de aproximadamente 3 ka (FAIRBANKS, 1989; 1990; BARD et al., 2010; DESCHAMPS et al., 2012; ABDUL et al., 2016). Somente a partir de 16,1 ka AP, a subida do nível do mar experimentou uma moderada aceleração tendo subido cerca de 14 metros, no período entre 16,1 a 14,6 ka AP (DESCHAMPS et al., 2012). Todavia, estes valores ainda não estão bem estabelecidos na literatura. Em seguida, durante o Pulso de Degelo 1A (MWP-1A), entre 14,6 – 14,3 ka AP, o nível do mar subiu cerca de 18 metros, representando uma taxa média de subida que pode ter alcançado 20 mm/ano (DESCHAMPS et al., 2012; ABDUL et al., 2016).

Após o MWP-1A, as taxas de subida do nível do mar diminuíram devido a um resfriamento rápido do planeta, que culminou no período frio conhecido como *Younger Dryas* (YD). Este período corresponde ao intervalo de tempo entre 13,9 a 11,7 ka AP (BARD et al., 2010; ABDUL et al., 2016). A teoria atualmente dominante diz que o YD foi causado por uma perturbação da circulação atmosférica que alterou a circulação circumpolar do Atlântico Norte (ABDUL et al., 2016). A abrupta liberação de grandes volumes de água doce represada em grandes lagos na América do Norte reduziu a salinidade e densidade de suas águas superficiais, resultando numa redução da circulação termohalina e consequentemente num menor transporte de calor do Atlântico Sul para o Atlântico Norte (ABDUL et al., 2016). Como resultado, ocorreu um resfriamento, e as taxas de subida do nível do mar diminuíram paulatinamente de 20 mm/ano para 7 mm/ano, chegando a alcançar uma taxa mínima de 4 mm/ano ao final do YD (BARD et al., 2010; DESCHAMPS et al., 2012; ABDUL et al., 2016). Ainda existem controvérsias se o YD foi um evento de caráter global, e sobre quais mecanismos poderiam ter causado um resfriamento tão catastrófico.

O segundo pulso de subida do nível do mar ocorreu entre o intervalo de aproximadamente 11,4 a 11,1 ka AP, como uma resposta direta, apesar de tardia, à deglaciação das grandes camadas de gelo do Hemisfério Norte, que marca o final do YD e início do período Holoceno (FAIRBANKS, 1989; 1990; LIU et al., 2004; BARD et al., 2010; HANEBUTH et al., 2011; DESCHAMPS et al., 2012; ABDUL et al., 2016). Embora alguns autores reconheçam a existência de um segundo pulso de degelo (MWP-1B) após o *Younger Dryas* (FAIRBANKS, 1989; 1990; LIU et al., 2004; BARD et al., 2010; HANEBUTH et al., 2011; DESCHAMPS et al., 2012; ABDUL et al., 2016), este segundo pulso foi confirmado somente na região de Barbados, onde aparece como um proeminente salto de aproximadamente 15 metros de magnitude na curva de variação do nível do mar, representando uma taxa de subida de 40 mm/ano. Por outro lado, os dados da região do Tahiti sugerem que este segundo pulso (MWP-1B) seria em realidade apenas uma aceleração nas taxas de subida do nível do mar, de caráter menos marcante que um pulso de degelo, não podendo assim ser considerado como tal (BARD et al., 2010).

Liu et al. (2004) com base na compilação de dados produzidos por vários autores propõem ainda a existência de mais dois pulsos de degelo (MWP-1C e MWP-1D), ocorridos em aproximadamente 9 e 7,5 ka AP, respectivamente. Entretanto, a falta de um consenso sobre a existência destes pulsos tem gerado uma grande controvérsia, de modo que, a existência ou não destes pulsos tardios é uma questão ainda em aberto e que precisa ser mais bem investigada. Apesar das controvérsias ainda existentes, permanece um padrão geral de comportamento do nível do mar desde a última glaciação, que exerceu um controle marcante na deposição e preservação dos estratos sedimentares durante a evolução dos deltas holocênicos (LIU et al., 2004; BIRD et al., 2007; TAMURA et al., 2009; HANEBUTH et al., 2011).

REFERÊNCIAS

ABDUL, N. A., R. A. MORTLOCK, J. D. WRIGHT, AND R. G. FAIRBANKS (2016) Younger Dryas sea level and meltwater pulse 1B recorded in Barbados reef crest coral Acropora palmata, Paleoceanography, 31, 330–344. doi:10.1002/2015PA002847.

AMOSORI, A., BRUNO, L., CAMPO, B.; MORELLI, A.; ROSSI, V.; SCARPONI, D.; HONG, W.; BOHACS, K. M.; DREXLER, T. M., 2017. Global sea-level control on local parasequence architecture from the Holocene record of the Po Plain, Italy. Marine and Petroleum Geology, 87, 99-111.

ANTHONY, E. J., 2015. Wave influence in the construction, shaping and destruction of river deltas: A review. Marine Geology. 361:53-78.

ARAÚJO, H. A. B.; DOMINGUEZ, J. M. L.; MACHADO, A. J.; RANGEL, A. G. A. N., 2018. Benthic foraminifera distribution in a deltaic clinoform (São Francisco Delta, eastern Brazil): A reference study. Journal of Marine Systems.

BANKS, K. W.; RIEGL, B. M.; SHINN, E. A.; PILLER, W. E.; DODGE, R. E., 2007. Geomorphology of the Southeast Florida continental reef tract (Miami-Dade, Broward, and Palm Beach Counties, USA). Coral Reefs, 26, 617–633.

BARD, E.; HAMELIN, B.; ARNOLD, M.; MONTAGGIONI, L.; CABIOCH, G.; FAURE, G.; ROUGERIE, F., 1996. Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. Nature 382, 241-244.

BARD, E.; B. HAMELIN.; DELANGHE-SABATIER, D., 2010. Deglacial Meltwater pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti. Science, 327(5970), 1235–1237.

BIRD, M. I.; FIFIELD, L. K.; TEH, T. S.; CHANG, C. H.; SHIRLAW, N.; LAMBECK, K., 2007. An inflection in the rate of early mid-Holocene eustatic sea-level rise: A new sea-level curve from Singapore. Estuarine, Coastal and Shelf Science, 71, 523-536

BLANCHON, P.; SHAW, J., 1995. Reef drowning during the last deglaciation: evidence for catastrophic sea-level rise and icesheet collapse. Geology, 23, 4–8.

BLANCHON, P.; JONES, B.; FORD, D. C., 2002. Discovery of a submerged relic reef and shoreline off Grand Cayman: further support for an early Holocene jump in sea level. Sedimentary Geology, 147, 253–270.

BOSKI, T.; BEZERRA, F. H. R.; PEREIRA, L. F.; SOUZA, A. M.; MAIA, R. M.; LIMA-FILHO, F. P., 2015. Sea-level rise since 8.2 ka recorded in the sediments of the Potengi-Jundiai Estuary, NE Brazil. Marine Geology, 365, 1-13.

CAINELLI, C., 1992. Sequence stratigraphy, canyons and gravity mass-flow deposits in the Piaçabuçu Formation, Sergipe-Alagoas Basin, Brazil: PhD Thesis, The University of Texas at Austin, 234p.

CAINELLI, C., 1994. Shelf processes and canyon/channel evolution controlling turbidite systems: examples from the Sergipe-Alagoas Basin, Brazil. GCSSEPM Foundation 15th Annual Research Conference, 39–50.

CBHSF, 2009. A bacia hidrográfica do rio São Francisco. Disponível em: http://www.saofrancisco.cbh.gov.br/

DESCHAMPS, P.; DURAND, N.; BARD, E.; HAMELIN, B.; CAMOIN, G.; THOMAS, A. L.; HENDERSON, G. M.; OKUNO, J.; YOKOYAMA, Y., 2012. Ice-sheet collapse and sea-level rise at the Bølling warming 14,600 years ago. Nature, 483(7391), 559–564.

DOMINGUEZ, J. M. L.; SILVA, R. P.; NUNES, A. S.; FREIRE, A. F. M., 2013. The narrow, shallow, low-accommodation shelf of central Brazil: Sedimentology, evolution, and human uses. Geomorphology.

FAIRBANKS, R. G., 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature 342: 637- 642.

FONTES, L. C.; KOWSMANN, R. O.; PUGA-BERNABÉU, A., (Editores) 2017. Geologia e geomorfologia na bacia de Sergipe-Alagoas. Editora Universidade Federal de Sergipe.

GILBERT, G.K., 1885. The topographic features of lake shores: U.S. Geological Survey, 5th Annual Report (1883–1884), p. 69–123.

GUIMARÃES, J. K., 2010. Evolução do delta do rio São Francisco – estratigrafia do Quaternário e relações morfodinâmicas. PhD Thesis, Federal University of Bahia.

HANEBUTH, T. J. J.; VORIS, H. K.; YOKOYAMA, Y.; SAITO, Y.; OKUNO, J., 2011. Formation and fate of sedimentary depocentres on Southeast Asia's Sunda Shelf over the past sea-level cycle and biogeographic implications. Earth-Science Reviews 104: 92-110.

HARRIS, P. T.; HEAP, A. D.; MARSHALL, J. F.; MCCULLOCH, M., 2008. A new coral reef province in the Gulf of Carpentaria, Australia: colonisation, growth and submergence during the early Holocene. Marine Geology, 251, 85-97.

HARRISON, S.; SMITH, D. E.; GLASSER, N. F., 2019. Late Quaternary meltwater pulses and sea level change. Journal of Quaternary Science. 34:1-15.

HERNÁNDEZ-MOLINA, F. J.; SOMOZA, L.; REY, J.; POMAR, L., 1994. Late Pleistocene-Holocene sediments on the Spanish continental shelves: Model for very high resolution sequence stratigraphy. Marine Geology, 120, 129-174.

HERNÁNDEZ-MOLINA, F. J.; FERNÁNDEZ-SALAS, L. M.; LOBO, F.; SOMOZA, L.; DÍAZ-DEL-RIO, V.; ALVEIRINHO DIAS, J. M., 2000. The infralitoral prograding wedge: a new large-scale progradational sedimentary body in shallow marine environments. Geo-Marine Letters, 20, 109-117.

HORI, K.; SAITO, Y., 2007. Classification, Architecture and Evolution of Large-river Deltas. Large Rivers: Geomorphology and Management.

LIU, J. P.; MILLIMAN, J. D.; GAO, S.; CHENG, PENG., 2004. Holocene development of the Yellow River's subaqueous delta, North Yellow Sea. Marine Geology 209: 45-67.

MEDEIROS, P.R.P.; KNOPPERS, B.A.; DOS SANTOS JUNIOR, R.C.; DE SOUZA, W.F.L., 2007. Aporte fluvial e dispersão de matéria particulada em suspensão na zona costeira do rio São Francisco (SE/AL). Geochimica Brasiliensis 21(2), 212-231.

MILLI, S.; D'AMBROGI, C.; BELLOTI, P.; CALDERONI, G.; CARBONI, M. G.; CELANT, A.; DI BELLA, L.; DI RITA, F.; FREZZA, V.; MAGRI, D.; PICHEZZI, R. M.; RICCI, V., 2013. The trasition from wave-dominated estuary to wave-dominated delta: The Late Quaternary stratigraphic architecture of Tiber River deltaic succession (Italy). Sedimentary Geology, 284-285: 159-180.

MUHS, D. R.; PRENTICE, C.; MERRITTS, D. J., 2003. Marine terraces, sea level history and Quaternary tectonics of the San Andreas fault on the coast of California: in Easterbrook, D., ed., Quaternary Geology of the United States, INQUA 2003 Field Guide Volume, Desert Research Institute, Reno, NV, p. 1-18.

MUTO, T.; STEEL, R. J., 1997. Principles of regression and transgression: the nature of the interplay between accommodation and sediment supply. Journal of Sedimentary Research, 994-1000.

NASCIMENTO, A.A., 2017. Estrutura da macrofauna bentônica de fundo inconsolidado na plataforma continental de Sergipe, Nordeste do Brasil (PhD dissertation) Federal University of Bahia, Brazil, (117p.).

PATRUNO, S.; HAMPSON, G. J.; JACKSON, C. A. L., 2015. Quantitative characterisation of deltaic and subaqueous clinoforms. Earth-Science Reviews. 142:79-119.

SOUZA-LIMA, W., ANDRADE, E. DE J., BENGTSON, P., GALM, P.C., 2002. A bacia de Sergipe Alagoas, evolução geológica, estratigrafia e conteúdo fóssil. Fundação Paleontológica Phoenix, Aracaju, 34 pp.

STANLEY, D.J.; WARNE, A.G., 1994. Worldwide initiation of Holocene marine deltas by deceleration of sea- level rise. Science 265, 228-231.

SUMMERHAYES, C. P.; FAINSTEIN, R.; ELLIS, J. P., 1976. Continental margin off Sergipe and Alagoas, northeastern Brazil: A reconnaissance geophysical study of morphology and structure. Marine Geology, 20, 345-361.

TA, T.K.O.; NGUYEN, V.L.; TATEISHI, M.; KOBAYASHI, I.; SAITO, Y.; NAKAMURA, T., 2002. Sediment facies and Late Holocene progradation of the Mekong River Delta in Bentre Province, southern Vietnam: an example of evolution from a tide-dominated to a tide- and wave-dominated delta. Sedimentary Geology 152: 313-325.

TAMURA, T.; SAITO, Y.; SIENG, S.; BEN, B.; KONG, M.; SIM, I.; CHOUP, S.; AKIBA, F., 2009. Initiation of the Mekong River delta at 8 ka: evidence from the sedimentary succession in the Cambodian lowland. Quaternary Science Reviews 28: 327-344.

TANABE, S.; SAITO, Y.; SATO, Y.; SUZUKI, Y.; SINSAKUL, S.; TIYAPAIRACH, S.; CHAIMANEE, N., 2003. Stratigraphy and Holocene evolution of the mud-dominated Chao Phraya delta, Thailand. Quaternary Science Reviews 22: 789-807.

TANABE, S.; SAITO, Y.; VU, Q. L.; HANEBUTH, T. J. J.; NGO, Q. N.; KITAMURA, A., 2006. Holocene evolution of the Song Hong (Red River) delta system, northern Vietnam. Sedimentary Geology 187: 29-61.

TÖRNQVIST, T.E.; GONZÁLEZ, J.L.; NEWSOM, L.A.; VAN DER BORG, K.; DE JONG, A.F.M.; KURNIK, C.W., 2004. Deciphering Holocene sea-level history on the U.S. Gulf Coast: a high- resolution record from the Mississippi Delta. Geological Society of America Bulletin 116: 1026–1039.

CAPÍTULO 2

ANTECEDENT TOPOGRAPHY CONTROLS PRESERVATION OF LATE QUATERNARY TRANSGRESSION RECORD AND CLINOFORM GEOMETRY: THE SÃO FRANCISCO DELTA (EAST-NORTHEASTERN BRAZIL)

Rangel, A.G.A.N., Dominguez, J.M.L.

ABSTRACT

The present study used shallow high-resolution seismic surveys to characterize the architecture and stratigraphic evolution of the last depositional sequence of the São Francisco Delta (SFDS), deposited during the late Pleistocene/Holocene. The SFDS accumulated in a bathymetric depression of the continental shelf associated with the São Francisco canyon head. The existence of this depression resulted in additional accommodation space measuring several tens of meters, which allowed the accumulation of a continuous stratigraphic record of eustatic sea-level rise since the Last Glacial Maximum on one of the shallowest and narrowest continental shelves in the world. Six major seismostratigraphic units were individualized. The lower units (Su1, Su2 and Su3) accumulated around the head of the canyon, limited by the walls of the bathymetric depression. Su1 was possibly drowned by MWP1A. Units Su2 and Su3 accumulated over the ensuing period of reduced rates of eustatic sea-level rise between the end of MWP1A and the Younger Dryas (YD). The top of Su3 transitions laterally to a wave-cut terrace engraved in the walls of the bathymetric depression. This terrace is located between 55 and 60 m below present sea level, and probably marks shoreline position during the YD. Su3 was drowned by MWP1B, as indicated by sediment wedges (Su6) that buried the top of this unit in the lateral portions of the bathymetric depression. Su4 was deposited still within a coastal embayment, during a period of low sea-level rise rates after MWP1B. Su4 was apparently drowned by MWP1C around 8.4 ka BP, marking the formation of the maximum flooding surface. The deposit of Su5, which corresponds to the most recent unit of the delta, only took place after the stabilization of eustatic sea level that began in 8-7.5 ka BP. This study

demonstrated how variations in eustatic sea-level rise rates combined with antecedent local morphology favored the creation of a continuous sedimentary record of Holocene transgression.

Keywords: wave-dominated delta; sea-level changes; seismic stratigraphy.

1 INTRODUCTION

Eustatic sea-level rise during the late Quaternary and the creation of accommodation space derived from this process hold fundamental control on the evolution of the large Holocene deltas, resulting in great similarities in their stratigraphic architecture (Ta et al. 2002; Tanabe et al. 2003, 2006; Liu et al. 2004; Törnqvist et al. 2004; Hori and Saito 2007; Tamura et al. 2009; Hanebuth et al. 2011; Milli et al. 2013; Anthony 2015; Patruno et al. 2015; Amorosi et al. 2017).

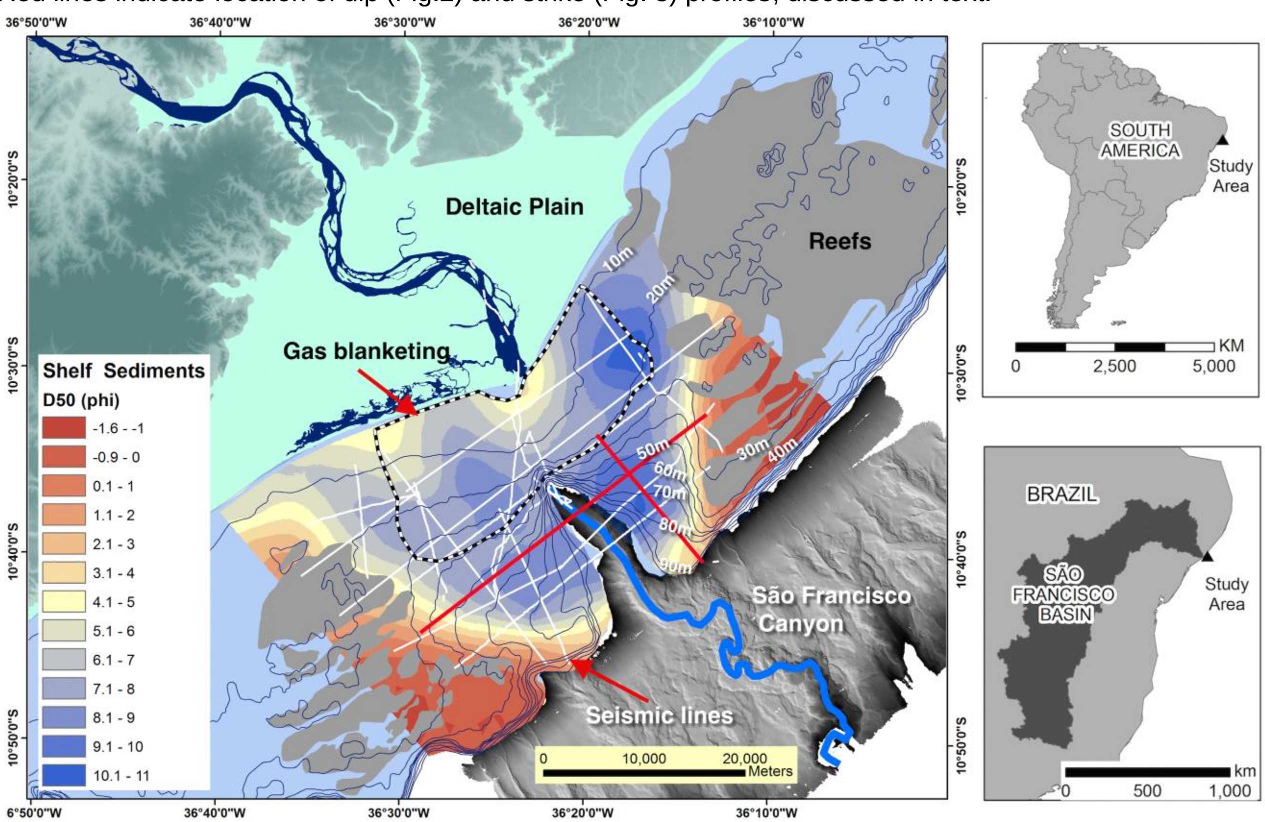
The rise in eustatic sea level since the Last Glacial Maximum (LGM) was marked by periods of accelerated rates (Meltwater pulses - MWP), alternated with periods of more reduced rates (e.g. Younger Dryas - YD) (Fairbanks 1989; Liu et al. 2004; Bird et al. 2007; Bard et al. 2010; Blanchon 2011; Hanebuth et al. 2011; Deschamps et al. 2012; Abdul et al. 2016; Harrison et al. 2019). During periods of acceleration, backstepping occurs in deltaic bodies if sedimentation rates cannot keep up with accommodation rates. In turn, during periods of deceleration or stillstand, sediment supply is usually sufficient to overcome the rates of sea-level rise, resulting in temporary progradation of transgressive packages (Hernández-Molina et al. 1994, 2000). Thus, the great deceleration of sea-level rise rates that occurred between 8.5 and 6.5 ka BP is considered the main factor responsible for triggering the construction of the great Holocene deltas around the world (Stanley and Warne 1994).

The shallowness (< 60 m) and narrowness (< 40 km) of the east-northeastern Brazilian continental shelf is a challenge to the investigation of effects of the last eustatic sea-level rise on the development of deltas located in this region. Moreover, high wave energy results in greater bottom shear stress, contributing to inhibit the development of deltaic clinoforms with classic geometry.

The wave-dominated delta of the São Francisco river presents very particular characteristics when compared to other Brazilian deltas (Figure 1). This delta was built on a topographically low region in the central portion of the continental shelf, associated with the head of the São Francisco canyon. This resulted in the creation of additional accommodation space of several tens of meters (Figure 1). In addition to favoring the development of 15 meters thick muddy clinoform associated with the current delta, this extra accommodation space also allows the investigation of how sea-level rise has influenced the architecture and stratigraphic evolution of the delta since the LGM.

The objective of the present study was to carry out a seismostratigraphic investigation on the deltaic clinoform of the São Francisco seeking to evaluate the role of variations in sea-level rise rates since the LGM on the development of the depositional architecture of the delta and its paleogeographic evolution.

Figure 1 - Location map of the study area showing: bathymetry, grain size of surface sediments, limits of area affected by gas blanketing, distribution of reef bottoms, and location of high resolution seismic lines. Red lines indicate location of dip (Fig.2) and strike (Fig. 3) profiles, discussed in text.



2 BACKGROUND INFORMATION

The headwaters of the São Francisco River are located at an altitude of 1,800 m and the river runs through 2,863 km until reaching the South Atlantic Ocean in northeastern Brazil. The river basin encompasses an area of 639,219 km^2, which corresponds to 8% of the national territory (CBHSF 2016) (Figure 1).

The mean flow rate and suspended sediment load at the mouth of the river was 3,010 m^3/s and 69 x 10^5 t/year, respectively, in the period before dams (1938-1973) (Medeiros et al. 2007). After a cascade of dams was built along the river, the values of mean flow rate and of suspended sediment load decreased to 1,760 m^3/s and 2.28 x 10^5 t/year, respectively (Medeiros et al. 2007).

The continental shelf adjacent to the delta integrates the Sergipe-Alagoas sedimentary basin, which was formed during the opening of the South Atlantic between the end of the Jurassic and beginning of the Cretaceous period (Souza-Lima et al. 2002). The mean width of the shelf is 30 km, with a 1:500 slope. The shelf break occurs abruptly with gradients of 7 to 8°, at a depth of approximately 50 m (Summerhayes et al. 1976; Cainelli 1994) (Figure 1).

The São Francisco canyon indents the continental shelf and is located only 8 km away from the river mouth, with a slope of 1:300 and noticeable influence from the isobath of 20 m onwards (Summerhayes et al. 1976). There is an area around the canyon head that is topographically low with abrupt lateral limits over which the delta has been built. Maximum depths of up to 670 m below sea level are recorded at the canyon head.

The continental shelf currently presents mixed siliciclastic-carbonate sedimentation, with siliciclastic sediments concentrated near the shoreline and in the deltaic clinoform. The areas to the side of the delta present sandy-gravelly carbonate sediments mainly composed of fragments and rhodoliths of crustose coralline algae associated with reef structures (Nascimento 2017; Fontes et al. 2017; Araújo et al. 2018) (Figure 1).

The deltaic plain of the São Francisco river covers an area of approximately 800 km^2 (Figure 1) and is formed by deposits of regressive shoreline sands (14m thick) and Pleistocene and Holocene dunes (Guimarães 2010). These sands prograde over the muddy clinoform covering an area of approximately 140 km^2, which extends from the isobath of 10 m to that of 80 m (Figure 1).

Since the Sergipe-Alagoas sedimentary basin is located in a passive margin at a mature stage, the region is considered tectonically stable, with very low subsidence rates from the coastal zone to the external continental shelf and slope (Cainelli 1992). Thus, the tectonic effects involved in creating accommodation space over the past 20 ka were assumed as insignificant.

3 MATERIALS AND METHODS

A total of 305 km of seismic records, including chirp (2 to 16 kHz) and sparker (0.3 to 1.5 kHz) acoustic sources, were used in the present study. This area covered the deltaic clinoform and the adjacent continental shelf (Figure 1).

Acoustic masking did not allow the visualization of subsurface features in the central portion of the clinoform (Figure 1). Thus, the data presented herein regards mainly the distal portions of the clinoform, both laterally and towards the shelf break.

4 RESULTS

4.1 Seismic stratigraphy

The depositional sequence in the subaqueous portion of the São Francisco delta (SFDS), which accumulated since the LGM, extends for about 22 km starting at the river mouth. Total thickness varies between 88 m (updip) and 27 m (downdip). The SFDS is delimited at its base by a subaerial unconformity and at the top by current seafloor. Six seismostratigraphic units were individualized (Figures 2 and 3). Moreover, a series of terraces carved in the lateral walls of the bathymetric depression where the SFDS accumulated were also mapped (Figures 2 and 3).

Figure 2 - Sparker dip seismic line oriented NW-SE showing major stratigraphic units discussed in text (see fig. 1 for location).

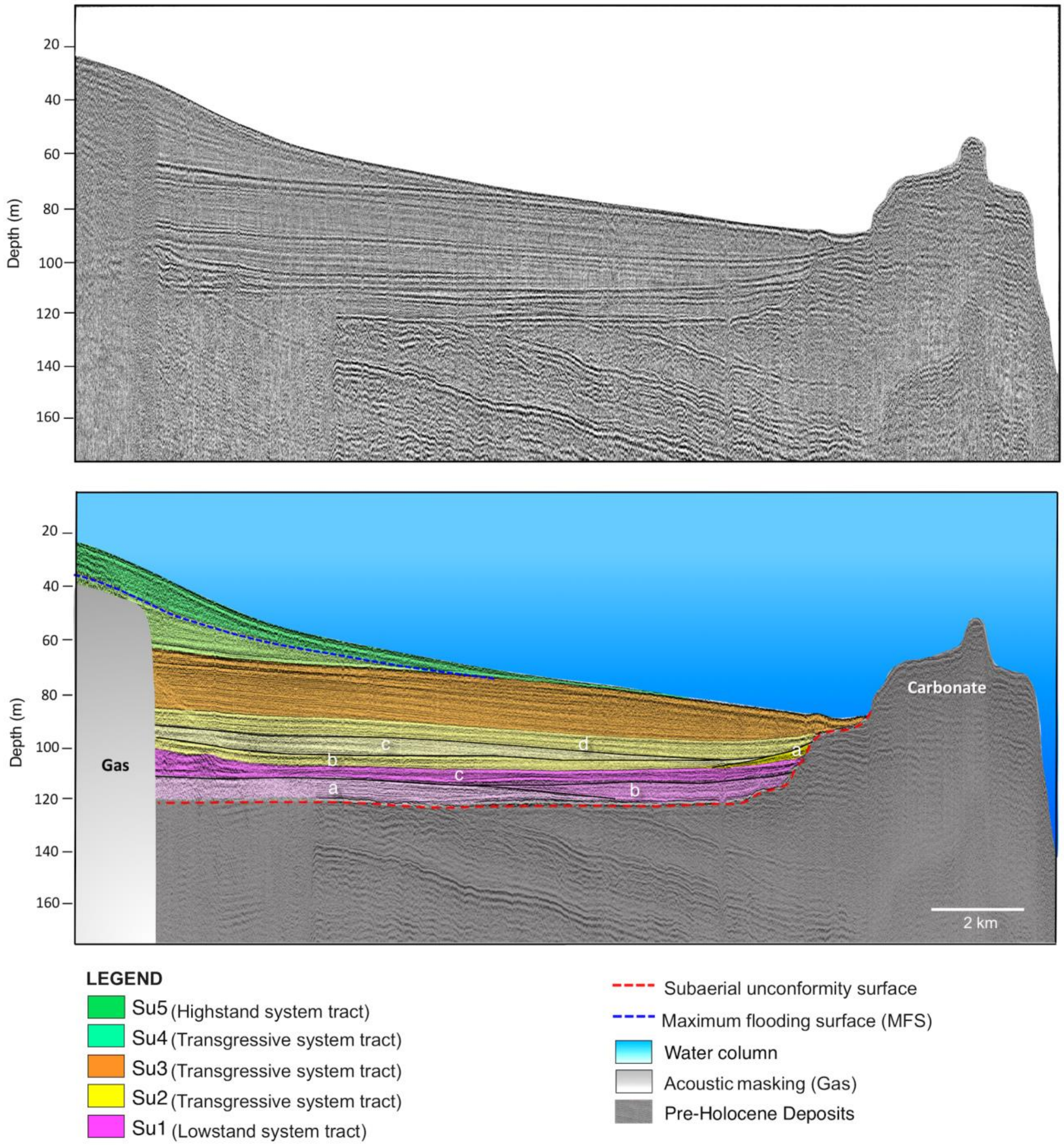
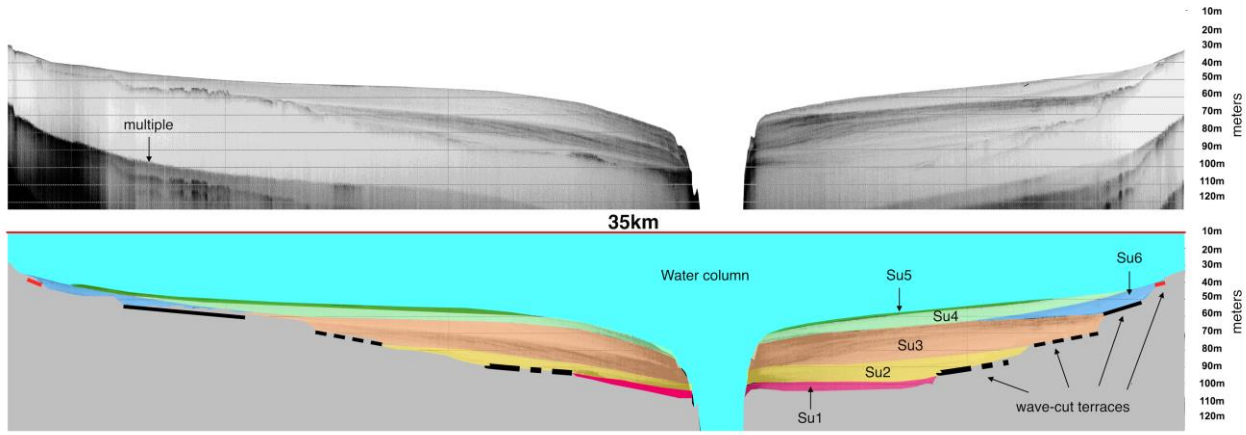


Figure 3 – Chirp strike seismic line oriented SW-NE showing major stratigraphic units discussed in text (see fig. 1 for location).



SEISMIC UNIT 1 (Su1) – this unit rests directly on the subaerial unconformity, restricting itself to the most distal portion of the delta, near the canyon head. The lower and upper limits of Su1 are located between 110 and 100 m below present sea level, respectively. The thickness of this unit is of the order of 10 m. Su1 can be internally subdivided into 3 smaller subunits (Su1a, Su1b and Su1c), whose thicknesses do not surpass 6 m (Figure 2). The two lower subunits (Su1a and Su1b) present a typically progradational geometry, growing from the margins of the bathymetric depression but with reduced lateral continuity. Subunit Su1a presents a downlap basal termination over the subaerial unconformity, while subunit Su1b onlaps subunit Su1a. From base to top, subunit Su1c presents a typically aggradational character, with large lateral continuity of its internal reflectors and onlap termination over the subaerial unconformity in the lateral walls of the bathymetric depression. Internal reflectors are continuous, regular, and considered parallel to subparallel.

SEISMIC UNIT 2 (Su2) – this unit is superimposed on the basal unit (Su1) and occurs fitted within the lateral walls of the canyon head, occupying a similar area to that of Su1. Its lower and upper limits are located between 100 and 80 m below present sea level, respectively, and the thickness of this unit does not surpass 20 m. This unit is composed internally of 4 subunits formed by high laterally continuous sedimentary bodies, with the exception of subunit Su2a (Figure 2). This subunit forms a wedge with maximum thickness of 3 m, with very limited lateral extension. The other subunits are formed by very low-angle thin clinoforms with large lateral continuity, with thicknesses that vary between 4 and 10 m. Subunits Su2b and Su2c show patterns of downlap termination over subunit Su2a. In turn, subunit Su2d, located at the top, exhibits a clinoform with great lateral continuity, slightly increasing in thickness offshore (downdip) and onlapping the subaerial unconformity in the lateral walls of the bathymetric depression (Figure 3). The internal reflectors of Su2 present upward concave geometry in the lower portion, transitioning to a progradational convex geometry towards the top. These internal reflectors are plane-parallel and exhibit great lateral continuity. This unit covers a well developed terrace carved in the lateral walls of the bathymetric depression at a depth varying from 85 to 90 m. Its top also terminates in another well-developed terrace at a depth of 80 m.

SEISMIC UNIT 3 (Su3) – this unit is also limited by the lateral walls of the topographic depression, but it extends over a relatively larger area of the embayment than that occupied by the underlying units (Su1 and Su2). The lower and upper limits of Su3 are located in its distal portion between 55 and 80 m below present sea level, respectively. The top of this unit reaches 55-60 m at the lateral portions of the clinoform (along-shelf) and passes to a prominent terrace carved in the wall of the topographic depression at this same depth (Figure 3). Thickness is of the order of 25 m, decreasing rapidly to only a few meters in the updip direction. This unit shows similar internal architecture as that of Su2. Concave reflectors at the basal portion become convex reflectors towards the top, exhibiting a more progradational geometry. This unit tends to be more seismically transparent at its base, with internal reflectors increasing in amplitude towards the top. Su3 presents a downlap termination over Su2 and an onlap over the subaerial unconformity on the lateral walls of the topographic depression.

SEISMIC UNIT 4 (Su4) – this unit consists of a well-developed sigmoidal clinoform, dipping 0.4°, extending from the river mouth to approximately half the width of the shelf. The lower and upper limits of this unit are located between 55 and 30 m, respectively. The thickness of Su4 does not surpass 25 m. The base of this unit is also seismically transparent, with increasing reflector amplitude towards the top. Its internal reflectors, when visible, are continuous and parallel.

SEISMIC UNIT 5 (Su5) – this is the most recent unit. Its top corresponds to the current seafloor. Similar to the underlying unit (Su4), Su5 also exhibits a pronounced sigmoidal clinoform dipping 0.5°. At its most distal portion, the unit's lower and upper limits are located between 30 and 20 m below present sea level, respectively. The internal reflectors of this unit are continuous, regular, parallel, and present high amplitude. Moreover, they present downlap termination over the maximum transgressive surface (MTS) that separates Su5 from the underlying unit (Su4). The most distal limit of this unit currently does not surpass the isobath of 80 m.

SEISMIC UNIT 6 (Su6) - besides the units described above there is a 6th unit, visible only in the strike profiles (Figure 3). The unit Su6, forms a wedge of sediments with very limited lateral continuity with an average thickness of 10m, deposited against the lateral walls of the bathymetric low. It rests on top of Su3 unit and it is partially or completely buried by unit Us4.

4.2 Terraces on the Lateral Walls

Terraces were identified carved on the lateral walls of the bathymetric depression that hosts the SFDS. These features are located at depths of 85-90m, 80-70m, 55-60m and 40m (Figure 3). The terraces are more noticeable on the northern wall of the bathymetric depression, which is also the steepest one. Moreover, these terraces end at the base of cliffs that measure between 3 to 10 m in height (Figure 3). The cliffs are sometimes associated with sediment wedges deposited against the depression. The most noteworthy of these wedges forms the Us6 unit mentioned above (Figure 3).

5 DISCUSSION

The SFDS shows a continuous record of deltaic sedimentation since the LGM. This unique situation, considering the shallow depths of the Brazilian east-northeastern continental shelf, was favored by the fact that this delta was built over a bathymetric depression associated with the head of the São Francisco canyon. This resulted in the creation of additional accommodation space. The existence of this depression also favored the trapping of fluvial sediments, especially mud, which has allowed the preservation of a complete stratigraphic history of this rise. Part of these sediments, however, has been lost to the São Francisco canyon.

Another important aspect is the existence of terraces carved in the lateral walls of the bathymetric depression. These features are possibly associated with either a temporary stabilization (stillstand) or a decrease in sea-level rise rates (Hernández-Molina et al. 1994). These terraces are similar to the wave-cut terraces present around lakes (e.g. Bonneville Lake, Utah) (Gilbert 1885) or in coastal regions that have suffered accentuated uplifting during the Quaternary (e.g. coast of California) (Muhs et al. 2003).

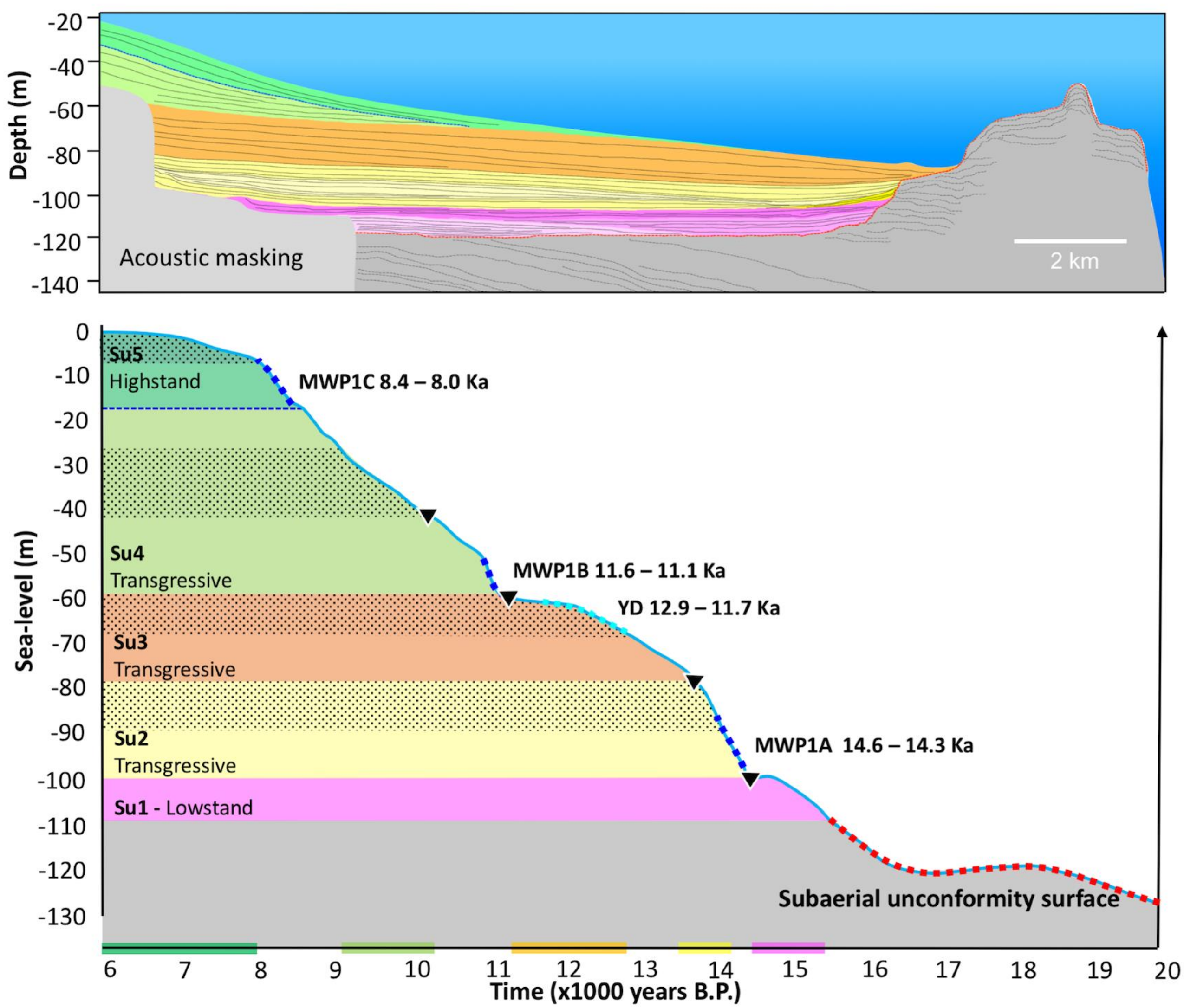
The basic assumption of the present study was that the depths at which these terraces occur, their lateral relationships with the individualized seismic units, and their integration with known eustatic sea-level curves since the LGM could help establish chronological limits for each seismostratigraphic unit identified in the sequence.

Moreover, the depths at which these terraces are located and their correlation with current bathymetry of the continental shelf adjacent to the bathymetric depression could contribute towards the paleogeographic reconstruction of the region. This is possible because the terraces represent the location of previous shorelines and accumulation rates of shelf carbonates that characterize this area are probably very low. Previous studies carried out on the eastern coast of Brazil on a section of the continental shelf with similar characteristics to that of the present study area have shown that in the last 10 ka, only 3-4 m of carbonate sediments accumulated in the outer region of the shelf (Dominguez et al., 2013).

However, establishing chronological limits based on seismostratigraphic surfaces is not entirely reliable. According to principles of sequence stratigraphy, the stratigraphic signature is nothing but the sedimentary record that resulted from the balance between sediment input and the accommodation space created by rates of sea-level variation (Muto and Steel, 1997). Thus, estimates of chronological limits are only reliable when using the lower limits of each seismostratigraphic unit, meaning when the process of sediment deposit began. It is therefore difficult to define the upper chronological limit, given that the paleo sea levels may have been much above the limit at the time of deposition.

Figure 4 illustrates the lower and upper limits of each seismic unit identified, the depths of the terraces engraved on the lateral walls of the depression, and the ice-volume global mean sea-level compiled by Harrison et al. (2019) which incorporates data from Liu et al. (2004), Bard et al. (2010), Deschamps et al. (2012) and Abdul et al. (2016). These curves were developed for locations that are distant from the great Quaternary ice sheets, regions that are therefore free from accentuated isostatic adjustments, as is the case with the São Francisco delta. Moreover, the tectonic stability of the region (Cainelli 1994) suggests that local effects related to tectonic subsidence can be ignored.

Figure 4– Upper and lower limits of the seismic units described in text plotted on the ice-volume global mean sea-level (Harrison et al. 2019). X axis color bars and stippled areas correspond to the time intervals during which deposition most probably took place. These intervals correspond to periods of reduced rates of sea-level rise, that followed the Melt Water Pulses (thick dark blue broken line). Black triangles indicate the average position of wave-cut terraces.



5.1 Evolutionary model

Seismic Unit 1, the first unit to be deposited, is restricted to a small area around the head of the São Francisco canyon. This area started being flooded when sea level surpassed 110 m below the current level, which according to the eustatic sea-level curves occurred around 15.2 ka BP. This unit, which consists of several amalgamated bodies of reduced lateral extension, was probably fed by the small water bodies that converged towards the topographic low, associated with the canyon head (Figure 5A). During the beginning of the flooding process there was no delta per se, and the sediments from the São Francisco river fed the canyon directly (Figure 5A). The upper limit of this unit, located 100 m below current sea level, coincides with the depth at the beginning of MWP1A, which took place around 14.6 ka BP. The deposition of Su1 terminated abruptly with MWP1A, which flooded this small recess in the shoreline.

With the end of MWP1A around 14.3 ka BP, eustatic sea-level rise rates decreased progressively, culminating in the YD with a near stabilization in sea level. Between the beginning of the MWP1A (14.3 ka BP) and the end of YD (11.7 ka BP), the eustatic level increased nearly 40 m, creating the accommodation space used for depositing units Su2 and Su3 (Figures 5B and C).

Su2 occurs over a slightly larger area than Su1, though it is still restricted to the head of the canyon (Figure 4B). When sea level surpassed 80 m, the flooded area at the head of the canyon increased considerably. This expansion was conditioned by the morphology of the head of the canyon and of the topographically low region around it. The flooding surface (downlap) that separated units Su3 from Su2 could have been caused by two factors. The first would be an acceleration in sea-level rise rates, for which there are still few evidences in other regions of the world. Bard et al. (1996) and Deschamps et al. (2012) suggest the occurrence of a rapid increase of 10 m between 13.8 and 13.4 ka BP for the region of Tahiti, right after MWP1A. The second cause would be the rapid broadening of the flooded area, which interrupted sedimentation given that a larger volume of sediments would be necessary to fill the accommodation space that has been created. The seismically transparent character of the base portion of unit Su3 and its transition towards the top to higher amplitude reflectors are indications of drowning followed by progradation.

Regarding its proximal portion, the top of Su3 laterally passes on to a marked terrace, carved in the walls of the topographic low, and visible mainly on the northeastern margin of this depression (Figure 3). This terrace is currently found at a depth of 55-60 m and was most likely carved by the action of waves during the near stabilization of sea level by the end of the YD, therefore marking the shoreline within the embayment that was becoming flooded. This geometrical arrangement suggests that the final accumulation of Su3 possibly coincides with the end of the YD (11.7 ka BP) (Figures 4 and 5C).

The accumulation of unit Su3 was abruptly interrupted by an acceleration in sea-level rise rates after the YD, during an event known as MWP1B (11.6-11.1 ka BP). At this time, sea-level rose around 10-15 m (Fairbanks 1989; Blanchon 2011). MWP1B was responsible for flooding an extensive area of the topographic depression, which led to a considerable retreat of the river mouth (Figure 5D). This rapid rise brought sea level to a position around -40 m, a depth in which another set of terraces, visible in the lateral walls of the bathymetric depression, occurs (Figure 3). A sediment wedge (Su6) was deposited associated with this terrace, covering the lateral margins of unit Su3, thus corroborating with the drowning of this unit (Figure 5D). These lateral wedges serve as a marker for the stratigraphic limit between units Su3 and Su4.

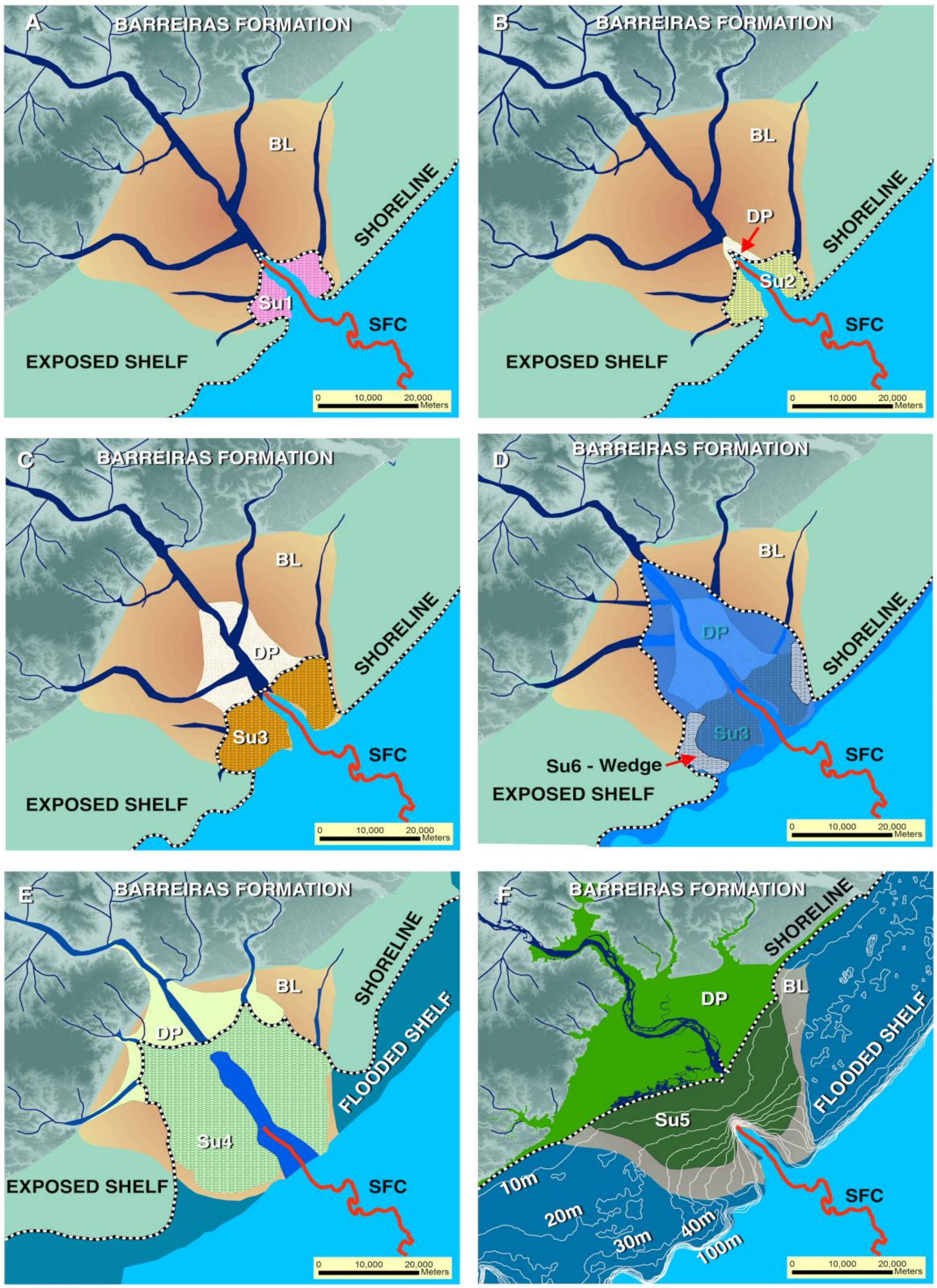
After MWP1B, the decrease in eustatic sea-level rise rates possibly favored the accumulation of unit Su4, which covered unit Su3 and the previously mentioned sediment wedges in the proximal region (Figure 5E). The top of unit Su4 reached 30 m below current sea level at the margins of the lateral walls of the topographic depression. Surveys conducted on the deltaic plain by Guimarães (2010) probably intercepted the top of this unit at nearly 20 m in depth associated with a bay-head delta that dates to between 8.8 and 8.4 ka BP.

The deposition of Su4 occurred still within the coastal embayment during the estuarine phase of the delta, given that the shelf at the lateral portions of the bathymetric depression had not been yet completely flooded (Figure 5E). However, this embayment occupied a much larger area than in previous stages.

Su4 was apparently drowned by another small eustatic sea-level rise pulse between 8.4 and 8.0 ka BP known as MWP1C. Although some authors recognize this period as only an acceleration of sea-level rise rates with less striking characteristics than a meltwater pulse, there are evidences in other regions of the world (Blanchon et al. 2002; Banks et al. 2007; Harris et al. 2008) and along the northeastern coast of Brazil of a rise of 6-7 m in sea level at a rate of 6.1 mm.year^{-1} (Boski et al. 2015). This period marks the formation of a maximum flooding surface that, in the deltaic plain, is located at a depth of 20 m and was dated at approximately 8.5-8.3 ka BP (Guimarães 2010). The depth at which the top of unit Su4 is found is compatible with the beginning of MWP1C, as shown in the curve proposed by Blanchon (2011) (based on curves by Fairbanks 1989; Blanchon and Shaw 1995), corroborating the idea that unit Su4 was drowned during MWP1C.

Only after eustatic sea-level stabilization began 8-7.5 ka BP, did the present delta began to prograde (unit Su5) (Figure 5F), as proposed by Guimarães (2010). The progradation of the current delta occurs in a scenario where the continental shelf is completely flooded, and the delta is exposed to high levels of energy. However, the topographic depression on the continental shelf where the delta is being built still presents a striking morphological expression in bathymetry, trapping fluvial sediments carried by the river and creating a significant accommodation space for the development of a muddy sigmoidal clinoform.

Figure 5 - Evolutionary model of the São Francisco delta. Su1, Su2, Su3, Su4, Su5 and Su6 are the seismic units discussed in text. DP – deltaic plain. BL – bathymetric low. SFC – São Francisco canyon. See text for details.



6 CONCLUSION

The last depositional sequence of the São Francisco river delta (SFDS) consists of six major seismostratigraphic units deposited since the LGM. Within the framework of stratigraphy sequences, unit Su1 comprises the lowstand systems tract (LST); units Su2, Su3 and Su4 comprise the transgressive systems tract (TST); and unit Su5 comprises the highstand systems tract (HST), when the current clinoform of this delta was formed after the stabilization of sea level around 8-7.5 ka BP. Unit Su6 forms a small wedge of sediments resting on the lateral walls of the bathymetric depression positioned between the units Su3 e Su4.

The depths of the abrasion terraces engraved in the lateral walls of the bathymetric depression, are compatible with the position of sea level during the deceleration or stillstand periods that occurred during post-LGM sea-level rise. This aided in the estimates of temporal limits for the individualized stratigraphic units.

The complete record of the sequence studied in the present investigation could only be preserved due to the very particular physiographic characteristics of the continental shelf adjacent to the delta, represented by a topographically low region associated with the head of the São Francisco canyon, which generated some tens of meters of additional accommodation space.

Similar to other deltas in the world, the formation of the present clinoform of the São Francisco delta (unit Su5) only began after the stabilization of sea level after 8-7.5 ka BP, when fluvial input was able to surpass the creation in accommodation space. Moreover, contrary to what has been published in the international literature, muddy clinoforms with classic sigmoidal geometry, which usually develop over extensive gentle low energy shelves (Patruno et al. 2015), can also develop in a shallow and narrow shelf with high wave energy due to particular aspects of submarine morphology.

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REFERÊNCIAS

ANTHONY, E. J., 2015. **Wave influence in the construction, shaping and destruction of river deltas: A review.** *Marine Geology.* 361:53-78.

BARD, E.; B. HAMELIN.; DELANGHE-SABATIER, D., 2010. **Deglacial Meltwater pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti**. *Science*, 327(5970), 1235–1237.

BHATTACHARYA, J. P., 2006. **Deltas.** In: Facies Models Revisited, Society for Sedimentary Geology, p. 237- 292.

BHATTACHARYA, J.P.; GIOSAN, L., 2003. **Wave-influenced deltas: geomorphological implications for facies reconstruction.** *Sedimentology* 50: p.187-210.

BITTENCOURT, A.C.S.P.; DOMINGUEZ, J.M.L.; FONTES, L.C.S.; SOUSA, D.L.; SILVA, I.R.; SILVA, F.R., 2007. **Wave refraction, river damming, and episodes of severe shoreline erosion: the São Francisco river mouth, northeastern Brazil.** *Journal of Coastal Research* 23 (4): 930-938.

BITTENCOURT, A.C.S.P.; DOMINGUEZ, J.M.L.; MARTIN, L.; FERREIRA, Y.A., 1982. **Dados preliminares sobre a evolução do delta do São Francisco (SE/AL) durante o Quaternário, influência das variações do nível do mar.** *Atas do IV Simpósio do Quaternário no Brasil*, p. 49-62.

BITTENCOURT, A.C.S.P.; MARTIN, L.; DOMINGUEZ, J.M.L.; SILVA, I.R.; SOUSA, D.L., 2002. A **significant longshore transport divergence zone at the northeastern Brazilian coast: implications on coastal Quaternary evolution**. *Anais da Academia Brasileira de Ciências* 74 (3): 505–518.

BROWN, S.; NICHOLLS, R. J., 2015. **Subsidence and human influences in mega deltas: The case of the Ganges-Brahmaputra-Meghna.** *Science of the Total Environment* 527-528:362-374.

CAINELLI, C., 1994. **Shelf processes and canyon/channel evolution controlling turbidite systems: examples from the Sergipe-Alagoas Basin, Brazil.** GCSSEPM Foundation 15th Annual Research Conference. 39–50.

CATUNEANU, O., 2002. **Sequence stratigraphy of clastic systems: concepts, merits, and pitfalls.** *Journal of African Earth Sciences (35):1-43.*

CATUNEANU, O., 2006. **Principles of sequence stratigraphy.** Editora Elsevier. The Netherlands, Amsterdam.

CATUNEANU, O., ABREU, V., BHATTACHARYA, J.P., BLUM, M.D., DALRYMPLE, R.W., ERIKSSON, P.G., FIELDING, C.R., FISHER, W.L., GALLOWAY, W.E., GIBLING, M.R., GILES, K.A., HOLBROOK, J.M., JORDAN, R., KENDALL, C.G.ST.C., MACURDA, B., MARTINSEN, O.J., MIALL, A.D., NEAL, J.E., NUMMEDAL, D., POMAR, L., POSAMENTIER, H.W., PRATT, B.R., SARG, J.F., SHANLEY, K.W., STEEL, R.J., STRASSER, A., TUCKER, M.E., WINKER, C., 2009. **Towards the standardization of sequence stratigraphy.** Earth-Science Reviews, (92):1-33.

CATUNEANU, O., GALLOWAY, W.E., KENDALL, C.G.ST.C., MIALL, A.D., POSAMENTIER, H.W., STRASSER, A., TUCKER, M.E., 2011. **Sequence stratigraphy: methodology and nomenclature**. Newsl. Stratigr. 44 (3):173-245.

CBHSF, 2009. *A bacia hidrográfica do rio São Francisco.* Disponível em: http://www.saofrancisco.cbh.gov.br/

COLEMAN, J.M.; WRIGHT, L.D., 1975. **Modern river deltas: variability of processes and sand bodies.** In: Broussard, M.L. (ed.), *Deltas, models for exploration*. Houston Geological Society, p. 99–149.

DESCHAMPS, P.; DURAND, N.; BARD, E.; HAMELIN, B.; CAMOIN, G.; THOMAS, A. L.; HENDERSON, G. M.; OKUNO, J.; YOKOYAMA, Y., 2012. **Ice-sheet collapse and sea-level rise at the Bølling warming 14,600 years ago**. *Nature,* 483(7391), 559–564.

DHN, 1989. *Carta Náutica n. 1000 Brasil – Costa Leste de Maceió ao rio Itariri*.

DOMINGUEZ, J. M. L, 1996. **The São Francisco strandplain: a paradigm for wave-dominated deltas?** *Geology of Siliciclastic Shelf Seas, Geological Society Special Publication* 117: 217-231.

DOMINGUEZ, J.M.L., 1986. **Evolução quaternária da planície costeira associada à fase do rio Jequitinhonha (BA).** In: Geologia e Recursos Minerais do Estado da Bahia, Secretaria de Minas e Energia, Superintendência de Geologia e Recursos Minerais, *Textos Básicos* 7: 81-121. Salvador.

DOMINGUEZ, J.M.L., 1987. **Quaternary sea-level changes and the depositional architecture of beach-ridge strandplains along the coast of Brazil.** Tese de Doutorado, University of Miami, Miami, Florida, USA.

DOMINGUEZ, J.M.L., 1990. **Deltas dominados por ondas: críticas às idéias atuais com referência particular ao modelo de Coleman & Wright.** Revista Brasileira de Geociências 20 (1-4): 352–361.

DOMINGUEZ, J.M.L.; BITTENCOURT, A.C.S.P.; MARTIN, L., 1981. **Esquema evolutivo da sedimentação quaternária nas feições deltaicas dos rios São Francisco (SE/AL), Jequitinhonha (BA), Doce (ES), e Paraíba do Sul (RJ).** Revista Brasileira de Geociências 11 (4): 227–237.

DOMINGUEZ, J.M.L.; BITTENCOURT, A.C.S.P.; MARTIN, L., 1983. **O papel da deriva litorânea de sedimentos arenosos na construção das planícies costeiras associadas às desembocaduras dos rios São Francisco (SE-AL), Jequitinhonha (BA), Doce (ES) e Paraíba do Sul (RJ).** Revista Brasileira de Geociências, (2): 98-105.

DOMINGUEZ, J.M.L.; BITTENCOURT, A.C.S.P.; MARTIN, L., 1992. **Controls on Quaternary coastal evolution of the east-northeastern coast of Brazil: roles of sea-level history, trade winds and climate.** Sedimentary Geology 80: 213–232.

FAIRBANKS, R. G., 1989. **A 17,000 year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation.** *Nature* 342: 637- 642.

FAIRBANKS, R.G., 1990. **The age and origin of the "Yonger Dryas Climate Event" in Greenland ice cores.** *Paleoceanography* 5 (6): 937-948.

FANGET, A. S.; BERNÉ, S.; JOUET, G.; BASSETTI, M. A.; DENNIELOU, B.; MAILLET, G. M.; TONDUT, M., 2014. **Impacto f relative sea level and rapid climate changes on the architecture and lithofacies of the Holocene Rhone subaqueous delta (western Mediterranean Sea).** *Sedimentary Geology 305: 35-53.*

FISHER, W. L., (1969). **Facies characterization of Gulf coast basin delta system, with some Holocene analogues.** Transactions of Gulf Coast Association of Geol. Soc., v. 19.

FOUFOULA-GEORGIOU, E., 2013. **A vision for a coordinated international effort on delta sustainability.** *In: Deltas: Lanforms, Ecosystems and Human Activities*, 358.

GALLOWAY, W. E., 1975. **Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional sustems.** In: Broussard, M. L. (ed.) *Deltas, models for exploration.* Houston Geological Society, Houston, p. 87-98.

GUIMARÃES, J. K, 2010. **Evolução do delta do rio São Francisco – estratigrafia do Quaternário e relações morfodinâmicas.** Tese de doutorado. Universidade Federal da Bahia, Instituto de Geociências.

HANEBUTH, T. J. J.; VORIS, H. K.; YOKOYAMA, Y.; SAITO, Y.; OKUNO, J., 2011. **Formation and fate of sedimentary depocentres on Southeast Asia's Sunda Shelf over the past sea-level cycle and biogeographic implications.** *Earth-Science Reviews* 104: 92-110.

HOLZ, M., 2012. **Estratigrafia de sequências: histórico, princípios e Aplicações.** *Editora Interciência, Rio de Janeiro. 212p.*

HORI, K., SAITO, Y., ZHAO, Q. AND WANG, P., 2002. **Architecture and evolution of the tide-dominated Changjiang (Yangtze) River delta, China.** *Sedimentary Geology*, 146, 249–264.

HORI, K.; SAITO, Y., 2007. **Classification, Architecture and Evolution of Large-river Deltas.** Large Rivers: *Geomorphology and Management*.

IBGE, 2009. *Banco de Dados – Cidades@.* Disponível em: http://www.ibge.gov.br/cidadesat/topwindow.htm?1

LIU, J. P.; MILLIMAN, J. D.; GAO, S.; CHENG, PENG., 2004. **Holocene development of the Yellow River's subaqueous delta, North Yellow Sea.** *Marine Geology 209: 45-67*MEDEIROS, P.R.P.; KNOPPERS, B.A.; DOS SANTOS JUNIOR, R.C.; DE SOUZA, W.F.L., 2007. **Aporte fluvial e dispersão de matéria particulada em suspensão na zona costeira do rio São Francisco (SE/AL).** *Geochimica Brasiliensis* 21 (2): 212-231.

MILLI, S.; D'AMBROGI, C.; BELLOTI, P.; CALDERONI, G.; CARBONI, M. G.; CELANT, A.; DI BELLA, L.; DI RITA, F.; FREZZA, V.; MAGRI, D.; PICHEZZI, R. M.; RICCI, V., 2013. **The trasition from wave-dominated estuary to wave-dominated delta: The Late Quaternary stratigraphic architecture of Tiber River deltaic succession (Italy).** *Sedimentary Geology* 284-285: 159-180.

PALAMENGHI, L.; SCHWENK, T,; SPIESS, V.; KUDRASS, H. R., 2011. **Seismostratigraphic analysis with centennial to decadal time resolution of the sediment sink in the Ganges–Brahmaputra subaqueous delta.** *Continental Shelf Research, 31, pp. 712–739*

PATRUNO, S.; HAMPSON, G. J.; JACKSON, C. A. L., 2015. **Quantitative characterisation of deltaic and subaqueous clinoforms.** *Earth-Science Reviews.* 142:79-119.

PIANCA, C.; MAZZINI, P. L. F.; SIEGLE, E., 2010. **Brazilian offshore wave climate based on NWW3 reanalysis.** *Brazilian Journal of Oceanography, 58(1):53-70.*

POREBSKI, S. J.; STEEL, R. J., 2003. **Shelf-margin deltas: their stratigraphic significance and relation to deep-water sands**. *Earth-Science Reviews*, v. 62, p. 283–326.

POREBSKI, S.J.; STEEL, R.J., 2006. **Deltas and sea-level change.** *Journal of*

Sedimentary Research 76: 390- 403.

RANGEL, A. G. A. N.; DOMINGUEZ, J. M. L., 2015. **Sedimentation and depositional architecture of the delta of the São Francisco River using high-resolution seismic data.** 14th International Congresso f the Brazilian Geophysical Society & EXPOGEF, Rio de Janeiro, Brazil, 3-6 August 2015: pp. 410-413.

SILVA, B.O., 2007. **Evolução, geometria e preenchimento do complexo de canyons de brejo grande, bacia de Sergipe-Alagoas.** Dissertação de Mestrado. Universidade Federal do Rio Grande do Sul, 88 pp.

SOUZA-LIMA, W., ANDRADE, E. DE J., BENGTSON, P., GALM, P.C., 2002. **A bacia de Sergipe Alagoas, evolução geológica, estratigrafia e conteúdo fóssil.** Fundação Paleontológica Phoenix, Aracaju, 34 pp.

SOUZA, W.F.L.; KNOPPERS, B., 2003. **Fluxos de água e sedimentos a costa leste do Brasil: relações entre a tipologia e as pressões antrópicas.** *Geochimica Brasiliensis* 17 (1): 57-74.

STANLEY, D.J.; WARNE, A.G., 1994. **Worldwide initiation of Holocene marine deltas by deceleration of sea- level rise.** *Science* 265: 228-231.

SUGUIO, K.; MARTIN, L.; BITTENCOURT, A.C.S.P.; DOMINGUEZ, J. M. L.; FLEXOR, J.-M.; AZEVEDO, A.E.G., 1985. **Flutuações do nível relativo do mar durante o Quaternário Superior ao longo do litoral brasileiro e suas implicações na sedimentação costeira.** Revista Brasileira de Geociências 15: 273-286.

SUMMERHAYES, C. P.; SESTINI, G.; MISDORP, R.; MARKS, N., 1978. **Nile delta: nature and evolution of continental shelf sediments.** *Marine Geology* 27: 43–65.

SYVITSKI, J. P. M; KETTNER, A. J.; OVEREEM, I; HUTTON, E. W. H.; HANNON, M. T.; BRAKENRIDGE, G. R.; DAY, J.; VOROSMARTY. C.; SAITO, Y.; GIOSAN, L.; NICHOLLS, R. J., 2009. **Siking deltas due to human activities.** *Nature Geoscience.*

TA, T.K.O.; NGUYEN, V.L.; TATEISHI, M.; KOBAYASHI, I.; SAITO, Y.; NAKAMURA, T.,

2002. **Sediment facies and Late Holocene progradation of the Mekong River Delta in Bentre Province, southern Vietnam: an example of evolution from a tide-dominated to a tide- and wave-dominated delta.** *Sedimentary Geology* 152: 313-325.

TAMURA, T.; SAITO, Y.; SIENG, S.; BEN, B.; KONG, M.; SIM, I.; CHOUP, S.; AKIBA, F., 2009. **Initiation of the Mekong River delta at 8 ka: evidence from the sedimentary succession in the Cambodian lowland.** *Quaternary Science Reviews* 28: 327-344.

TANABE, S.; SAITO, Y.; SATO, Y.; SUZUKI, Y.; SINSAKUL, S.; TIYAPAIRACH, S.; CHAIMANEE, N., 2003. **Stratigraphy and Holocene evolution of the mud-dominated Chao Phraya delta, Thailand.** *Quaternary Science Reviews* 22: 789-807.

TANABE, S.; SAITO, Y.; VU, Q. L.; HANEBUTH, T. J. J.; NGO, Q. N.; KITAMURA, A., 2006. **Holocene evolution of the Song Hong (Red River) delta system, northern Vietnam.** *Sedimentary Geology* 187: 29-61.

TÖRNQVIST, T.E.; GONZÁLEZ, J.L.; NEWSOM, L.A.; VAN DER BORG, K.; DE JONG, A.F.M.; KURNIK, C.W., 2004. **Deciphering Holocene sea-level history on the U.S. Gulf Coast: a high- resolution record from the Mississippi Delta.** *Geological Society of America Bulletin* 116: 1026–1039.

WRIGHT, L.D.; COLEMAN, J.M., 1973. **Variations in morphology of major river deltas as functions of ocean waves and river discharge regimes.** *The American Association of Petroleum Geologists Bulletin* 57 (2): 370-398.

CAPÍTULO 3
CONCLUSÕES

The last depositional sequence of the São Francisco river delta (SFDS) consists of five seismostratigraphic units deposited since the LGM. Within the framework of stratigraphy sequences, unit Su1 comprises the lowstand systems tract (LST); units Su2, Su3 and Su4 comprise the transgressive systems tract (TST); and unit Su5 comprises the highstand systems tract (HST), when the current clinoform of this delta was formed after the stabilization of sea level that began 8-7.5 ka BP.

The depths at which the abrasion terraces engraved in the walls of the bathymetric depression are found are compatible with eustatic paleo sea levels during the deceleration or stillstand periods that occurred during post-LGM sea-level rise. This aided in the estimates of temporal limits for the individualized units.

The complete record of the sequence studied in the present investigation could only be preserved due to the very particular physiographic characteristics of the continental shelf adjacent to the delta, represented by a topographically low region associated with the head of the São Francisco canyon, which generated some tens of meters of additional accommodation space.

Similar to other deltas in the world, the formation of the current clinoform of this delta (unit Su5) only began after the previously mentioned stabilization of sea level after 8-7.5 ka BP, when fluvial input was able to surpass the increase in accommodation space. Moreover, contrary to what has been published in the international literature, muddy clinoforms with classic sigmoidal geometry, which usually develop over extensive shelfs with smooth dips and low energy (Patruno et al. 2015), can also develop in a shallow and narrow shelf with high wave energy due to particular aspects of submarine morphology.