INTRODUCTION AND GEOLOGICAL SETTING

Deepwater slope channels are the significant sediment-delivery conduits from shallow shelf areas to deepwater basin floor, and the main reason for the sand-mud-sand grain-size partitioning, between shelf, slope and basin floor lithosomes, in the distal segment of the source-to-sink system. For large, shelf-margin scale systems, seismic data offers a holistic view but with only limited resolution, especially regarding vertical resolution.
on slope channels (e.g. Sylvester, Deptuck, Prather, Pirmez, & O’Byrne, 2012). High-resolution slope outcrops provide only fragmented views (e.g. Malkovsky, Jobe, Sharman, & Graham, 2018’s Magallanes clinoforms). Smaller systems allow a complete trace, but are mostly focused on the shelf topsets and basin floor fan deposits without the detailed description of the facies and architecture of the linkage, that is the slope channels, between the two (e.g. Lewis-Fox Hills (Carvajal & Steel, 2006), Spitsbergen Norway (Plink-Björklund, Mellere, & Steel, 2001)).

Here we present an outcrop study on Early Jurassic slope-channel facies and architectural variation across the length of the slope from clinoforms in Bey Malec Estancia, La Jardinera area in the southern Neuquén Basin, Argentina.

Neuquén Basin, with its 4-km-thick near continuous Late Triassic-Early Cenozoic stratigraphy covering 120,000 km² in Argentina and central Chile, is bounded by the Andes on the west, the Pampeano-Sierra Pintada massif to the northeast and the North Patagonian massif to the southeast (Howell 2005, Figure 1a). The evolution of the southern Neuquén Basin has been divided into three stages (Howell, Schwarz, Spalletti, & Veiga, 2005; Ramos, 1999a, 1999b; Vergani, Tankard, Belotti, & Welsink, 1995). (a) Late Triassic to Early Jurassic fault-controlled extension and subsidence that created isolated depocentres of pre-Cuyo strata on the western margin of Gondwana; (b) Early Jurassic to Early Cretaceous back-arc stage, where thermal sag gradually outtook fault-controlled subsidence by Middle Jurassic. Marine deposition initiated with merging of isolated depocentres by Pliensbachian (Vicente, 2005), and induced multiple transgressive-regressive cycles in the Cuyo, Lotena and Mendoza groups; (c) Late Cretaceous to Cenozoic transition to a shallowly dipping subduction zone associated with the Andean orogeny.

Onshore seismic data show a series of NW prograding slope clinoforms of the Cuyo Group in south-central Neuquén Basin (Brinkworth et al., 2017). Those clinoforms mapped NE of the field area flatten out downslope as deepwater bottomsets and eventually flatten also upslope as clinoform topsets that form shallow-marine deposits and their coeval proximal fluvial deposits. The basinward-migrating clinoforms are therefore composed of diachronous, broadly coeval facies belts of Los Molles Fm. (basin floor and slope), Las Lajas Fm. (shelf) and Challaco Fm. (fluvial) (Paim et al., 2008, 2011; Zavala, 1996) (Figure 1b).

Bey Malec Estancia in La Jardinera area in SW Neuquén Basin exposes outcrops of the oldest of these clinoforms in dip-oblique section (Figure 1c), for brevity termed the Bey Malec clinoforms herein after the name of the Estancia. Bottomset deposits of one of the Bey Malec clinoforms is dated by ammonites and volcanic ashes to the early Toarcian (182.4 ± 2.3 Ma) (Naipauer, Garcia, Manassero, Valencia, & Ramos, 2018) and the entire section is estimated to be deposited within a 3rd-order sea level cycle (Paim et al., 2008). The maximum water depth of the basin was estimated to be approximately 300 m (undecompacted) from the clinoform height. The objective and highlight of this study is to document and interpret several inclined stratal lines that connect neighbouring sand bodies within the Los Molles slope deposits of the Bey Malec clinoforms. Those lines are interpreted as clinoform timelines, representing contemporaneous surfaces, resembling hard reflectors with clinoform shape in seismic data mentioned above. The changing downslope channel geometry and sediment density flows within the channels found along those clinoform timelines is the subject of the analysis herein (Figure 2).

2 | METHODOLOGY AND RESULTS

Eleven vertical sedimentological sections, between 13 m (Log 62) and 374 m (Log 53) long, are measured at 10 cm vertical resolution over a 4 km length, covering most of

### Highlights

In this paper, we present a novel case study for sediment delivery in sediment supply dominant, water-depth limited (<500 m), prograding dominant basins, focusing on facies and architectural variability in deepwater slope channel fill with an outcrop from La Jardinera area in Jurassic Neuquén Basin, Argentina. Using sedimentological methods, remote sensing images and simple grain size analysis, the authors find that in a high-sediment supply system with not-too-deep-water depth (<1 km),

1. The basin-filling clinoforms are prograding-dominant;
2. The slope channels are relatively simple in geometry (single-storey channel) and small in size (<50-m thick and < 1 km in width). The channel forms are eroded and filled within a 4th order sea level cycle.
3. In a coarse sediment supply case like this one, the majority of the coarse grains are retained within the upper slope channels, although pebble-bypass to deeper water occur at times.
4. Upper slope channel fill contains a greater variety of facies and grain size variations than lower slope channel fill.
5. Architecture of upper slope channels possess certain degree of horizontal migration, whereas the lower slope channel fill is dominated by vertical aggradation.
6. Large amount of sediments is bypassed into deep water, gave rise to thick basin floor sequence.
the Los Molles deepwater slope and partially the top of the underlying basin floor fans of the Bey Malec clinoforms. A flat-lying, parallel-laminated, burrowed green sandstone unit less than 1-m thick is part of the clinoform topset and is marked as the correlation datum (Figure 2b–c, 3a). A 3D model constructed from drone images was used to map and correlate coarser sandstone bodies to highlight basinward-dipping clinoform timelines. Seven such timelines are visible on the exposed mountain side below the topset datum. Drone images are used for high-resolution mapping of the internal structure of the slope channels (Figures 2c–e and 4a, b). Nine lithofacies and four facies associations are distinguished based on grain size, sorting and sedimentary structures (Tables 1 and 2; Figure 3). Proportions of each grain-size range (e.g. fine sand, medium sand, course sand) are added from stratigraphic logs for each slope channel fill. The slope gradient is measured in general to be less than 4° between the sub-horizontal topset datum and the dipping timelines punctuated by slope channels, but can be locally as much at 10° just beyond the interpreted shelf edge. The internal character of the slope-channel fill is analysed for each channel element that is readily identified from sedimentary log correlations. The channel element is defined as a channel form surface plus the sediment that fill it (McHargue et al., 2011). The mapping of the slope-channel element architecture is aided by remote sensing images.

### 3 | OUTCROP CHARACTERISTICS

The slope-channels are recognized as sharp, erosive-based, amalgamated conglomeratic and sandstone units up to tens of metres thick with an overall upward-finishing fill pattern. All slope channels appear to be single-storey. The channel fill itself displays great heterogeneity, from muddy intervals, to very fine-grained, well-sorted sandstone with parallel lamination, to clast-supported, sandy matrix conglomerate showing subtle imbrication. There is a general downslope trend of channels becoming less deeply cut, and channel fill becoming finer grained and less conglomeratic, although without a predictable facies pattern (Figure 5a, b). However, the upward fining grain-size tendency within each individual single-storey channel is clear, and sorting improves upward within each channel element. The slope itself, when exposed, is thinly bedded mudstone with occasional mm to cm bedded very fine to fine sandstone. They are typical slope mudstone facies with intermittent sediment gravity flow deposits (Figure 4c).

#### 3.1 | Shelf deposits, upper slope-shelf transition and shelf-edge slope channel

On the shallow-water topset (including the shelf) of the clinoforms, sediment transport is dominated by fluvial, tidal and wave processes, whereas beyond the shelf edge, on the slope, the sediment transport is dominated by gravity flows. In this section, we will briefly describe the nature of the shelf portion of the topset of the clinoform units, with an emphasis on a slope channel identified just beyond the shelf edge on the slope. In outcrop, the shelf edge is often sharp but as a zone of interplaying fluvial, tidal, wave and gravity controls,
and this zone is sometimes termed the ‘shelf-slope transition’ (Dixon, 2013; Gomis-Cartesio, Poyatos-Moré, Hodgson, & Flint, 2017a; Gomis-Cartesio, Poyatos-Moré, Hodgson, & Flint, 2017b). Details of topset shelf deposits and the shelf-slope transition is further described in De Almeida Jr. et al. (2019), in review, where interaction of the shelf deltas with shelf-edge incisions is documented along both depositional dip and strike sections.

### 3.1.1 | Description

Logs 51, 52 and 53 transect the proximal end of the studied outcrop area (Figure 2a). The transition between the topset (shelf delta) and the slope is identified by an increase in slope gradient, seen between timelines 5 and 7 in Figure 2b.c. From sedimentary log correlations, two thick sandstone bodies in the slope portion of two clinothems are in close proximity but show different characteristics (Figure 2b). Body 1 is composed of massively amalgamated conglomerate and sandstone extending 400 m in length. The thickness varies laterally; the maximum is 60 m where it appears to have an erosive base, to 20 m where the incision is minimum. It is amalgamated by several conglomerate-rich, normally graded, lenticular bodies each about 10-m thick, and has an overall coarsening upward trend (Figure 2c). Body 2 is 40-m thick and contains matrix-supported conglomerates, very poorly sorted gravelly sandstone, very poorly sorted sandstone, and poorly sorted medium to coarse-grained sandstone facies (Figure 3c, d). The most common facies association is lenticular and flat-lying, thick-beded conglomerates interbedded with poorly sorted pebbly sandstone. The pebbles are matrix-supported in a very poorly sorted sandy matrix. Sand matrix-supported mudstone clasts (clustering within up to 50-cm-thick beds) and amalgamated beds are abundant. The upper surface of Body 2 shows a drastic change in the angle between Log 52 and Log 53 (timeline 7 in Figure 2c), but unfortunately the point of angle-change (left most red arrow in Figure 2c) is not accessible.

On top of Body 1 and Body 2, there are two sub-horizontal levels of sandstones are interpreted as topset shelf deposits because of their sub-horizontal attitude and their position along the clinoform. Bench 1 has the characteristics of distributary channels and mouth bars of a shelf-edge delta (Figure 2d, e), Bench 2 is interpreted as shelf deposits consisting of hyperpycnites with partial tidal reworking on the shelf during the following transgression and highstand of relative sea level. The truncation of the prograding clinoform timelines below by basal Bench1 implies a severely erosive transgressive surface (dashed green line, Figure 2c–e), subsequently enhanced by erosion and reworking during the following topset regression with the deposition of Bench 1 and its time equivalent (between the dashed and solid green line, Figure 2c–e). Unlike the bioturbated transgressive layer above the regressive Bench 1 (with a green colour that suggests likely glauconite content, Figure 3a), the transgression at the base of Bench 1 is not identified in the outcrop but interpreted because of the very rapid mudstone thinning (between timelines 6 and 7) and the toplap erosion (Figure 2c–e).

### 3.1.2 | Interpretation

The two sub-horizontal levels of sandstones are interpreted as topset shelf deposits because of their sub-horizontal attitude and their position along the clinoform. Bench 1 has the characteristics of distributary channels and mouth bars of a shelf-edge delta (Figure 2d, e), Bench 2 is interpreted as shelf deposits consisting of hyperpycnites with partial tidal reworking on the shelf during the following transgression and highstand of relative sea level. The truncation of the prograding clinoform timelines below by basal Bench1 implies a severely erosive transgressive surface (dashed green line, Figure 2c–e), subsequently enhanced by erosion and reworking during the following topset regression with the deposition of Bench 1 and its time equivalent (between the dashed and solid green line, Figure 2c–e). Unlike the bioturbated transgressive layer above the regressive Bench 1 (with a green colour that suggests likely glauconite content, Figure 3a), the transgression at the base of Bench 1 is not identified in the outcrop but interpreted because of the very rapid mudstone thinning (between timelines 6 and 7) and the toplap erosion (Figure 2c–e).

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**FIGURE 2** (a) A Google Earth perspective view (looking ESE) of the Bey Malec clinoforms, marking the position of measured logs (yellow), and mapped clinoform timelines (red). Vertical exaggeration is 1.5. (b) Simplified sedimentary logs, their correlation, and the Bey Malec clinoform timelines through them, with 1 being the oldest, 7 being the youngest. Nine lithofacies (Table 1) are simplified into four major facies associations (Table 2). Eight depositional environments are interpreted. (c) Detailed mapping of the shelf-slope transition area in 3D model constructed from done images. No vertical exaggeration. Note the three truncated clinoform timelines (red arrows) by an interpreted transgressive surface (green lines), and one more transgressive surface above it. In the outcrop, only the top transgressive surface is well-preserved (d) Perspective view (looking ENE) from drone images of the interpreted slope channel right at the shelf edge, and the deltaic distributary mouth bars. No vertical exaggeration. (e) The line drawing highlights the two different internal architecture of the channel fill on the slope and the delta on the shelf. The delta deposits have lenticular cross bedding and cross stratification, whereas the slope channel is filled by amalgamated (metre thick) sandstone with pebbles and gravels.
Body 1, with its overall upward-coarsening trend and internally normal-graded conglomeratic lenses, is interpreted as a shelf-edge delta with distributary channels. Body 2 is interpreted as a slope channel fill right beyond the shelf edge, with the top truncated by a transgressive erosion surface underlying renewed shelf-edge delta.
progradation (Bench 1). Some stacked pebbly sandstones with horizontally aligned gravels resemble those shown in Figure 6b and c in Kneller and Branney (1995), have been likely deposited by progressive aggradation from quasi-steady flow. Body 2 is henceforward referred as an ‘upper slope channel’.

Although not described from the Bey Malec clinothems, there are two other younger shelf edges mapped in the La Jardinera area, that show a sigmoidal rather than an oblique clinoform pattern, with different types of conglomeratic transitions onto the slope. They exhibit well-stratified, well-imbricated and fairly well-sorted, channelized fluvial conglomerates on the shelf, passing quickly into thick-bedded, mainly non-graded, poorly sorted sand matrix-supported conglomerate found beyond the shelf edge on the upper slope. In this case there was therefore a continuous transition from fluvial channels to subaqueous slope channels, in contrast to the toplap truncated clinoforms, and dis-connectedness in the current Bey Malec location. The details of this transitional case aid our interpretation of conglomeratic bodies in Bey Malec clinoforms.

### 3.2 | Slope channel fill

The ‘slope channels’ described here are all single-storey sandstone elements identified in the slope segment of the outcrop, but excluding the Body 2 described above. The slope channels described below are referred to as ‘lower slope channels’ in the following discussion. The slope channels identified are from clinoform timeline 4 to timeline 7 in Figure 2b, and those from older timelines are automatically more distal than those of younger clinoform timelines because of the clinoform progradation.

#### 3.2.1 | Description

Medium-grained sandstone dominates in the channels. Amalgamated structureless sandstone units and upward-fining sandstone units are the two most common facies associations (Figure 5b), composed of gps, vps, ps, wss and fs facies (Table 1). The thickness of these slope channel elements is 6.4 m at minimum (Log 62 in Figure 2a), 26.4 m at maximum (Log 59), and the net to gross ratio varies between 43.1% and 100%. The single-storey channel intercepted by Log 55 on the east end & Log 57 on the western end serves as a type example for slope channel elements (Figures 2a and 4b). The channel element comprises amalgamated beds of two categories: (a) structureless sandstone with vps and ps facies (Table 1) with occasional floating pebbles (counts towards 56% of slope channel facies associations), and (b) normally graded, coarse sandstone beds with a pebbly base (grain size...
maximum 3 cm) and a very poorly sorted sandstone (vps facies) top, with reducing clast density and size upwards in the bed. In every slope channel (Log 59, 60, 62), the graded beds fine from coarse to medium sand without the presence of clasts. Together, normally graded beds count for 28% of facies associations in the slope channels (Figure 5b). In addition to normal-graded beds, coarse-tail graded beds are also present. The bed thickness of structureless beds varies between 0.1 and 1.8 m. However, the thicker beds appear to be amalgamated and the true bed thickness within such structureless beds is only discernible with mud clast layers. Graded bed thicknesses vary between 0.3 and 2.1 m, again with the error of amalgamation.

### Table 1: Lithofacies identified in the field area and their interpretations

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Description</th>
<th>Bed thickness</th>
<th>Interpretation/depositional process</th>
</tr>
</thead>
<tbody>
<tr>
<td>cgl&lt;sub&gt;in&lt;/sub&gt;</td>
<td>Conglomerate. Rounded to subrounded clasts, dominant grain size 2–5 cm, maximum grain size 8 cm, matrix-supported. Matrix is poorly sorted sandstone with gravel, very-coarse and coarse sand. Structureless or fining upward.</td>
<td>Decimetre to metre scale</td>
<td>Debrite and high-density turbidite</td>
</tr>
<tr>
<td>cgl&lt;sub&gt;c&lt;/sub&gt;</td>
<td>Conglomerate. Rounded to subrounded clasts dominant size 2–5 cm, maximum grain size 8 cm, clast-supported. Faint imbrications. Matrix is poorly sorted sandstone with gravel, very-coarse and coarse sand. Structureless or fining upwards. Sometimes inverse grading in the basal few centimetres of the beds.</td>
<td>Decimetre to metre scale</td>
<td>Debrite and high-density turbidite.</td>
</tr>
<tr>
<td>gps&lt;sub&gt;(par)&lt;/sub&gt;</td>
<td>Very poorly sorted sandstone, maximum grain size up to granule, matrix-supported. Gravel grains are subangular. Gravel clasts can be aligned in long axes. Some 2- to 3-cm-sized pebble clasts ‘float’ in a matrix of poorly sorted sand.</td>
<td>Up to metre scale</td>
<td>High-density turbidite (S&lt;sub&gt;1&lt;/sub&gt;, S&lt;sub&gt;2&lt;/sub&gt;, Lowe, 1982)</td>
</tr>
<tr>
<td>vps</td>
<td>Very poorly sorted sandstone, dominant grain size range upper-medium to upper-coarse, maximum grain size very-coarse-upper or even 5-mm-scale gravel. Structureless or fining upwards (coarse tail grading). Some 2- to 3-cm-sized pebble clasts ‘float’ in a matrix of poorly sorted sand.</td>
<td>Up to metre scale, massively bedded</td>
<td>High- and/or low-density turbidity current deposits (S&lt;sub&gt;1&lt;/sub&gt; Lowe, 1982, T&lt;sub&gt;a&lt;/sub&gt; Bouma, 1962)</td>
</tr>
<tr>
<td>ps&lt;sub&gt;(par)(x)&lt;/sub&gt;</td>
<td>Poorly sorted sandstone, grain size often medium-lower to coarse-upper sand, dominated by medium sand grain size. Structureless or fining upwards.</td>
<td>Up to metre scale</td>
<td>High- and/or low-density turbidity current deposits (S&lt;sub&gt;1&lt;/sub&gt; Lowe, 1982, T&lt;sub&gt;a&lt;/sub&gt; Bouma, 1962)</td>
</tr>
<tr>
<td>wss</td>
<td>Well-sorted sandstone, grain size often upper-fine to lower-coarse and span two grain size ranges (e.g. upper fine to lower medium, lower medium to upper medium). Structureless, parallel-laminated or fining upwards.</td>
<td>Decimetre</td>
<td>Low-density turbidite (T&lt;sub&gt;a&lt;/sub&gt; Bouma, 1962)</td>
</tr>
<tr>
<td>fs&lt;sub&gt;(par)(x)&lt;/sub&gt;</td>
<td>Well-sorted sandstone, grain size very fine to lower-medium. (With parallel lamination (cross-bedded)).</td>
<td>Decimetre</td>
<td>Low-density turbidite (T&lt;sub&gt;a&lt;/sub&gt; Bouma, 1962). Sheet turbidite. Hyperpycnite.</td>
</tr>
<tr>
<td>mdc</td>
<td>Horizons of mudstone and siltstone rip-up clasts in a matrix of poorly sorted sandstone. The grain size of the sandstone matrix is consistent with the overlying sandstone. Mud clasts are often deformed and sheared, length (long axis) can be up to 20 cm. Mostly found at base of beds, very rarely at the top.</td>
<td>&lt;20 cm</td>
<td>Mudstone clasts entrained from erosion of underlying mudstone and siltstone during flow scouring.</td>
</tr>
<tr>
<td>mud</td>
<td>Mudstone, parallel-laminated, can be interbedded with thin (&lt;2 cm) siltstone and sandstone.</td>
<td>mm-and cm-scale beds, amalgamate into hundreds of metres</td>
<td>Slope mud settled from the suspended fine grains of sediment gravity flows and hemipelagic fallouts</td>
</tr>
</tbody>
</table>

#### 3.2.2 Interpretation

The structureless and graded beds of the channel fill were deposited by high-density and low-density turbidity currents respectively (Bouma, 1962; Lowe, 1982; Middleton & Hampton, 1973; Talling, Masson, Sumner, & Malgesini, 2012). The lack of inversely graded beds in slope channels suggests that the sediment concentration was not high enough to form traction carpets (Lowe, 1982; Sohn, 1997). No debrites have been seen in the lower slope channels. Since the grain size in sediment transport is proportional to the height above the channel base (Jobe et al., 2017), the grain size variation at the base of the channel element is likely caused by grain size variation in the initial flow, associated with trigger
events at the shelf edge. Since the investigated cross section is oriented at an angle to the depositional dip, some of the variations could also be caused by channel thalweg migration.

To the left of the channel intercepted by Log 57 & Log 55, 7.3 m of amalgamated sandstone in thinly bedded vps and ps facies, shows bedding boundaries inclined at an angle to those interpreted in the channel element right next to it (Figure 4b). The inclined beds could be proximal levees accumulated immediately next to the channel itself, or splay deposits, which would explain the coarse grain size.

### Table 2: Facies associations in the Bey Malec clinoform

<table>
<thead>
<tr>
<th>Facies associations</th>
<th>Description</th>
<th>Bed thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conglomeratic</td>
<td>conglomeratic and sandstone dominant with gps and vps towards the top of the bed. Structureless or fining upwards (in pebble size and pebble amount).</td>
<td>In upper slope and close/on the shelf, metre-thick. In lower slope channels, bypassed conglomeratic beds are ~ 10 cm and displays rapid fining upwards pattern.</td>
<td>Upper slope conglomerates are coarse-grained debris flows (structureless and minor fining upward), lower slope conglomerates are high-density turbidites.</td>
</tr>
<tr>
<td>Mixed conglomeratic and sandstone</td>
<td>gps and vps dominant, with occasional floating pebbles (1–2 cm) in matrix in lenticular horizons. For thicker beds (&gt;10 cm), gravel size and amount reduce upward.</td>
<td>Alternating conglomeratic and coarse sandstone bed each at ~ 10 cm.</td>
<td>Mixed debris flow and high-density turbidite.</td>
</tr>
<tr>
<td>Structureless sandstone</td>
<td>vps, ps with floating gravels and pebbles. Amalgamation marked by mud clast-rich layers. Can have cm-scale parallel laminations at places</td>
<td>Individual bed thickness in decimetres. Amalgamated thickness metre scale.</td>
<td>High-density turbidite and low-density turbidite. Hyperpycnite.</td>
</tr>
<tr>
<td>Fining upward sandstone</td>
<td>vps, ps, wss and fs facies. Bed thickness in decimetres (&lt;50 cm). Top of the bed can have cm-scale parallel laminations.</td>
<td>Decimetre</td>
<td>High-density turbidite and low-density turbidite.</td>
</tr>
</tbody>
</table>

**Figure 4** Internal architecture of slope channels, location shown in Figure 2. (a) In the upper slope, slope channels are composed of thick-beded, mixed conglomeratic and sandstone facies. Beds are metre scale, with a migration direction at an angle to the local palaeocurrent direction, indicating lateral accretion of the slope channel fill. Palaeocurrents \( n = 7 \) is measured from pebble imbrications in the deltaic distributary channels nearby (b) Lower slope channel. The beds are thinner (decimetre scale) and lack the inclining-migration pattern observed in upper slope channels. Note the difference in vertical scale. (c) Muddy levees in the slope about 10 km N of the study area. Yellow arrows mark downlapping surfaces, person in red circle for scale.
The upper-slope channel close to the shelf edge is notably different in architecture to the single-storey channels on the lower slope. This upper-slope channel (Body 2, Figure 2c, e) displays internal scouring and erosional surfaces showing a slight downslope prograding pattern of internal surfaces downlapping onto the base of the channel, but at an angle to the local palaeocurrent direction (Figure 4a), indicating the incise-and-fill nature of the channel with a degree of lateral accretion or even some lateral migration of the incisions themselves. The lower slope channels, in contrast, are aggradational in their internal architecture, as the internal amalgamation surfaces are often parallel to the base of the channels (Figure 4b). The upper-slope channel fill is much thicker than the fill of the lower slope channels (Figure 5c). This can have two explanations that are not mutually exclusive of each other: either the upper slope was incised much more than the lower slope, due to the intensity of erosive sediment density flows right at the shelf edge where the gradient is the steepest and the sediment density is maximum without any notable deposition;
or, the lower slope channels have a higher proportion of muddy fill towards the top of the channel, and this is less detectable in the outcrop, whereas the upper slope receives most of the coarse sediments and is thus filled with sediments more readily preserved. A schematic plot of channel architecture is given in Figure 6 summarizing the results concluded from Figures 2, 4 and 5.

### 3.4 Basin floor deposits

#### 3.4.1 Description

Several of the measured logs include thin-bedded sandstone units in the Los Molles Fm occurring below the slope segment (clinoform timelines 1, 2, 3 in Figure 2). Compared to the slope channel fills, the turbidite bed thickness of these bottomset units is in general markedly thinner (on the scale of tens of centimetres, mostly less than 1 m). Finer grain-sized facies, such as ps, wss and fs (Table 1), are prominent. Occasionally, gps and cgl facies (Table 1) are present. There are more structureless beds ($T_a$) than normally graded beds ($T_b$). Occasional cm-scale parallel laminations are observed on top of fine-grained beds (Figure 3i). Mudstone clasts are common at the bed boundaries, marking the base of amalgamated sandstone beds.

#### 3.4.2 Interpretation

Based on their stratigraphic position, these sandstone units are interpreted as basin floor fan deposits. The depositional process was a combination of high-density and low-density turbidity currents. The lateral thickness variations suggest some low-relief channelization (Timeline 2, Figure 2b). The coarse grains are associated with coarse-grain bypass from the shelf. The dominance of the structureless beds can be explained by hindered coarse-grain settling and fast deposition at the proximal head areas of the basin floor fans (Normark, 1978; Walker, 1967). Mudstone clasts at bedding boundaries suggest that a density increase caused by erosive slope sediment entrainment is also a plausible explanation. For a more detailed study on Los Molles basin floor fans, see Giacomone, Olariu, Steel, & Shin, 2019.

### 4 DISCUSSION

#### 4.1 Slope channel facies and architecture

A grain size analysis performed for each slope-channel element in Bey Malec clinoforms (Figure 5a) shows that the system is relatively coarse-grained, especially close to the shelf break. Medium-grained sandstone dominates the system. Even in the lower slope channels, fine sand grain size is present only occasionally. Most of the channel fills are dominated by medium and coarse-grained sandstone. Downslope fining of the grain size distribution is observed as the conglomerate/sand ratio decreases within slope channels (Figure 5a). A schematic trend of the slope channel fill is plotted in Figure 6.

The slope channels in Bey Malec clinoforms are single-storey, each cut and filled by deposition from multiple sediment density flows as suggested by the amalgamation of channel fill. Channelization is possibly related to periods of low relative sea level, when the shelf is completely absent or narrow and large volumes of sediment arrive at the shelf edge and transit directly into slope channels through the shelf-edge delta (Paim et al., 2008; Porębski & Steel, 2006) which also promote delivery of the coarsest sediments. Occasional shelf-edge collapse can initiate channel formation as well, but vegetation cover on the mud-dominated slope prevents observation of such features in the outcrop. The amount of sediments delivered to the shelf edge and into upper slope channels would have reduced during periods of rising sea level and widening of the shelf, promoting deposition within the channels. Yet, in other areas such as areas close to steep fluvial systems, channelization and sediment bypass would have persisted, governed by the direct sediment supply from the river. Channelization could also occur simply during strong flood periods with high water/sediment ratios, irrespective of sea-level stand.

Delivery of pebble-sized grains down onto the basin floor suggests relatively simple and straight channel systems without much sinuosity to develop inner terraces and bends to trap coarse grains as channel lag. However, some degree of meandering was present, as seen in the lateral accretion of the upper slope channel and the inclined levees of some lower slope channels. Most
flow transitions occurred at the shelf break and in the upper slope channels where the slope gradient was steepest, apparently approaching 10°, potentially as a combination of originally steep shelves and differential compaction on the slope mudstone and shelf sandstone. Entrainment of water into the sandy debris flows in high-velocity upper slope channels could have sufficiently diluted flows to create high and low-density turbidity currents, as documented by downslope facies change within the slope channel fills (Figure 5) (Lowe, 1982; Mutti, 1992; Talling et al., 2012). This also explains the absence of debrites in lower slope channels. There is a notable lack of dewatering features everywhere along the Los Molles slope channel deposits, potentially related to the coarse grain sizes, so that intragranular pressure did not rise high enough to produce dewatering structures.

Amalgamation of event beds within the single-storey channels is significant for both upper and lower slope channels (Figures 3c, d and 4b), but amalgamation of channel elements into channel complexes is not observed at these outcrop locations. There is no statistical correlation between the channel position below the shelf edge and the thickness of the channel element despite the general trend (Figures 2, 5c, 6).

Levees are identified in the lower slope by the presence of mm to cm thick, very fine and fine sandstone beds that dip and thin laterally with downlapping geometry (Figure 4b, c). Most of the lower slope has been constructed by such muddy overspills (with sandier content in the proximal levee) from the small but densely distributed slope channels. The lower slope gradient is moderate at 2–4°, and no large-scale mass transport complexes are present, although small scale (<10 m) slumps of the slope mud facies are seen in the lower slope, where there is more loading onto the slope deposits. This can suggest that the sediment supply drove relatively frequent but small sediment gravity flows so that the slope maintained a relatively graded profile (Pirmez et al., 2000; Prather, O’Bryne, Pirmez, & Sylvester, 2017), which is consistent with hosting simple channel geometries.

4.2 Sediment delivery

Figure 2a shows that slope channel elements are associated with the prograding portion of the Bey Malec clinoforms. There are only minor sandstone bodies (thickness < 5 m) between clinoform timelines 7 and the Datum, as the interpreted transgression between the timelines (dashed green lines in Figures 2 and 5) reduced sediment supply to the shelf edge. The sediment source for the Bey Malec slope channels and their updip coeval shelf deposits were derived mostly from the upstream fluvial Challaco Formation, with a provenance from a combination of the North Patagonia massif to the southeast and some from the active volcanic arc to the west (Naipauer et al., 2018). In the Middle Jurassic southern Neuquén Basin, large sediment volumes were generated by a tectonically active hinterland (Kim, Malla, Gutierrez, & Malone, 2014). The transportation of coarse grains over the shelf into the slope channels was aided by shelf-edge delta formation, especially with early-stage narrow shelf, either caused by periods of low sea level (Porębski & Steel, 2006) or simply because of high sediment yield from tectonically active regions (Carvajal & Steel, 2012). The presence of pebbles in slope channels limits the river mouth-to-shelf-edge distance to less than 10 km, as this is the suggested maximum distance pebble-sized grains should be transported across the shelf into the deep water slope channels (Sweet & Blum, 2016). This is also consistent with steep hinterland slopes and limited distances of shelf transgression during a greenhouse climate (Steel, Olariu, Zhang, & Chen, 2019).

An intermittent connection to the Pacific Ocean introduced a mesotidal range to the Neuquén back-arc basin (Dean, 1987) that has left extensive tidal deposits on the shelf. These cross-shelf tidal current fills (Rossi et al., 2016) contributed to sediment delivery into the Bey Malec slope channels, especially aiding the delivery of fine sediments during relative sea level highstands, albeit modest greenhouse highstands. Variations of palaeocurrent direction suggest that there were local changes of sediment source potentially associated with delta lobe switching and channel feeder avulsion, as well as autogenic evolution of deepwater slope channels (e.g. Reimchem et al. 2016).

Single-storey channels in the study area suggest a lack of time for channels to evolve into complex meandering terraced systems before being filled up or abandoned. The strongly progradational shelf edges of the main distributary fairways mapped from the same outcrop (Olariu et al., 2019, this volume) resemble those associated with sediment supply dominated margins where multiple contemporaneous slope channels form and fill in relatively short timescale due to large flood-generated sediment supply from the shelf. This short timescale can be less than or equal to a 4th order sea level cycle, even in Jurassic greenhouse time, as suggested by high-frequency shore line transits in Wilcox formation (Zhang, Steel, & Ambrose, 2016) that was also severely modulated by variable sediment discharge.

Since the slope channels hold volumetrically insignificant volumes of sandy deposits, the majority of the sediments, both fine and coarse-grained fractions, bypassed the shelf edge and ended up on the Los Molles basin floor, and produced thick sandstone bodies with even conglomeratic occurrence (Figure 2b) (Table 3) (Paim et al., 2008, Olariu et al., 2019 this volume, Giacomone et al., 2019), despite the relative constant thickness of the slope section (~300 m) and the small dimensions of the slope channel conduits (less than 50-m thick).

Overall, the La Jardinera system serves as a good example the coarse-grained, line-sourced concept in Reading and Richards (1994)’s facies model.
4.3 Clinoform configuration and comparison with other systems

The Bey Malec clinoforms were deposited during the early thermal sag phase that gradually healed and extended the isolated depocentres of Neuquén Basin from Late Triassic to Middle Jurassic (Vicente, 2005). More regional studies show that Jurassic sedimentation of the Cuyo Group was still coeval with subsidence and pulsed contractional deformation associated with the activity of Huincul High, on which the La Jardinera area sits (Naipauer et al., 2012). The Bey Malec clinoforms are likely to represent the early stages (early Toarcian) of basin fill where mature clinoforms (with a clear shelf-edge trajectory) just developed from an initial ramped (without shelf edge) basin margin, and were strongly driven by a locally controlled sediment source, explaining some westward palaeocurrent directions deriving from the overall northward sediment transportation direction (c.f. Zavala, 1996).

Similar progradation-dominated clinoforms are found elsewhere, especially in relatively shallow back-arc, foreland, and lacustrine basins, where high sediment supply from the nearby mountains promoted sand delivery into deep water. Known examples include the Lewis-Fox Hills system of the Maastrichtian Washakie Basin (Carvajal & Steel, 2006), the Miocene Dacian Basin (Fongngern et al., 2018) and the Neocomian Siberian Basin (Pinous, Levchuk, & Sahagian, 2001). The Spitsbergen system, although of similar scale, is interpreted to have four types of clinothems distinguished by architecture and sediment partitioning as the result of interplay between accommodation and amount of sediment supply (Plink-Björklund et al., 2001). These supply dominated systems, usually with isolated and short-lived slope channels, contrast with other systems of large river and associated long-lived canyons fixed at a single shelf-edge location, such as the Amazon fan (Pirmez & Flood, 1995), the Congo-Zaire canyon (Babonneau, Savoye, Cremer, & Klein, 2002) and the Bengal fan (Curray et al., 2003).

There's no obviously matching modern analogue to the Bey Malec clinoforms, however, the Golo system on the eastern slope of Corsica in the Mediterranean (Table 3) bears the most resemblance. Mountainous granitic and sandstone-rich terrane (max height 2,700 m) of Corsica is drained by the Golo river that then drives several kilometres of delta progradation over a 5–10 km wide shelf (Deptuck, Piper, Savoye, & Gervais, 2008). The slope gradient is 2–3° and no mass transport deposits are present (M. Sweet, 2018, personal communication). Coarse sand to clay sediments are delivered to deep water through slope channels, and gullies associated with shelf-edge collapse form at the shelf edge. Using McHargue et al.’s (2011) aspect ratio of slope channels as 20–30, the Bey Malec slope channels, with thickness 10–50 m, are calculated to be 150–800 m.
in width, being on the same order of magnitude as gullies of the Golo system. Both the Golo and Bey Malec slope channels are associated with multiple basin floor fan complexes of similar dimensions (Deptuck et al., 2008). However, one has to bear in mind the noted sea level control of the Golo system, as the canyons of a couple of kilometres width may have been active only during past falling or low sea level stages (Gervais, Mulder, Savoye, & Gonthier, 2006).

5 | CONCLUSIONS

In a 4-km-long N-S exposure, in Bey Malec Estancia in the La Jardinera area in Jurassic Neuquén Basin, Argentina, deepwater slope channels are mapped along clinoform timelines from the shelf to the basin floor, revealing the downslope facies and architectural changes. All identified channel elements are less than 50 m in thickness and estimated to be under 1 km in width with a simple internal architecture (i.e. single storey). There is a general trend of downslope grain-size fining in slope channel fills. The lower slope channel architecture has a greater degree of internal amalgamation. Those slope channels are typical for supply dominated systems, where the channel configuration is relatively simple, composed of only single-storey channels filled with an upward fining sequence of amalgamated high- and low-density turbidites, indicating multiple episodes of sediment bypass. Despite the relatively small dimension of the slope channels, aside from a significantly thicker incision at the upper slope with the coarsest grain size, they are still able to deliver large volumes of coarse-grained sand into the deepwater.

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DATA AVAILABILITY STATEMENT

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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