Geometry and continuity of deep-water sandstones and siltstones, Brushy Canyon Formation (Permian) Delaware Mountains, Texas

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Abstract
Sandstones and siltstones of the Permian (Guadalupian) Brushy Canyon Formation were deposited in 50 to more than 300 m of water in the Delaware Basin of west Texas. The transition from slope to basin-floor palaeoenvironments is exposed in a 24 km long transect of extensive outcrops that occur along the western flanks of the Guadalupe and Delaware mountains.

Several types of deep-water channels occur in the study area. Individual channels were filled with thick-beded sandstones that were deposited by high-density sediment gravity flows, thin-beded classical turbidites that were deposited by low-density turbidity currents, or siltstones that were deposited from suspension. Many channel-fills change facies laterally from amalgamated turbidite sandstones in channel axes to thin-beded turbidities and siltstones along channel margins. Cross-beded sandstones that occur as lag deposits at the bases of some channels are interpreted to have been deposited tractingly by high-density sediment gravity flows that carried most of their sediment loads further basinward. Siltstone drapes occur along the bases and margins of some channels, and are common in slope and proximal basin-floor palaeoenvironments and are interpreted to reflect channel abandonment.

Levee deposits are common in the Brushy Canyon Formation. Stratal geometries along many channel margins indicate aggradation of adjacent levee deposits during channel filling. Levee deposits commonly include laterally continuous, thin-beded classical turbidites that were deposited by low-density flows. However, near channel margins, levees can include thick-beded, amalgamated turbidites that were deposited by high- and low-density flows and cross-beded sandstones that were deposited tractingly by high-density flows. Along channel margins, palaeocurrents were commonly oriented parallel to slightly oblique to the channel trend and lenticular sandstone beds are common. In levee deposits, palaeocurrents generally were oriented 20°-60° away from the adjacent channel margin.

Introduction
Deep-water sandstones and siltstones of the Brushy Canyon Formation were deposited in the Delaware Basin of western Texas and southern New Mexico during Late Permian (Lower Guadalupian) time. Outcrops of Permian strata along the western escarpments of the Guadalupe and Delaware mountains provide an oblique cross-section of the transition from the shelf into the north-western part of the Delaware Basin during deposition of the Brushy Canyon Formation (Figs. 25.1 and 25.2). During the Late Permian, the Delaware Basin was a site of deposition of both siliciclastic and carbonate sediments. Shallow-water carbonates, evaporites and minor siliciclastics were deposited on the broad shelves that surrounded the basin; the deep-water basin was filled primarily with siliciclastics.

The Brushy Canyon Formation consists of basinally-restricted sandstones and siltstones that attain a maximum thickness of 360 m (Figs. 25.1, 25.3 and 25.4). At the basin margin the Brushy Canyon thins and onlaps underlying slope and shelf-margin carbonates of the Victorio Peak Dolomite, Bone Spring Limestone and Cottonwood Formation (King 1948). The onlap surface at the base of the Brushy Canyon Formation is an erosional unconformity that is interpreted to be of submarine origin (Harms and Pry 1974). The surface exhibits 410 m of relief and slopes of 5°-10° in the basin-margin area. This erosional surface is interpreted to correlate updip with a surface of subaerial exposure that occurs within carbonate bank strata of the San Andres Formation (Sarg and Lehmann 1980). The base of the Brushy Canyon Formation is interpreted to be a sequence boundary.

The Brushy Canyon Formation is interpreted to have been deposited during an overall lowstand of relative sea level (Silver and Todd 1969, Sarg and Lehmann 1986). Alternations of widespread siltstone- and sandstone-rich units within the Brushy Canyon Formation are interpreted to represent high-frequency depositional sequences.

Basinal deposits of the lower Brushy Canyon Formation are interpreted to have been deposited in 400-600 m of water, based on relief at the basin margin (410 m) and on clinof orm in the San Andres carbonate bank (150 m). Water depths at the basin margin probably decreased during Brushy Canyon time as a result of basin filling and may have been as little as 50-150 m in the basin margin area during deposition of the uppermost Brushy Canyon Formation (Sarg et al. 1988). Because the Brushy Canyon Formation onlaps the basin margin, coeval shelfal deposits have not been identified. The absence of clay and overall fine-grain size of sandstones in the Brushy Canyon Formation suggest eolian transport of sand and silt across the shelf (Fischer and Sarneith 1988).

Trends of submarine channels, ripple marks and cross-beds in the Brushy Canyon Formation indicate that the dominant direction of sediment transport in the Guadalupe Mountains area was to the south-east (King 1948; Harms 1974; Rossen 1985). This direction is normal to the trend of the basin margin in the Guadalupe Mountains area and is oblique to the trend of the Brushy Canyon outcrop belt (Figs. 25.1 and 25.3). In the Delaware Mountains the average direction of sediment transport in the Brushy Canyon was to the east. This direction is normal to the western margin of the basin and is oblique to the trend of the Brushy Canyon outcrop belt.

Lithofacies
Several lithofacies occur in the Brushy Canyon Formation. Rip-up clast conglomerates (Facies 1) occur throughout the study area and are common at the bases of channels (Fig. 25.21(a)). The conglomerates consist of pebble- to cobble-size siltstone rip-up clasts that are supported by a matrix of fine- or medium-grained sandstone that commonly contains abundant fusulinids or other skeletal carbonate grains. Rocks of this lithofacies typically weather reddish brown to medium orange. Mud-clast conglomerates are lenticular in geometry, commonly contain internal surfaces of erosion and are typically cross-beded. They are interpreted to be lag deposits that were deposited tractingly from the lower parts of high-density sediment gravity flows.

Massive and cross-beded sandstones (Facies 2A and 2B) are also common in channel-fills of the Brushy Canyon Formation. Massive sandstones lack internal sedimentary structures and range in thickness from 1 m to >6 m. Grain size in the massive and cross-beded sandstones ranges from lower fine to upper medium sand; the cross-beded sandstones commonly contain abundant fusulinids. Cross-beded sandstones are characterized by scoop-shaped erosional scours with fills that are well-to-poorly-laminated. In beds with well-defined laminations, laminae onlap one or both flanks of the trough or drape the trough. Aggradational or climbing dune forms occur in some channel-fills, including those at Popo Channel (Fig. 25.16A-(d)). Massive sandstones are interpreted to have been deposited rapidly from suspension within high-density flows. Cross-beded sandstones also are inferred to have been deposited from high-density flows but at more moderate rates of deposition (Lowe 1982).

Classical turbidites that exhibit one or more subdivisions of the Bouma sequence (Facies 3) occur in overbank deposits and channel-fills in the Brushy Canyon Formation. In non-amalgamated classical turbidities, thin, laminated siltstones are present between sandstone beds. In amalgamated classical turbidities, sandstone beds are absent. Thick-beded classical turbidites contain sandstone beds that are more than 0.3 m thick, are fine- to medium-grained and commonly consist of Tds, Tsb, Tdp, Tsc and Tsk beds. Thick-beded turbidites are interpreted to have been deposited from high-density (Tds subdivisions) and low-density (Tsb and Tsc subdivisions) flows (Lowe 1982). Thin-beded classical turbidites consist of very fine- to fine-grained sandstones that are <0.3 m thick.
They typically exhibit base-missing Bouma sequences (Tbs and Tsb beds) and are interpreted to have been deposited from low-density flows. Non-amalgamated, thin-beded turbidites occur throughout some channel-fills but more commonly occur along the margins of channels (Planar Crash Canyon, Fig. 25.14(c) and (f); Permeater Channel, Fig. 25.21) or in overbank areas (Buena Vista, Fig. 25.10(g) and (h)).

Current-rippled sandstones (Facies 4) commonly are fine- to very fine-grained and consist of thin, (1 cm to 0.3 m thick) beds of sandstone that are interbedded with laminated siltstones. The sandstone beds are laterally continuous (Facies 4A) or lenticular (Facies 4B). Lenticular, current-rippled sandstones in the Brushy Canyon Formation have been suggested to represent contourites (Muti 1992). However, this lithofacies commonly occurs in the Brushy Canyon Formation at the margins of channels in channel-fills and in overbank deposits (Figs 25.14(e) and 25.21(e)). The close association of this lithofacies with channel-margin and overbank processes suggests that the current-rippled sandstones were deposited by low-density turbidity currents.

Siltstones (Facies 5) in the Brushy Canyon Formation range from light grey to dark grey in colour (Fig. 25.10(a)) and in thickness as much as 25 m thick. They exhibit alternating thin (mm-scale) laminae of light-coloured, silt-rich and dark-coloured, organic-rich laminae (Fig. 25.10(a)). Dark-grey siltstones contain as much as 2 wt% of total organic carbon (Wenger, personal communication) and commonly include phylloclastic nodules and siderite or limestone concretions. Beds and laminae of siltstone characteristically drape or parallel underlying relief such as channel margins or the tops of current ripples (Fig. 25.18(1)) and occur as units as much as 25 m thick. They exhibit alternating thin laminae of stratified, fine- to medium-grained sandstone with abundant fusulind and other skeletal grains. The pebbly sandstones are interpreted to have been transported by debris flows and, where stratified, by small-scale, high-density sediment gravity flows. Medium-thick grey siltstones occur as thin (0.3 m thick) graded beds that commonly contain large, angular siltstone clasts (Fig. 25.19(c)). These beds occur in outcrops of the Brushy Canyon Formation at the base of the axial channels that they make up a small proportion of the strata. The siltstone beds are interpreted to represent slurry beds.

Channel types

A wide variety of channel morphologies occur in the Brushy Canyon Formation and the channels are filled with a variety of lithofacies. Basal channel surfaces range from simple, single erosional surfaces to multiple channel cuts that appear to be genetically related. Styles of channel-fill include beds that onlap the margins and bases of channels without lateral facies changes and folds that change facies laterally from channel axis to channel margin. Although some channels appear to have been cut and filled without the development of significant levees, stratigraphic relationships along the margins of many other channels indicate that adjacent levees aggraded during the filling of the channels.

The fills of many channels, especially in the basin margin area, are not genetically related to the processes that cut the channels. For example, in the interpreted palaeoarea between Bone Canyon North and Guadalupe Canyon, large-scale, erosional surfaces that have as much as 85 m of erosional relief which are directly overlain or draped by laminated beds of siltstone are common. Some of these erosional surfaces may represent abandoned channels that were draped with siltstone (Harms 1974). However, many of these erosional surfaces are associated with slumped or contorted siltstone units and are interpreted as slump scars.

In some sandstone-filled channels the vertical succession of lithofacies suggests that the processes that filled the channels are not genetically related to the processes that cut the channels. Channel-fills of this type occur at Permeater Channel (Figs 25.20 and 25.21), Delaware Mountain (Fig. 25.18(i)) and other areas where lenticular units of rip-up clast conglomerate or cross-beded, fusulinid-rich sandstone at the base of the channel are directly overlain by siltstones or sandstones that change facies from channel axis to channel margin. The basal beds of stratified rip-up clast conglomerate and fusulinid-rich sandstone typically contain numerous internal erosion surfaces and are interpreted to have been deposited by many flows. These relatively coarse-grained lithofacies are interpreted to be lag sands that were deposited tractively from the basal parts of high-energy sediment gravity flows that transported most of their sediment loads further basinward. These flows may be genetically related to the flows that cut the channels. Many basal channel lags are overlain by lower-energy deposits, such as siltstones, thin-beded turbidites or classical turbidities, that change facies completely Bouma sequences, e.g. facies containing the basal ‘a’ subdivision (Tbs, Tsb beds). Many of the siltstones and sands of the channel axis, to thin-beded Bouma sequences that lack the basal ‘a’ subdivision (Tbs, Tsb beds) along channel margins. In these cases, a significant period of time may have elapsed between channel erosion/lag formation and channel infilling when the channel had an bypass zone for sediment that was transported further into the basin.

Some of the channels in the Brushy Canyon Formation appear to have been cut and filled without the development of significant levees. However, stratigraphic relationships along the margin of many channels indicate that adjacent overbank strata aggraded during filling of the channels. These overbank sediments help to fill the cutoff channels within the channels and are interpreted to represent levees. For example, aggradation channel complexes with thick levee/overbank deposits occur in the middle and lower Brushy Canyon Formation at Pogo Channel (Figs 25.15 and 25.16), Plane Crash Canyon (Figs 25.13 and 25.14), Colleen Canyon (Figs 25.22 and 25.23) and other basal localities. The axial channels in these complexes lack a single, basal surface of erosion.

Instead, individual channel-fills are amalgamated in the axial parts of the channel complexes and split apart toward the margins of the complexes. Palaeocurrent measurements from the unconfined, overbank deposits indicate that flow was commonly oriented 20–60° oblique to the trend of the axial channel complex.

At Colleen Canyon, the most distal Brushy Canyon Formation outcrop studied to date, erosional relief at the channel margin was only 3–9 m. The outcrop at Colleen Canyon represents the transition between channelized areas that were dominated by erosional processes and non-channelized areas in which flows were relatively unconfined and depositional processes predominated.

**Geometry and continuity of sandstones and siltstones**

Within channel-fills, beds and bedsets of sandstone have a wide range of lateral continuity and vertical connectivity. Sandstones that occur within the axial parts of channel-fills commonly are laterally continuous within the fills (Figs 25.7, 25.8, 25.12, 25.14, 25.16, 25.19 and 25.23). In these axial areas where connectivity between sandstones is enhanced by erosion of siltstones and vertical amalgamation of sandstones. In contrast, isolated, lenticular sandstone bodies are common along the margins of channels with fills that exhibit lateral facies changes. Lenticular sandstones that represent erosional remnants are common along complex channel margins that exhibit multiple, successive periods of erosion and deposition. These features can be observed at the Guadalupe Canyon (Fig. 25.16(a)) and Buena Vista localities (Fig. 25.19(i)) and elsewhere in the upper Brushy Canyon Formation. Other lenticular sandbodies preserved at channel margins, such as lenticular beds containing ripple marks and larger bedforms, are predominantly the result of depositional processes and can be observed at channel margins north of Delaware Mountain (Fig. 25.18(i)) and elsewhere.

In overbank areas a high degree of lateral continuity can be demonstrated for many sandstone beds. For example, thin (0.3 m thick) sandstone beds were traced for scores of metres in overbank areas at Pogo Channel that were measured from scores of metres in overbank areas at Pogo Channel and thicker (1 m thick) sandstone beds extend for more than 300 m in the relatively unconfined deposits that occur to the east of the axial channels at Colleen Canyon (Fig. 25.25). Channels with limited