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Confined Turbidite Sand-Sheets in the Paraná and Paganzo Basins.

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> Dissertação de Mestrado apresentada como parte das exigências para a obtenção do título de Mestre, pelo Programa de Pós-Graduação em Geologia da Universidade do Vale do Rio dos Sinos (UNISINOS).
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ACKNOWLEDGMENTS

ABSTRACT

During the Late Paleozoic, Gondwana was home for a protracted glaciation that generated high sediment supplies and ample accommodation space within incised valleys and fjords, usually laterally confined and longitudinally extensive geomorphological features. These glacial valleys became so overdeepened below sea level that they filled up with hundreds of meters of seawater, as glacio-eustatic sea level rose during and following deglaciation, therefore producing deep-water environments prone to gravity-flows such as turbidity currents. This study assesses the behavior and morphology of sand sheets deposited by turbidity currents that instead of flowing outward onto an unconfined basin floor, were confined within a limited space up to a spill point. This study compare three examples of turbidite sand sheet deposited on settings presenting distinct degree of confinement, including two areas in the Paganzo Basin (Paganzo Group) and one area in in the roughly coeval Paraná Basin (Itararé Group) The Argentinean case studies are located in the eastern Precordillera of the Andes, near San Juan city, and include a palaeofjord (Quebrada de Las Lajas) and a more open valley (Quebrada Grande), both incised into Early Paleozoic shelf carbonates (San Juan Formation) and olistostrome (Rinconada Formation). The Brazilian counterpart is located in Vidal Ramos (Santa Catarina State, Southern Brazil) and it is a very wide palaeovalley carved into Proterozoic schists and secondly marbles. After a general description and interpretation of the valleys fills and a more detailed description and interpretationof the associated turbidite sand sheets, their similarities and differences are stressed to define which features ewere dependent or independent of the glacial valley morphology.

1. INTRODUCTION

Deglacial marine sedimentary sequences record the melting retreat of glaciers. Along the Late Palaeozoic, glaciation affected Gondwana over 100 million years, and glaciogenic deposits can be found in all present day continents formerly assembled in the Gondwanaland (Crowell, 1983).

Fjords (long, narrow inlets with steep sides or cliffs) and valleys can provide especially detailed records of environmental changes associated with deglaciation due to the large amount of accommodation space available within these glacial, overdeepened features. Their typical high deposition rates provide a high-resolution record (e.g. Eyles *et al.*, 1997; Kneller *et al.*, 2004). These local basins are overdeepened during glacial excavation, and commonly left under hundreds of meters of seawater following glacial retreat and relative sea level rise (e.g. Powell and Molnia, 1989; Eyles *et al.*, 1990; DaSilva *et al.*, 1997; Benn and Evans, 1998). Also, high sedimentation rates during and following deglaciation can emplace hundreds of meters of stratigraphy in relatively short time spans (hundreds to thousands of years), therefore enhancing resolution of their sedimentary records (e.g., Cowan *et al.*, 1988; Powell and Molnia, 1989; Eyles *et al.*, 1990; Lysa *et al.*, 2004).

Deposition tends to respond to seasonal cycles, as snowmelt, rainy seasons or seasonal thawing of a valley glacier (Dykstra *et al.*, 2007). Spring or summer melts periods can be responsible for the greatest sediment pulses. Sediment failures may also contribute significantly to sediment gravity flows in these confined settings. Commonly produced from delta-fronts, usually related to Gilbert-type deltas, or rock fall off the steep fjord walls, they can evolve into turbidity currents upon entrainment of ambient water and flow for very long distances down to local basins (Prior *et al.*, 1987; Benn and Evans, 1998; Kneller *et al.*, 2004; Dykstra, 2005).

At last, it is worth to notice that Gondwana glacial deposits are important hydrocarbons producers in Oman (Levell *et al.*, 1988), Argentina and Bolivia, and are being actively explored in Brazil, Saudi Arabia and Australia.

1.1 Geographic Location

The data set was acquired in the well exposed Paganzo Basin and in the densely covered Paraná Basin in order to use the first one as a possible analogue to interpret the second one. As both study areas in the Paganzo Basin represent more (Quebrada de Las Lajas - QLL) or less (Quebrada Grande - QG) confined depositional settings, the area that was chosen in Brazil (Vidal Ramos - VR) was also supposed to represent some sort of depositional confinement.

The Argentinean areas of the Paganzo Basin include a palaeofjord (QLL) and a palaeovalley (QG), both located in the western sector of San Juan city (Fig1.1), which were rapidly filled subsequently to the Pennsylvanian Gondwana glaciation (Dykstra, 2005). They show excellent exposures, minimal structural overprint, and a three-dimensional view that allow a detailed observation of the evolution of their thick turbidite fill.

On the other hand, the Brazilian counterpart comprises a previously mentioned palaeovalley (Santos, 1996) situated near the south-eastern border of the Paraná Basin in the Vidal Ramos (Santa Catarina State) surroundings (Fig1.2). Its rock record is poorly exposed due to dense vegetation cover. A reliable representation of the turbidite facies geometry, architecture and stratigraphy can only be assessed from scarce outcrops.
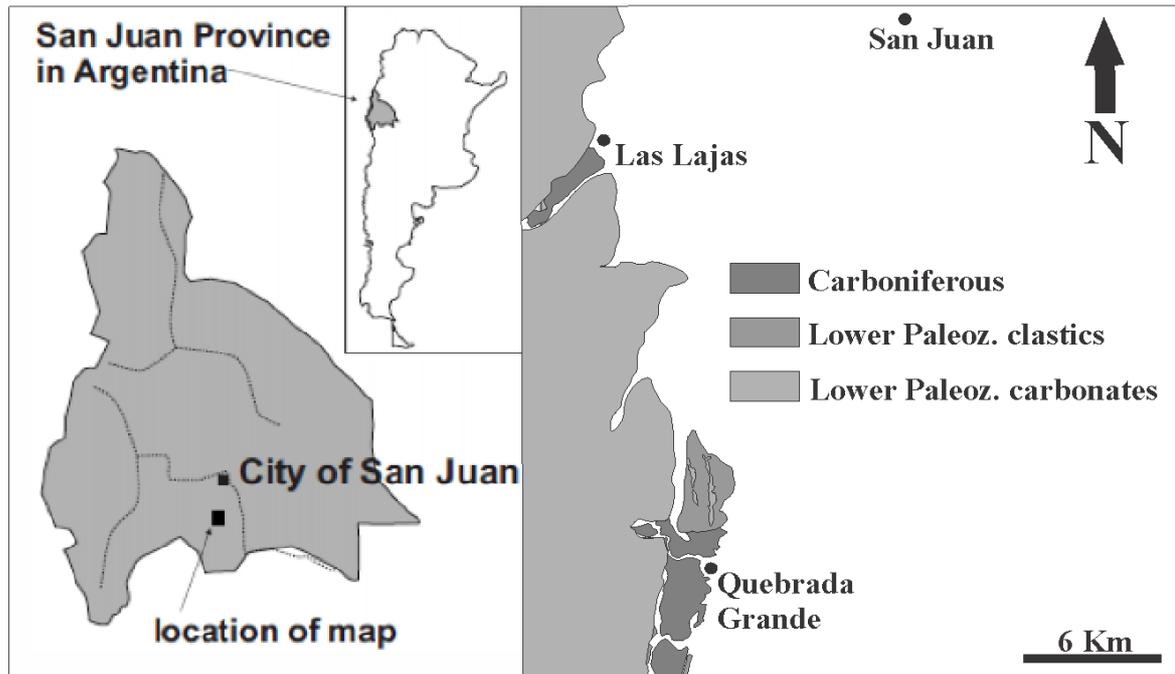
Figure 1.1: Location map of the Quebrada de las Lajas (palaeofjord) and Quebrada Grande (palaeovalley).
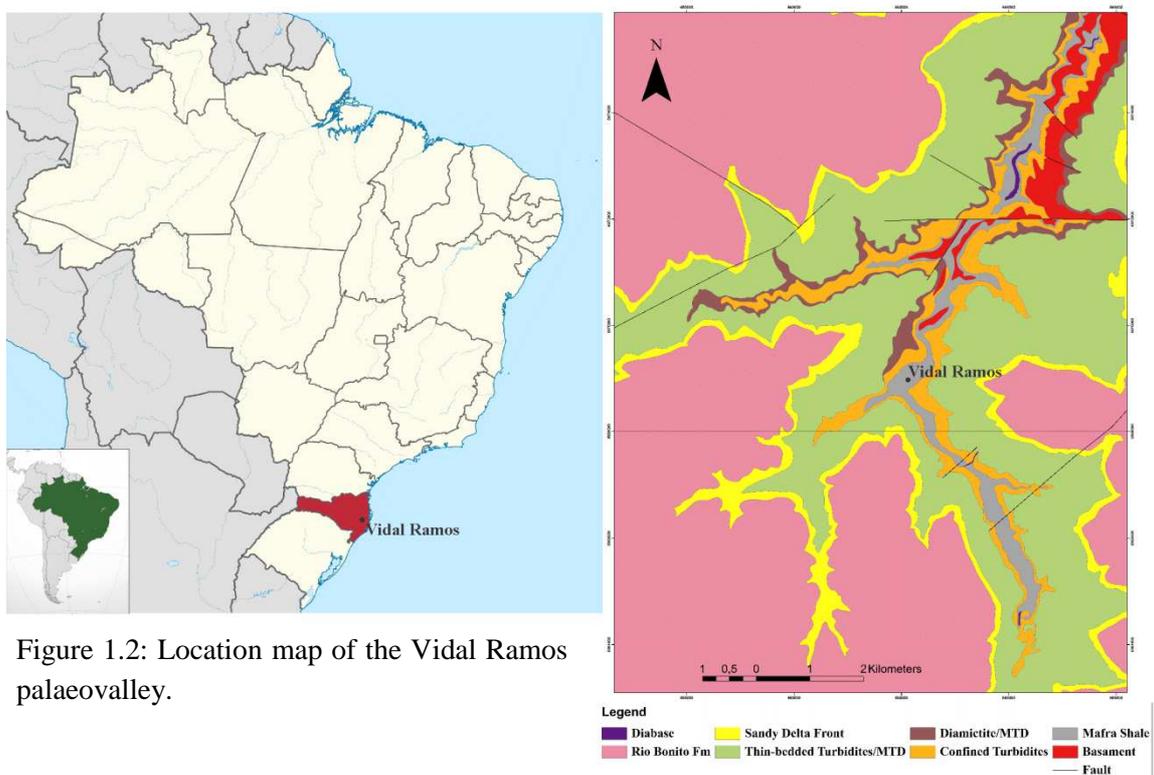
Figure 1.2: Location map of the Vidal Ramos palaeovalley.

1.2. Methodology

Data collection included three month of field work in the Paganzo Basin and two months and a half in the Paraná Basin. Field data was later processed in lab during a 2 year period. A geological map of each study area was produced in order to understand and organize individual sedimentary successions.

The lack of topographic maps as well as the arid climate in the Paganzo Basin lead to the use of satellite and aerial images, GPS marker-points and outcrop description to produce the QLL and QG geological maps (1:10,000). Their incorporation into ArcGis project has enabled later input of data acquired from laser-shots used to locate unreachable points and delineate inaccessible stratigraphic surfaces. It is worth to notice that the QG northern sector had been previously mapped (Kneller, 2004), so only its southern sector, a 2km x 1km area mostly comprising gravity flow deposits, was actually mapped.

In the Paraná Basin (VR), topographic-map, GPS points, laser-shots and outcrop features were used to produce a larger-scale (1:100,000) geological map in an area characterized by dense vegetation and usually largely weathered exposures.

Both in Argentina and Brazil, several sedimentary logs were acquired in strategic points during the mapping to describe facies, facies association and later to define turbidite stratigraphic architecture. To improve depositional elements characterization, the analysis of photo-mosaic and 3D images (supported by laser-shooting) was performed in both QLL and QG.

To demonstrate glacial movement and sediment transport paths within the confined local basins, paleocurrent readings were taken from all mapped units, usually presented as rose diagrams and / or mean vectors, and from the basal surface scoured into the Precambrian basement. To better differentiate distinct turbidite intervals and to bring into light some pieces of evidence related to source areas, some samples were taken to petrographic analysis. A few, dark grey to black shales were sampled to TOC analysis.

1.3. Objectives

This MSc dissertation aims to describe and compare turbidite sand sheets deposited during the Gondwana, late Palaeozoic de-glaciation on distinct, poorly- to highly-confined depositional settings. To accomplish this main goal, three objectives were pursued:

- To demonstrate the confined nature of the turbidite succession and to quantify the degree of confinement of each area;

- To erect individual models of facies association, stratigraphic architecture and overall geometry of turbidite strata in each area;

- To compare the turbidite systems taking into account their specific geological settings.

As a result this work may also be useful for exploration and exploitation in the Paraná Basin as well as in other comparable successions, including Brazilian offshore. Its results may also be used for training of subsurface teams in deep-water systems using field trips and digital-outcrop data.

2. CONCEPTUAL REVIEW AND GEOLOGICAL SETTING

2.1 Glacier-Valleys

A glacier corresponds to a mass of ice formed by the compaction and recrystallization of the precipitated snow that moves downslope by the action of gravity. According to their occurrence, glaciers can be subdivided into valleys-glaciers and continental-glaciers (Miller, 1996).

Valley-glaciers, also known as alpine-glaciers, have a dendritic drainage (Miller, 1996), and are formed by the drift of ice accumulated at high altitudes through snow precipitation (fig 2.1).
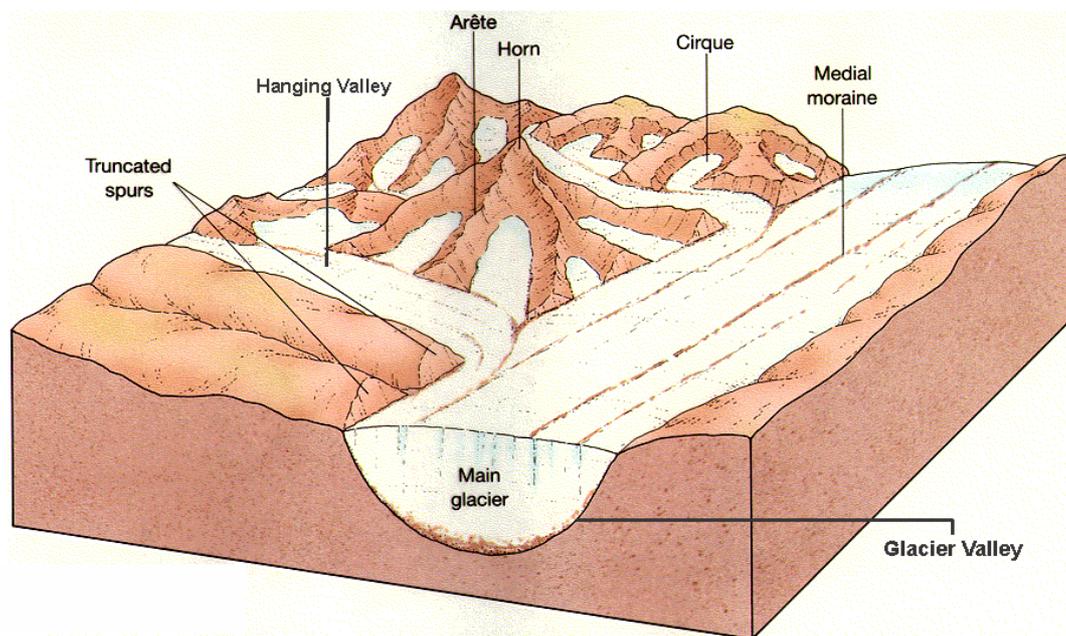
Figure 2.1: Glacial-valley and associated landscapes. Main glaciers carve the basement to create the main glacier-valley and adjacent hanging-valley (modified from Surachit, 2007)

The shape of the valley changes as the glacier moves downhill. A V-shaped valley is deepened and widened, and therefore transformed into a U-shaped valley through the glacial erosion processes of plucking and abrasion, creating perfectly confined setting.

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The mass balance profile (fig 2.2) and the valley geometry combine to set the pattern of glacial modification of the valley.
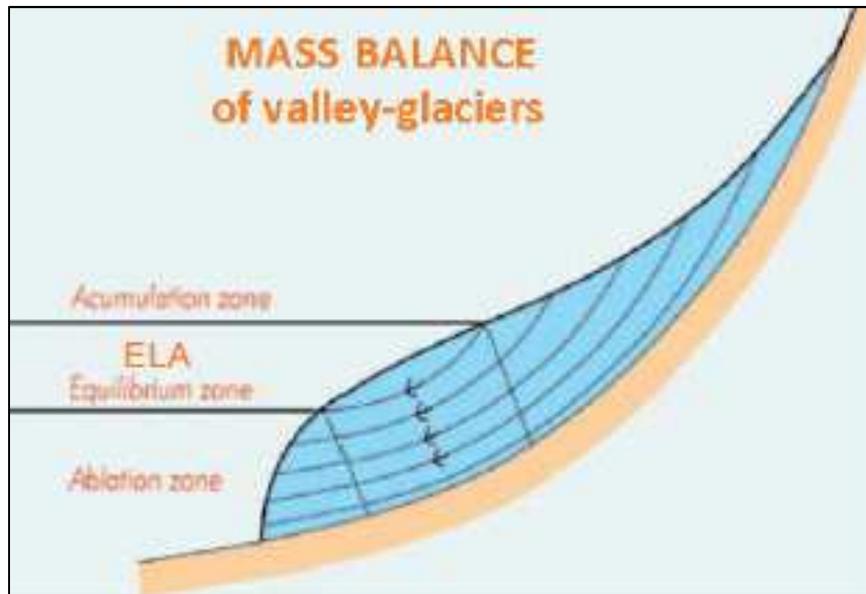
Figure 2.2: Mass balance in a valley-glacier (Embleton & King, 1975)

The interplay of the dynamic properties of the eroding glacier and the mass strength of the subglacial bedrock control the valley morphology. When the pattern of ice discharge reaches its maximum at the equilibrium line altitude (ELA), the pattern of erosion ought to mimic this trend (Anderson *et al.*, 2006). The longer it stays in the ELA, the deeper the glacier-valley tends to be (Sugden and John, 1976). Another major control on valley morphology is the bedrock strength as there is a strong positive correlation between rock mass strength and Fr number (Fr = D/2w), where D is the valley depth and w is the valleys half width. This correlation suggests a trend toward narrower, steeper-sided valleys being developed on bedrock of high mass strength. An increase in the Fr number implies on narrower valley.

2.2. Glaciation and turbidity currents

Glaciers covered most of Gondwana during the Carboniferous and Permian, leaving in the rock record clues of a second major ice age in the Phanerozoic, *i.e.* the Karoo ice age (360-260 Ma). The Panthalassa Ocean and the Paleotethys Sea influenced a progressive cooling of the summer, and higher rates of snow accumulation in winter, causing alpine glaciers to form and spread out to form continental glaciers (Berner, 1999; Beerling and Berner, 2000; Fig.2.3).
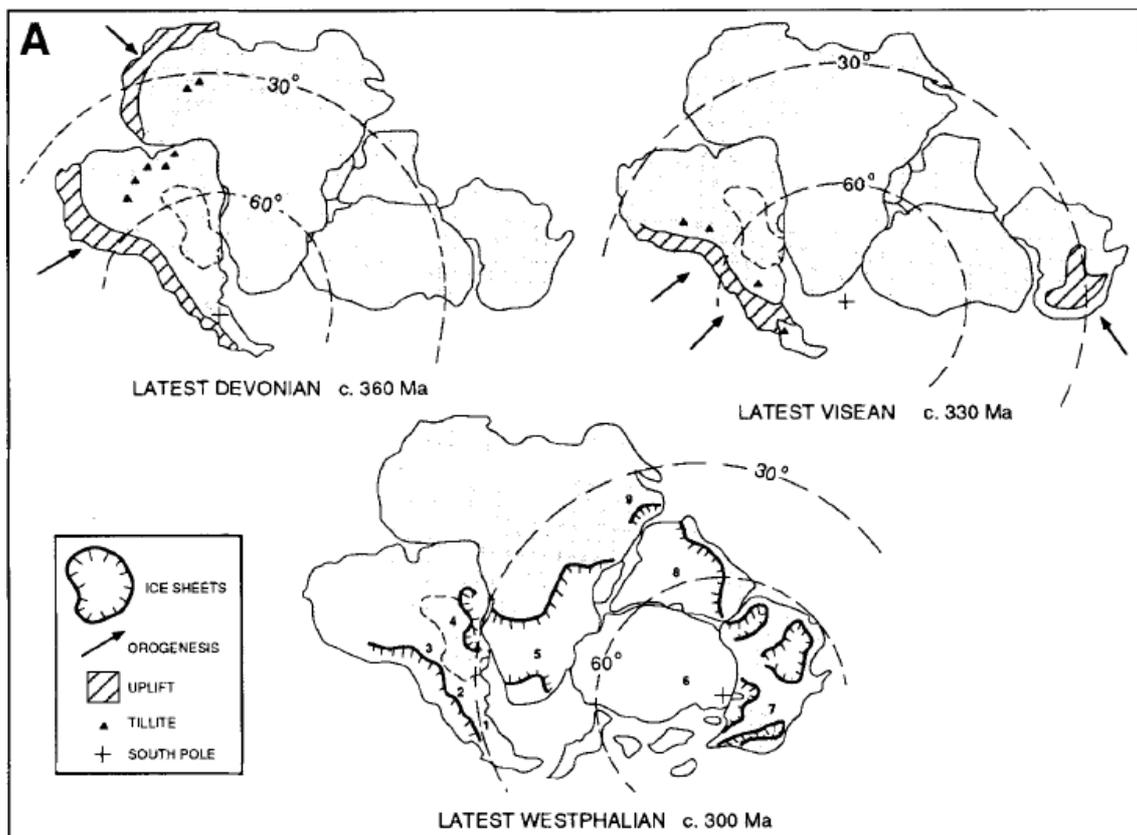
Figure 2.3: Paleogeographical maps of the Gondwana for the latest Devonian, latest Visean and latest Westphalian. Permo-carboniferous glaciated basins are numbered 1-9: 1) Argentina; 2) Paraguay; 3) Bolivia; 4) Paraná Basin; 5) Southern-African; 6) Antarctica; 7) Australia; 9) Indian S C; 9) Arabian P. (modified from Veevers& Powell, 1987).

Glaciers displacement and deposition are well recorded in the Paganzo and Paraná basins, usually as striation on basement rock surfaces due to moving ice, an extensive process that can form fjords and valleys due over-deepening by glacial scour. Such depressions are commonly left under hundreds of meters of water following glacial

retreat and relative sea level rise (Powell and Molnia, 1989;Eyles *et al.*, 1990; DaSilva *et al.*, 1997; Benn and Evans, 1998; Dykstra *et al.* 2007).

Though postglacial rebound can remove a great portion of the accommodation space, these incised valleys are commonly so over-deepened that most of the valley remains below sea level, where sedimentary preservation is usually high (Kneller *et al.*, 2004; Dykstra *et al.*, 2007).

Sediments in this deep fjord and valleys are largely made up of turbidites: they generally consist of sediments eroded from land surface, transported by rivers to the shoreline, and ressedimented into deeper water by turbidity currents (surge-type). Besides, during glaciation turbidity flows were also favoured by seasonal cyclic changes. Spring/summer warmer temperatures can spur melting periods, producing important sediment pulses of very cold and dense, sediment laden, usually hyperpycnal, and sustained flows into confined settings (Prior *et al.*, 1987; Benn and Evans, 1998; Kneller *et al.*, 2004; Dykstra, 2005).

Ancient sandy turbidites tend to be deeply buried and compacted, and form an important class of hydrocarbon reservoirs (Weimeir & Slatt, 2007) or a host rock for a particular type of gold deposit (Keppie *et al.*, 1987). The structure of the turbidity currents can be differentiated into a frontal region (head) that is followed by the body (Fig. 2.4). The motion of the fluid behind the head can be approximated with the modified formula of the Chezy equation for flow in open channels, using reduced gravity (Middleton, 1993), and it is slope-dependent. Thus, the buoyance flux (Turner, 1973) into the head increases with increasing slope, with a concomitant effect on mixing. Finite-volume releases (surge-type currents) may be dominated by the properties of the head (Hacker *et al.*, 1996) in contrast to sustained underflows.

The deposits of turbidity current mostly depend of the behaviour of the particles in terms of spacial and temporal acceleration within the flow (Fig. 2.5). Accepting that when both accelerations are zero, the fluid and particles will be moving in a constant flow, if it has a substantive acceleration the particles will not fall from flow suspension (bypass/erosion). When the fluid starts decelerating, it will eventually reach a suspension threshold forcing the sediment to fall-out (Bouma, 1962).
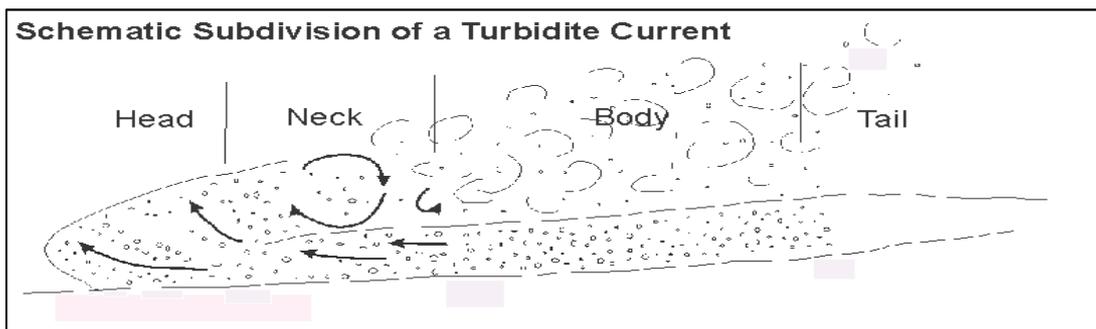
Figure 2.4: Diagram representing a schematic subdivision of a turbidite current.

The acceleration of a turbidite can be subdivided into five possible types of flow, differentiated by the type of acceleration (Fig. 2.6).

- Waxing/Depletive flow: it presents a positive temporal acceleration (waxing) and negative spatial acceleration (depletive) flow. At a specific point the flow becomes faster with time whereas it is slower downstream. At that specific point the grain size material falling from suspension from the flow tends to become larger with time as the current velocity increases, producing inverse graded beds. Eventually, if the flow velocity becomes too high, it will maintain all the sediment in suspension (bypass/erosion).
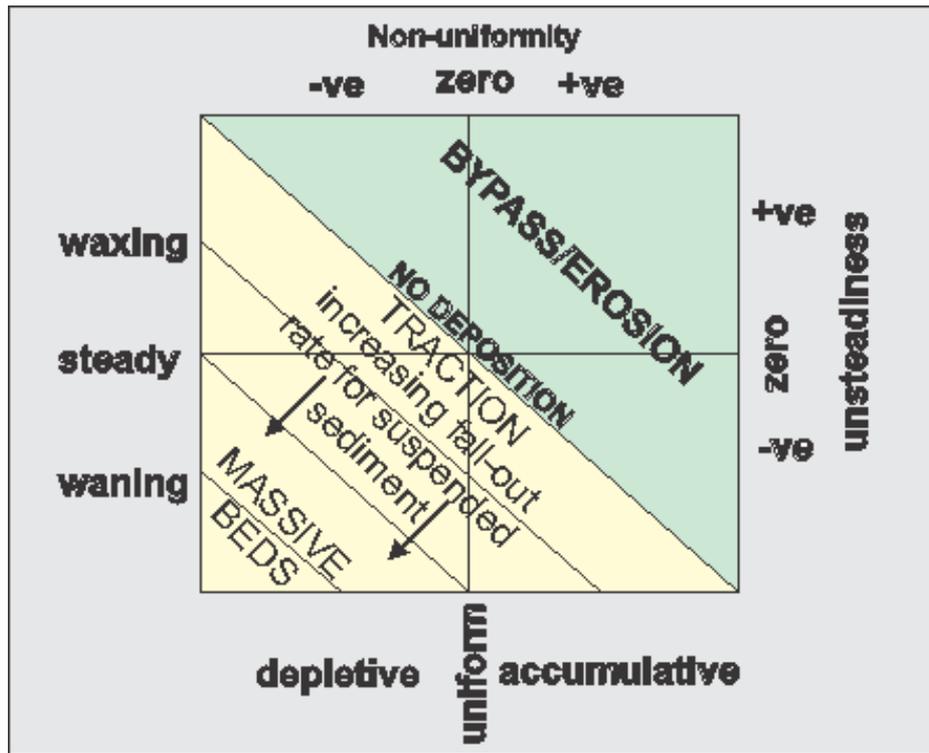
Figure 2.5: Turbidite acceleration graph, representing the type of deposits according to the type of temporal and special acceleration (Kneller, 1995).
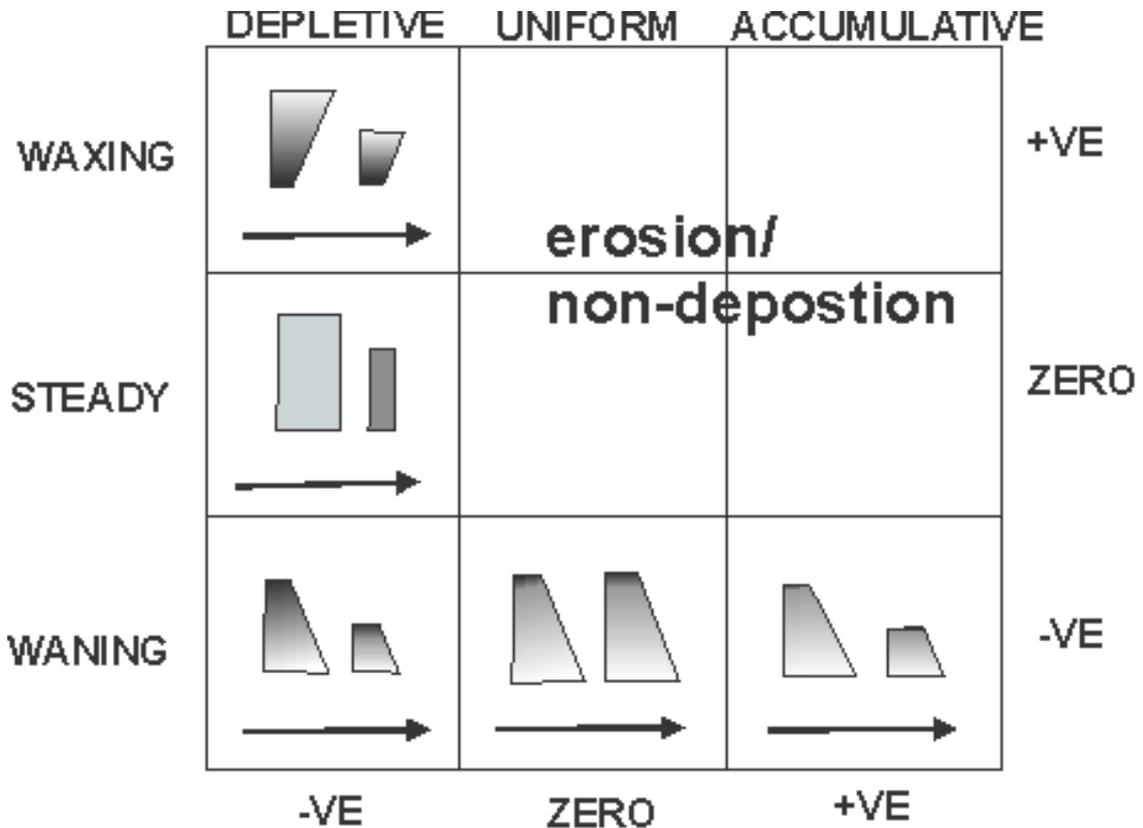
Figure 2.6: Types of turbidite deposits based on the type of flow acceleration (Kneller, 1995)

- Steady/Depletive flow: it shows no variation in temporal acceleration (steady), as the flow has a constant velocity, although the negative spacial acceleration (due confinement) allows particles to decelerate and deposit. This flow tends to deposit sediments from flow suspension so rapidly that produce massive ungraded beds with occasional incorporation of mudclasts at a specific point and finer material downstream.

- Waning/Depletive flow: it displays negative flow acceleration in which both temporal (waning) and spatial (depletive) accelerations are negative. As the flow decelerates it produces graded beds. At an upstream position the sediments will be coarser-grained and the beds thicker, at a downstream position grains will be finer-grained and bed thinner. The higher negative acceleration, the harder will be to produce traction structures.

- Waning/Accumulative flow: it shows a positive spatial acceleration (accumulative) that will give rise to changes in the bed shear stress, but due the waning properties it will generate a graded bed. Altogether the flow will produce a well-sorted graded bed, commonly with flute marks/cast.

In many cases, the development of turbidite systems has been moderately to strongly controlled by pre-existing bounding slopes (Lomas and Joseph, 2004). Van Andel & Komar (1969) used the term *ponding* to refer to a situation in which turbidity currents of sufficiently large volume are fully contained by an area of subdue bathymetry. Similarly, Pickering & Hiscott (1985) used the expression "contained turbidites" to describe beds deposited from turbidity currents that were confined within a basin too small to allow sustained unidirectional flow. These terms describe situations in which gravity flow sediments fill the floor of an enclosed depression and are unable to

surmount the bounding sloped which bound that depositional area. Closely associated with the concept of confined deposition are the notions of flow reflection and deflection and fill-and-spill common features related to turbidity currents confinement, (Pickering and Hiscott, 1985) that can also include:

- presence of thick siltstone/mudstone caps overlying or interbedded with sandstone beds;

- complex multidirectional paleocurrent directions;

- dominance of thick beds;

- complex grading patterns;

- soft-sediment-deformation;

- onlap of sheet-like beds onto confining-basement.

The upper part of confined sedimentary successions tends to present an increase in the sand-to-mud ratio due to flow stripping (Fig 2.7), *i.e.* the denser, sand-rich flow travels along the basin floor eroding and depositing amalgamated sandstones, while the finer-grained material stays in suspension, therefore escaping from confinement (Sinclair and Tomasso, 2002).
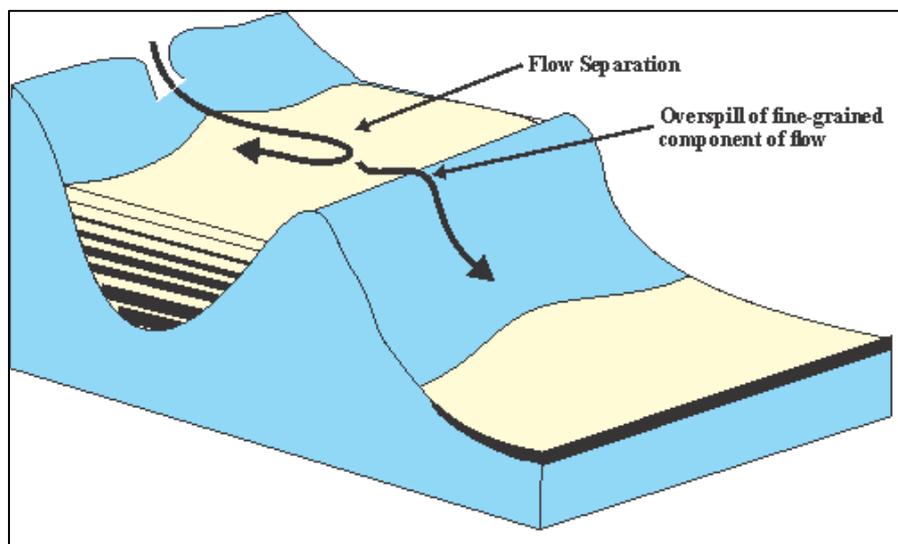
Figure 2.7: Depositional model for the infill of a confined turbidite basin at the flow stripping stage (modified from Sinclair and Tomasso 2002)

2.3 Turbidites Onlap

The development of stratigraphic traps, a key feature of many hydrocarbon reservoirs, is favoured at the margin of confined turbidites sheet systems (*sensu* Haughton, 1994a). According to McCaffrey & Kneller (2001) there are two main types of pinch out (types A and B). In the first case (Type A), turbidites thin onto the confining surface – although the final sandstone pinch-out is commonly abrupt – and individual beds tend not to erode into earlier deposits. In type B, turbidite sandstones commonly thicken toward the confining slope, and bed may incise into earlier deposits. These two types may occur in combination, so producing a wide spectrum of pinch out characteristics (fig. 2.8).

Figure 2.8: Classification of sand body terminations after (a) McCaffrey & Kneller (2001); and (b) Smith &Josep (2004). *Apud* Gardiner (2006).

Sandbody pinchout occur in a range of depositional environments and can take a variety of forms (Fig. 2.9). In the case of glacial-valley, individual beds may preserve their thickness, but usually as they approach the onlap surface sandbody thickness tends to decrease. (Gardiner, 2006). In some cases, as sand beds thin at the onlap surface, they continue up the surface for some distance (McCaffrey & Kneller, 2006; Smith & Joseph, 2004), often with a thickness of only a few cm (Fig. 2.9 b, d).
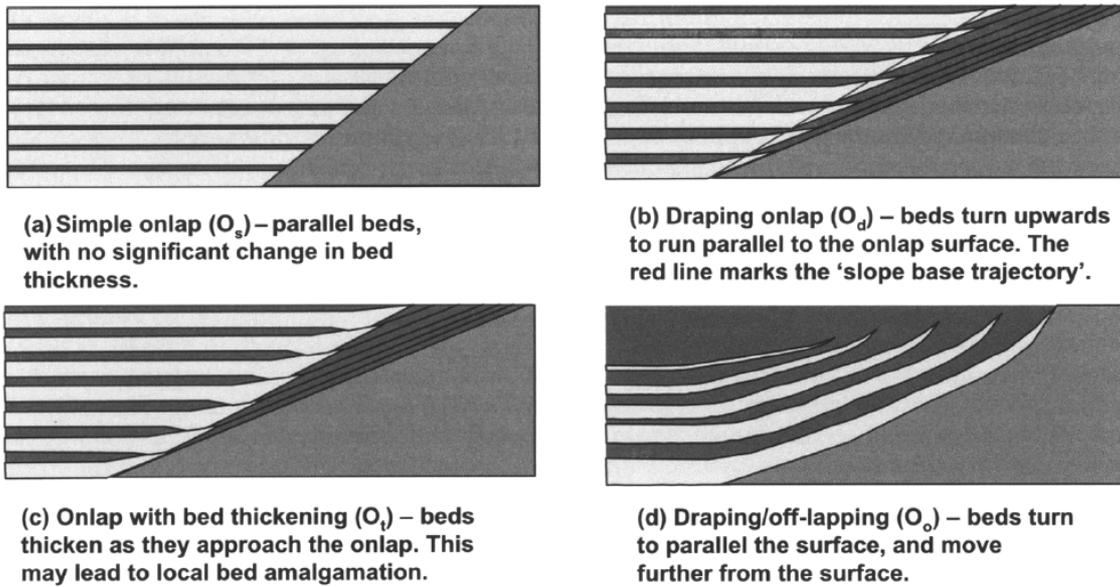
Figure 2.9: Examples of sandbody pinchout geometry, (a-d) show a range of possible geometries when sandstone body onlaps an existing surface (modified from Gardiner, 2006).

2.4. The Paganzo Basin

Western Argentina is an area of recent and constant deformation resultant from the convergence of the Nazca and South American plate, reflected in modern GPS velocity data and constant seismicity in the region, including large-magnitude, catastrophic seismic events (Brooks *et al.*, 2003; Costa and Finzi, 1996; Perucca and Paredes, 2003; Ramos *et al.*, 2002; Siame *et al.*, 2002). Recent uplift and arid climate favour excellent rock exposure in the Paganzo Basin, one of the largest (140,000 Km^2) upper Palaeozoic depositional areas documented along the western margin of Gondwana (Saltify and Gorustovich, 1983; López Gamundí at al., 1994). The Paganzo Basin is bounded to the east by the Sierras-Pampeanas cratonic area and to the west by an orogenic belt composed of thrusted and folded, Lower Palaeozoic and Lower Carboniferous units (Limarino *et al.*, 2002).

This pericratonic basin received glacial-derived sediments during and just after the Late Carboniferous glaciation of Gondwana (Kneller *et al.*, 2004; Eyeles *et al.*, 1995). It lied close to the active western margin of the supercontinent near paleolatitude 60ºS (Fig 2.3). Its tectonic setting has been ascribed either to a retro-arc foreland basin (Ramos, 1998) or to a strike-slip basin (Fernández Seveso & Tankard, 1995). Paganzo Basin was isolated from a coeval, open marine basin to the west (Calingasta-Uspallata and Rio Blanco basins) by one or more coastal ranges (e.g. Salfity & Gorustovich, 1983) uplifted in the Devonian and geographically roughly equivalent to the modern Precordillera (Fig. 2.10).

Postglacial, early-Namurian to Westphalian transgressive strata have been identified in several localities (Limarino *et al.*, 2002). In the western and central domains of the Precordillera, transgression is recorded by open marine facies and transitional fjord type environments (López Gamundí, 1987). The Precordillera represents the transition between open marine environments (western domain) and thin flooding facies typical of the Sierra Pampeanas (eastern domain). This transgression is well exposed in at least five localities: Huaco, Rio Francia, San Juan River Valley, Guandacol (Sierra de Maz), Cerro Bola, Quebrada de Las Lajas and Quebrada Grande and Cuesta de Huaco.
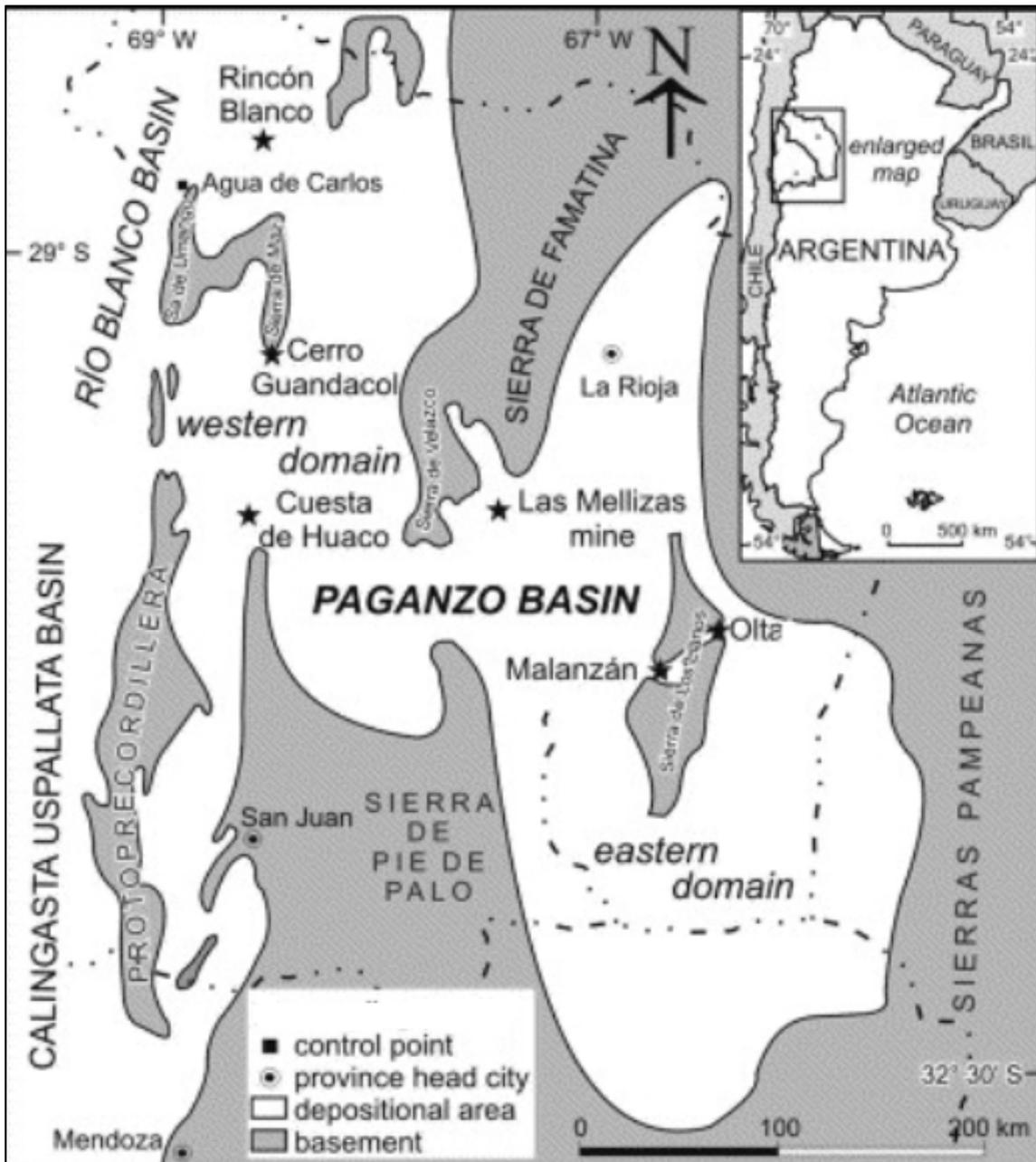
Figure 2.10: Late Carboniferous (Pennsylvanian) paleogeography of the Paganzo and adjacent basins, modified from Buatois and Mángano (1995).

The palaeovalleys situated along the western Precordillera were apparently connected to the open ocean during the Carboniferous transgression, which is recorded in the lower portion of the Jejenes Formation and characterized by gravity-flow deposits and shales with marine fossils (López Gamundí, 1983, 1989), spores and monosacate pollens, subordinated Namurian acritarchs and marine microfossils represented by abundant *Prasinoficeae* algae and few *Veryhachyum* specimens (Césari and Bercowski, 1998).

These palaeovalleys also show clear evidence of glacial erosion, with well-preserved striations and basal glacial pavements (Milana and Bercowski, 1990, 1993; López-Gamundí and Martínez, 2000; Kneller *et al.*, 2004; Dykstra *et al.*, 2007).

The Quebrada de Las Lajas, situated near San Juan, in the western Argentina (Fig. 1.1), represents one of these palaeovalleys and comprises one of the study areas. This modern valley exhumes a deeply incised, Mississippian palaeofjord developed on the eastern side of the Andean proto-Precordillera. Carved during the Gondwana glaciation, cutting over 1000 m down into Lower Palaeozoic shelf carbonates of the San Juan Formation (Kneller *et al.*, 2004; Dykstra, 2005; Dykstra *et al.*, 2007), it was later filled with a siliciclastic succession known as the Jejenes Formation.

Fjord basement was probably already dipping steeply by the Carboniferous. It was deformed during a Devonian accretion event and incorporated into a mountain range at the time (Beresi and Bordonaro, 1985). The unconformity between the San Juan and Jejenes formations displays a geometry that is similar to the modern glacial-valleys (Fig. 2.11), *i.e.* U-shaped with steeply dipping walls (60º-70º) that extend more than 1,000m above the valley floor (Dykstra, 2005; Dykstra *et al.*, 2007; Fig. 2.10).
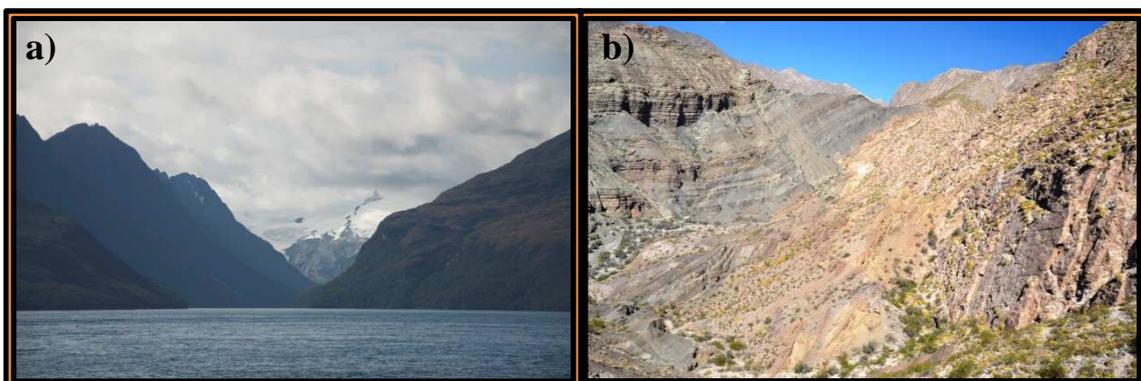
Figure 2.11: Two photos taken from carved fjords: a) a modern fjord, Chile (Milbre, 2011) and b) a well exposed palaeofjord filled with Carboniferous turbidites in the Quebrada de Las Lajas, (Carvalho, 2012)

Abundant fossil and palynological evidence demonstrates that marine transgression progressed up into the inner mountains valleys such as the Quebrada de Las Lajas (Martinez, 1991; Césari and Berkowski, 1997; Lech *et al.*, 1998; Limarino *et al.*, 2002; Pazos, 2002; Kneller *et al.*, 2004; Dykstra *et al.*, 2007).

The QLL palaeofjord is now oriented to SW-NE and contains Pennsylvanian clastic strata preserved over about 5 km along the valley with total thickness close to 297 m. Its width varies from ~200 m at its head up to 1 km at its mouth (Fig. 2.12). The thickness of marine deposits is greater than the magnitude of a eustatic sea-level rise, suggesting a significant over-deepening of the palaeofjord below sea level by glacial erosion prior to glacial collapse and flooding (Dykstra *et al.*, 2006). Several sills are present in the base of the palaeofjord and at least three hanging valleys with Carboniferous fill are present along its margins (Dykstra *et al.*, 2007).

According to Dykstra (2006) the sedimentary succession that fills the QLL palaeofjord can be subdivided into four main evolutionary stages (Fig 2.12) represented by distinct facies associations:

- Stage I, an ice-contact facies association related to subaqueous outwash and subaqueous fans, and deep lacustrine strata including a channel levee system and mass-transport deposits (MTD).
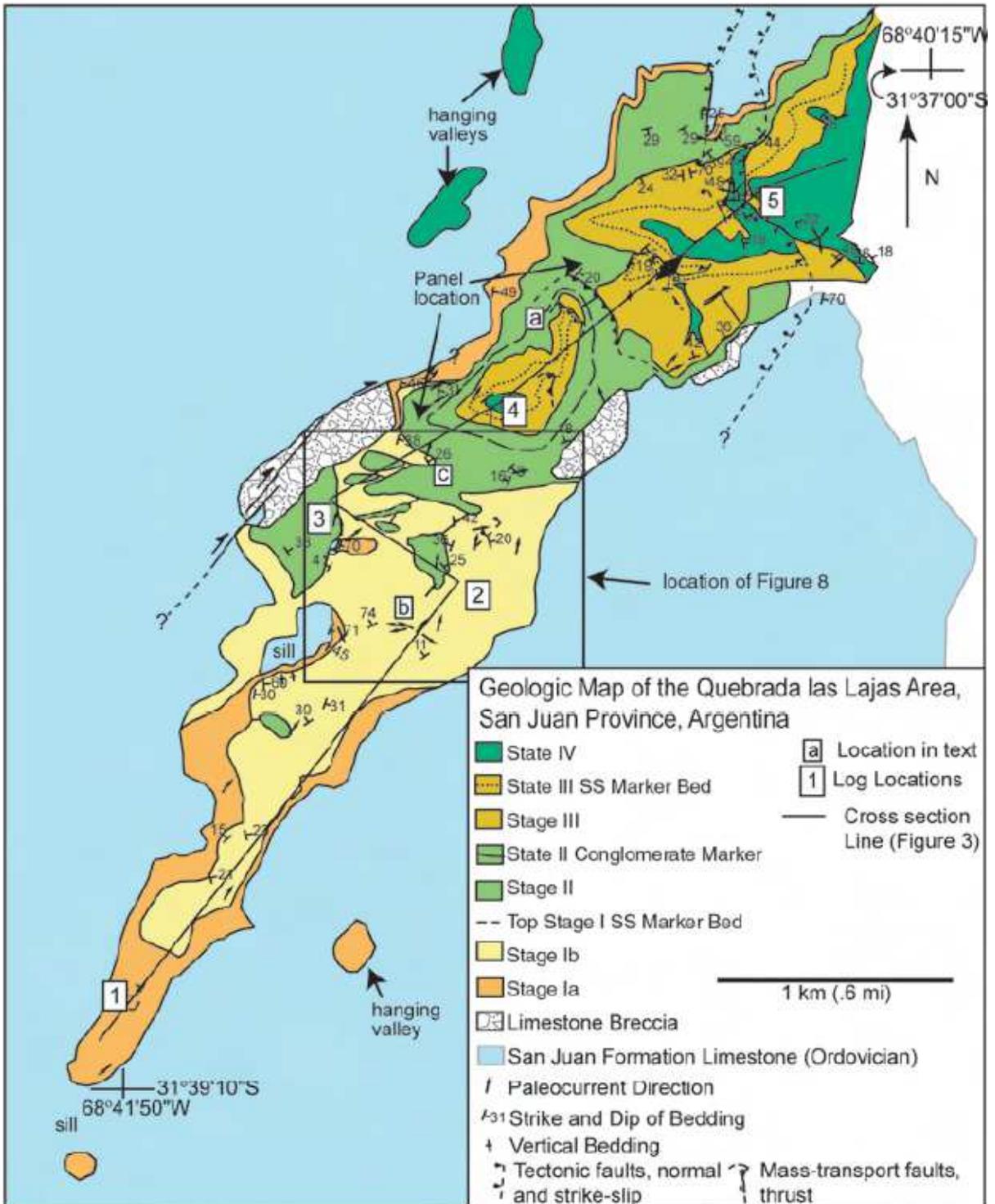
Figure 2.12: Quebrada de las Lajas geological map (Dykstra, 2006)

- Stage II, a postglacial, marine transgressive succession including basal, chaotic MTDs and associated ponded channels, few slope channels (channel bodies incised into black shales), and coherent, slide-sheet-dominated MTDs.

- Stage III, which encompasses sheet-like, sandy turbidites (the main target of this dissertation).

- Stage IV, coarse grained, delta sheet sandstones and conglomerates and abundant, small-scale MTDs.

The Quebrada Grande (QG) is situated 13 km to the south of QLL and about 25 Km from San Juan, on the eastern margin of the Zonda range, in the foothills of the eastern Precordillera in western Argentina (Fig 2.13). The area lies on the foreland side of a Quaternary back-thrusting zone (the Eastern Precordillera), with practically no structural complications. Like the QLL, it also exposes the filling of a palaeovalley and comprises the second study area of the Paganzo Basin.

The succession (Jejenes Formation) exposed in the Quebrada Grande dips (35º-40º to ESE) transversely to the roughly NNE-SSW oriented valley axis (Fig. 2.12). The Quebrada Grande area is extensively dissected by modern valleys and creeks, providing superb sections through parts of the palaeovalley fill (Fig. 2.12). The local subcrop to the Carboniferous deposits consists of Ordovician platform limestones of the San Juan Formation (Berese and Bordonaro, 1985), stratigraphically overlain to the east by an Ordovician-Silurian olistostrome (Rinconada Formation; Amos, 1954; Peralta, 1990; Peralta *et al.*, 1994) composed of highly deformed, Ordovician-Silurian siliciclastic turbidites and olistoliths of the San Juan Formation (Kneller *et al.*, 2004). Both lower Palaeozoic units and their boundary dip very steeply to the east. The Carboniferous Jejenes Formation crops out as a ~1km wide strip that is unconformably overlain to the east by Neogene and Quaternary deposits (Fig 2.14).
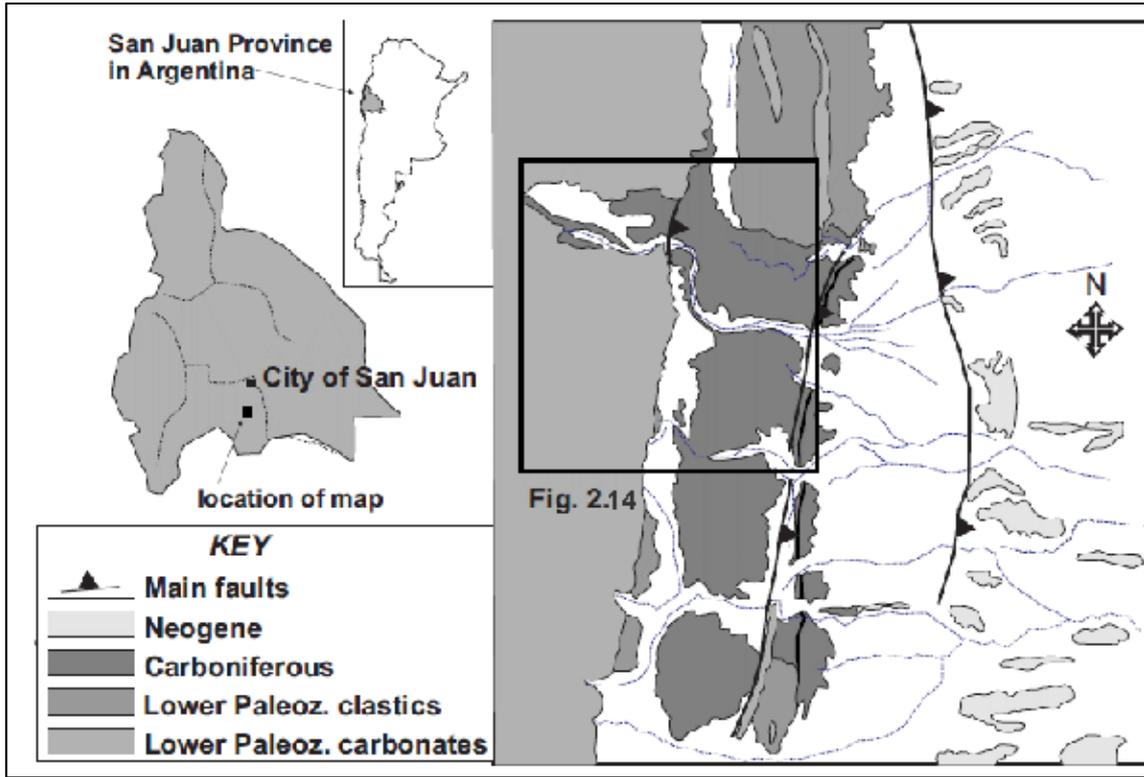
Figure 2.13: Location of the Quebrada Grande and outline of local geology (modified from Kneller, 2004). Carboniferous strata (dark grey) represents depositional areas situated to the east of the Protoprecordillera (medium grey area). Rectangle in the right hand side map shows approximate location of figure 2.14.

The valley-fill was subdivided by Kneller (2004) into five stages (Figs. 2.14; 2.15):

· Stage I: Glacial deposits comprising massive, stratified diamictites and matrix-supported conglomerates;

· Stage II: Shallow-marine strata represented by water-laid sandstone, including shallow-water delta facies;

· Stage III: Proglacial, deep-water deposits, including gravel-rich submarine fans and dropstone-bearing basin-floor mudstones with few, thin-bedded turbidites and two mega-beds (Kneller, 2004; Rodrigues, in prep);

· Stage IV: Confined turbidite sand sheets (the focus of this thesis);

· Stage V: Mass flow deposits and turbidite-channelized complexes associated with the development of Gilbert-type deltas.

Figure 2.14: Geologic map of the Quebrada Grande. The north sector is modified from Kneller (2004) whereas the central and southern areas were mapped during this work. Location shown in fig. 2.13.

Geologic map of the Quebrada Grande region,
San Juan Province, Argentina

Stage III
- T_2 tuffite
- T_1 tuffite
- GM marker
- Slump unit below upper megabed
- T_6 tuffite
- Upper megabed
- Conglomerate CGT_2
- Lower megabed
- Slumped base of lower megabed
- Conglomerate CGT_1
- Lower profan shales with thin sands
- Mainly Shales
- Sandy conglomerate

Stages I & II
- Finely Laminated sands
- Diamictite with channels and laminations
- Massive Diamictite

- Olistostrome   Rinconada Formation : Silurian
- Limestone   San Juan Formation : Ordovician

- Alluvium (Quaternary)

Stage V
- Ripple laminated sandstones
- Pebbly sandy mudstone
- Thin/medium bedded turbidite sands and shales
- Slumped/chaotic turbidites
- Thick-bedded massive turbidite sandstone
- Conglomerates

Stage IV
- Turbidite sandstone unit Ts3
- Turbidite sandstone unit Ts2
- Turbidite sandstone unit Ts1

Figure 2.15: Fence diagram showing thickness changes within the northern Quebrada Grande palaeovalley fill (Kneller, 2004)

2.5. Paraná Basin

The Paraná Basin (Fig. 2.16), named after the Paraná River (Eyles *et al.*, 1993) and one of the largest intracratonic basin of South America, is situated in the central-eastern part of the continent (Barberena *et al.*, 1991). It is present from the Mato Grosso do Sul to the Rio Grande do Sul state, in Brazil, and also in eastern Paraguay, Argentina and northern Uruguay. The shape of the depression is elliptical and covers about 1,500,000 km^2 (Souza *et al.*, 2007; Milani *et al.*, 2007), with 2/3 of its surface covered by Mesozoic basaltic lavas of the Serra Geral Formation (Melo, 1988; Zalán *et al.*, 1991; Souza *et al.*, 2007).

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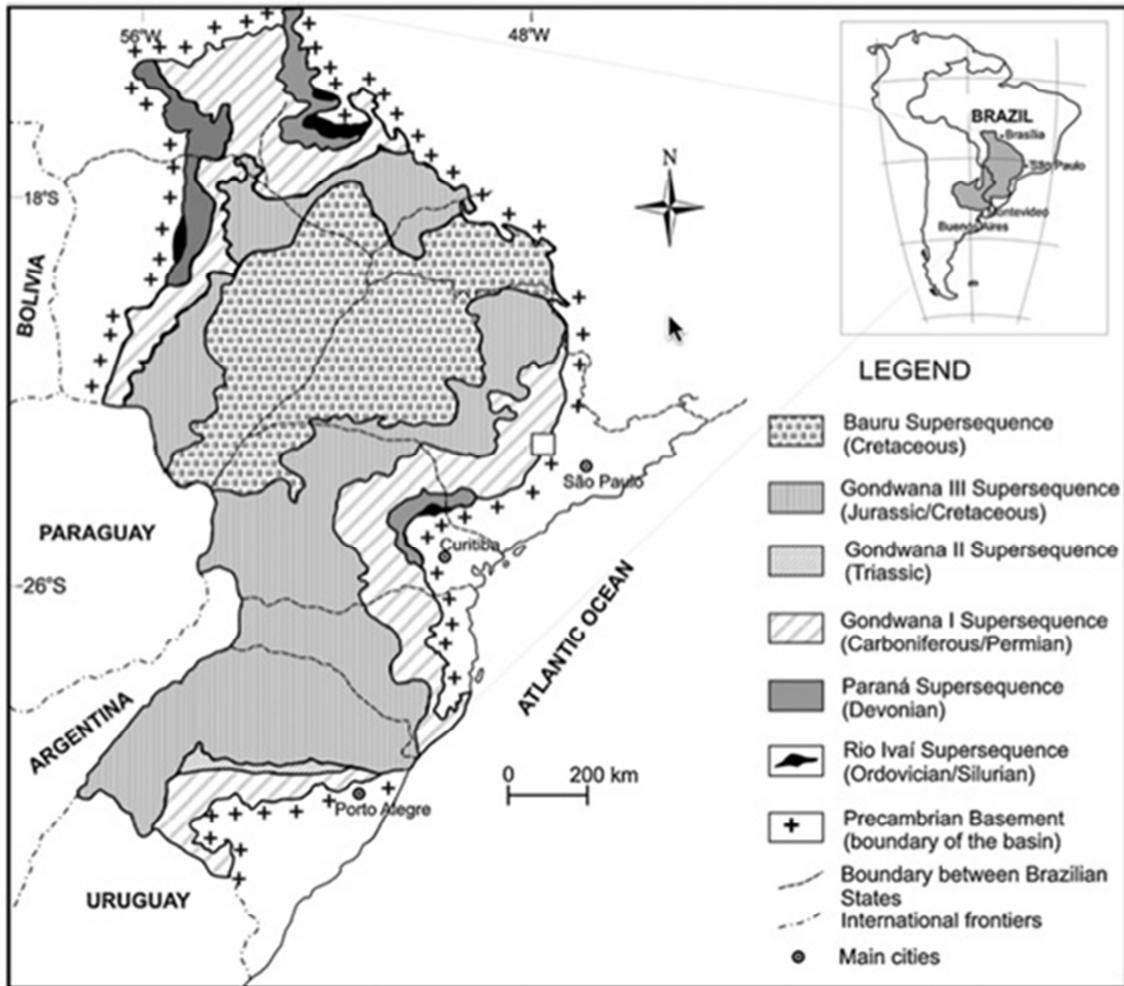
Figure 2.16: Geologic map of the Parana Basin in Brazil, and its location and extension in South America.

The eastern flank of the Paraná Basin corresponds to a crustal region deeply affected by the South Atlantic rifting and drift developed on continental crust of the former Gondwana supercontinent (Milani and Zalán, 1999). The Precambrian basement that underlies the volcano-sedimentary succession is complex and highly structured (Fig. 2.16). It consists of cratonic nuclei bounded by upper Proterozoic thrust and fold belts composed of metasedimentary and metavolcanic rocks intruded by granites (Eyles & Eyles, 1993; Fig 2.17).
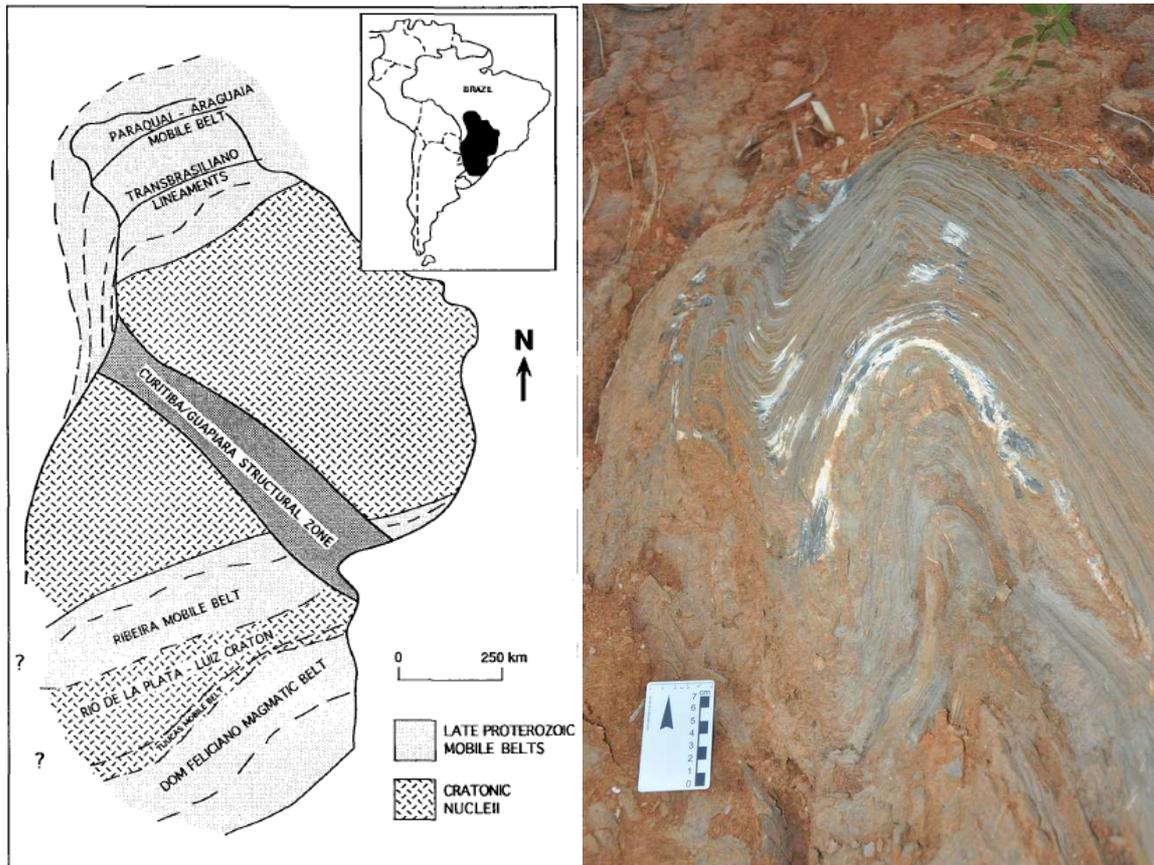
Figure 2.17: Geological provinces (left) and one example (right) of the Paraná Basin basement rock.

The stratigraphy of the Paraná Basin comprises a total thickness over 6 km and consists essentially of six sequences (Milani, 1993) ranging from Ordovician to Late Cretaceous (Fig 2.18) that record distinct episodes of tectonic subsidence and sediment preservation within the basin.

The stratigraphic interval directly related to this thesis is the Gondwana I, a Late Carboniferous to Early Triassic Sequence that includes the glaciogenic Itararé Group. Together with similar deposits of the Carnarvon Basin (Lyons Group, 2500m thick) in Western Australia (Veevers & Powell, 1987), the Itararé Group holds one of the thickest accumulations of late Palaeozoic, glacially-influenced strata. It reaches up to 1300 m in thickness (França & Potter, 1988) in the central-western portion of the basin and comprises three lithostratigraphic units (Milani, 1997, 2005; Bartoreli, 2005; Marques

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and Ernesto, 2005): the Campo do Tenente, Mafra and Rio do Sul formations
(Schneider *et al.*, 1974), roughly equivalent to the subsurface-based units proposed by
França (1987) and França & Potter (1988, 1991), namely the Lagoa Azul, Campo
Mourão and Taciba formations, respectively (Fig. 2.18). Despite of the data source,
each formation comprises a lower, coarser-grained (sandstone to conglomerate) and an
upper, diamictite-dominated unit (Fig. 2.19).
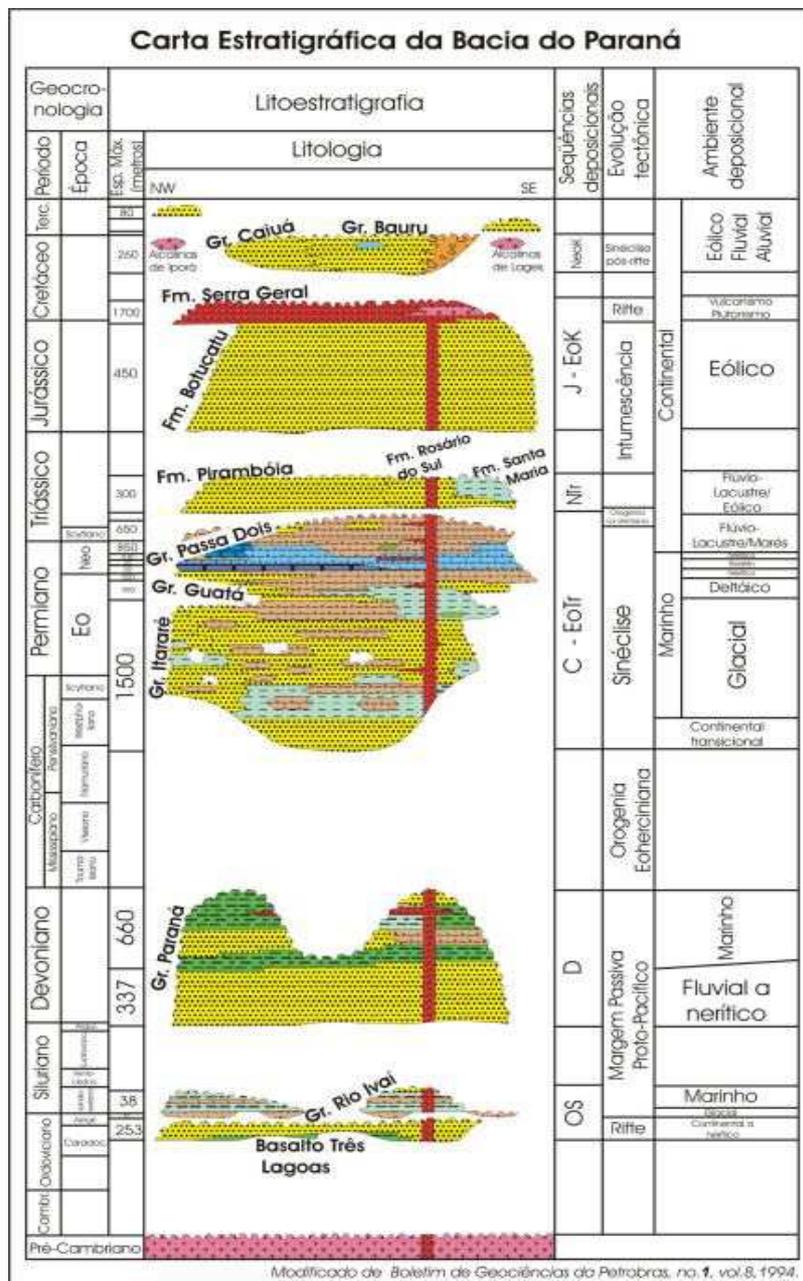
Figure 2.18: Depositional sequences and lithostratigraphy of the
Paraná Basin (Milani, 1993).

Schneider et al. (1974)	França & Potter (1988)	Weinschütz (2001)	Seqüências
Fm, Rio do Sul (RS)	Ciclo/Fm. Taciba	Sup. (s): Siltito e folhelho	Seqüência Rio do Sul s/m
		Méd. (m): diamictito, arenito e siltito	
Fm. Mafra (M)	Ciclo/Fm. Campo Mourão	Inf. (I): folhelho "Lontras"	Seqüência Rio do Sul i/Mafra s
		Sup. (s): arenito e diamictito	
		Méd. (m): varvito, diamictito e arenito	Seqüência Mafra m/i
		Inf. (I): arenito e diamictito	
Fm. Campo do Tenente (CT)	Ciclo/Fm. Lagoa Azul	Folhelho, varvito e diamictito marron-avermelhado	Seqüência Campo do Tenente

Figure 2.19: Stratigraphic correlation of the Itararé Group, with depositional sequences (Weinscthuz and Castro, 2004)
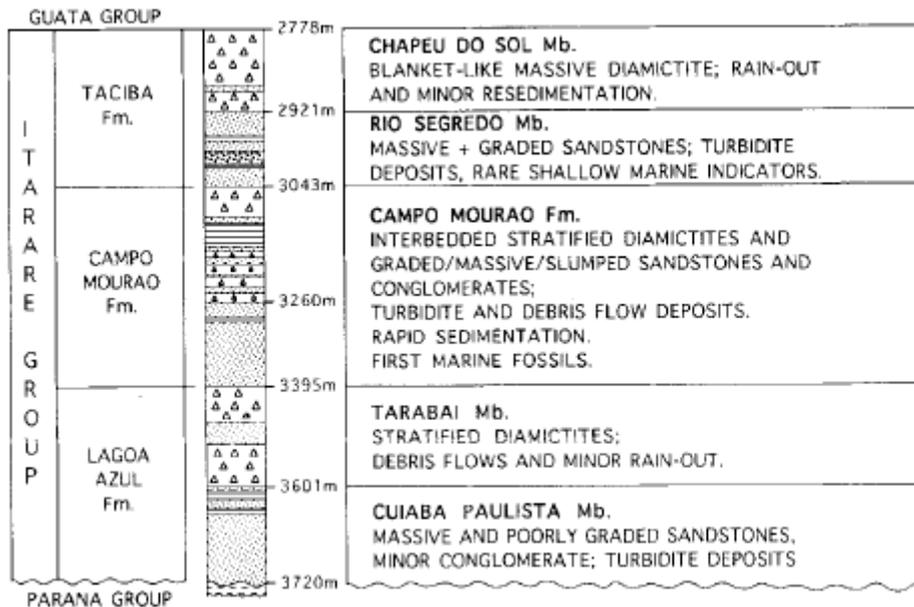
Figure 2.20: Subsurface stratigraphy and associated lithological column of the Itararé Group according to Franca &Potter, 1988.

Although the oldest Itararé strata record glacial-lacustrine or brackish water settings, an increasingly marine influence can be identified upwards through the Itararé Group (Eyles *et al.*, 1993). Fully marine conditions are recorded in the overlying deltaic deposits related to the top of Taciba Formation (Puigdomenech, 2013).

The Itararé Group is poorly constrained relative to other Gondwana glacial successions and there is no firm ground, as yet, for establishing the Carboniferous-Permian boundary within the Paraná Basin (Rocha-Campos and Rösler, 1978). Preliminary findings using revised scheme suggest that the Itararé Group ranges from the

latest Westphalian to Early Kungurian, so it is essentially Permian (Daemon, 1991). The late start of the glaciation in the Paraná Basin compared to adjacent basins in Bolivia, Argentina and Paraguay may simple reflect a different tectonic regime in southern Brazil: glacial strata only began to accumulate in the Paraná Basin after 300 Ma with the onset of tectonic subsidence (Eyles *et al.*, 1993). A long period of glacial erosion in the basin would explain both the lack of depositional record and the existence of glacially carved valleys. At this time, the major glacial influence on sedimentation in southern Brazil was from the Kaokoveld area of southern Africa, such glaciers coming from southwest Africa and reaching the westernmost portions of the Paraná Basin (Barbosa, 1940; Gesicki *et al.*, 2001). Glacial abrasion features carved onto Precambrian basement rocks (Barbosa, 1940; Carvalho, 1940; Dequech, 1948; Canuto, 1993) are common and indicate a north-north-westward sense of ice movement. Besides, intraformational striated surfaces scoured on sandstone and diamictites were found in Cachoeira do Sul, Pinheiro Machado (Tomazelli and Soliani, 1982) and Suspiro (Tomazelli and Soliani, 1997), as well as in other areas of the south of Brazil and São Paulo (Caetano-Chang *et al.*, 1990). More recently, a striated surface developed on glacial deposits of the Aquidabán Formation, which is comparable to the Itararé Group, was found in the village of Escobar, Eastern Paraguay (Riccomini and Velázquez, 1999). In the Santa Catarina state, iceberg scours, dump structures, striations as well as debris left by grounded icebergs on rhythmites, probably generated during the last, early Permian deglaciation episode of the Paraná Basin, were also described (Canuto, 1991; Santos *et al.*, 1992). The orientation of the subglacial abrasion features ranges from NW to NNE, but it is dominantly oriented to NNW, revealing that the southeastern terrains of the Paraná Basin where the main glacier source area during most of the Itararé Group deposition (Santos *et al.*, 1992).

The study area in Brazil represents a relatively wide paleovalley exposed in the Vidal Ramos (VR) county, central portion of the Santa Catarina State (Fig. 1.2). It is poorly exposed due to extensive and dense vegetation cover, and deep weathering. There is almost no structural complication, apart from the diabase intrusive rocks (Serra Geral Formation) that locally dislocate the strata from their original position. The local subcrop to the Permian-carboniferous package consists of Pre-Cambrian gneiss, marble and meta-sedimentary rocks that were originally carved by glaciers producing a striated surface that displays evidence of north-northwestward (345º) sense of ice movement. The palaeovalley has an average width of 12 km and walls that dip 10º to 15º. Its fill includes a thin succession of the Mafra Formation, that crop out along a~2km wide strip. Once the palaeovalley was filled (Mafra Formation), deposition of the Rio do Sul formation took place in a non-confined setting including deep marine to deltaic context (Puigdomenech, in prep) (Fig 2.21).

Stratigraphic Column
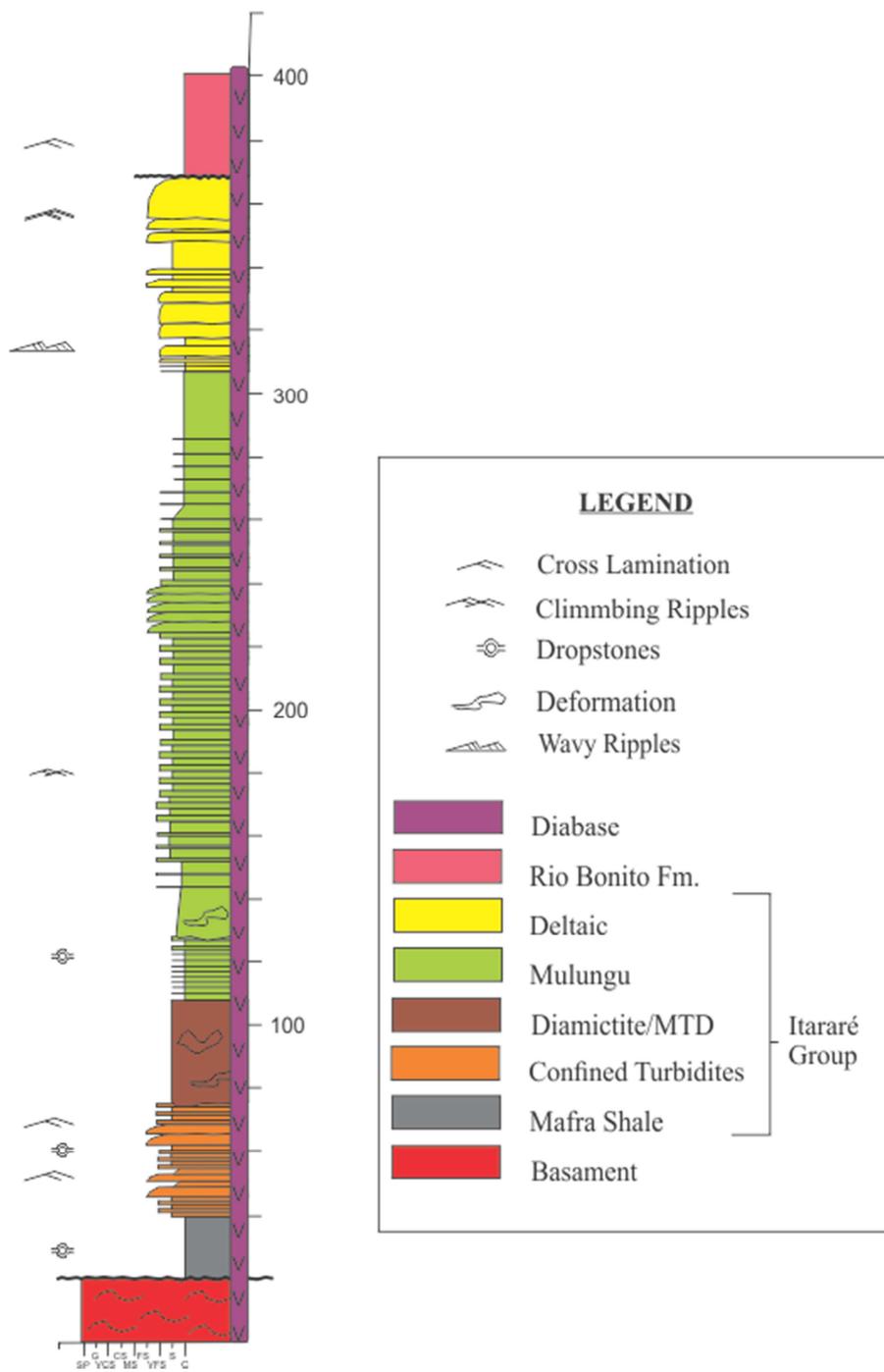
Figure 2.21: General sedimentary log of Vidal Ramos

CHAPTER 3 – RESULTS, DISCUSSION AND CONCLUSION

This chapter presents most of the results, their discussion and the conclusions of the dissertation, It is organized as paper submitted to the Journal of South American Earth Sciences. Therefore, although minimized as much as possible, some duplication in terms of objectives, methodology, scientific background and geological setting were unavoidable. But in general the paper presents the results, discussion and conclusions and the previous chapter discuss further subjects that are important in terms of a dissertation, but to long to be included in a paper. Below, the confirmation of the submission of the article.

THE ROLE OF CONFINEMENT ON TURBIDITE SAND-SHEET DEVELOPMENT: A COMPARISON OF THREE CASE STUDIES FROM PAGANZO AND PARANÁ PALAEOVALLEYS

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ABSTRACT

During the Late Paleozoic, Gondwana was home for a protracted glaciation responsible for high rates of sediment supply and generation of accommodation space within laterally confined, longitudinally extensive glacial valleys and fjords. These features became so overdeepened below sea level that they filled up with hundreds of meters of seawater, as glacio-eustatic sea level rose during glaciation and early deglaciation. Deep-water environments and high sediment supply thus favored gravity-flows, including large-scale mass transport, debris flows and turbidity currents. This study assesses the morphological setting, sedimentology and architecture of the sand sheets deposited by turbidity currents that instead of flowing onto an unconfined basin floor were confined within glacial palaeotopography. In this context, this paper describes and compares three turbidite systems deposited into settings presenting different degrees of confinement. The main aim was to identify some sort of signature on turbidite strata that could be correlated to their confined depositional settings. This study was based on three outcrop case studies, two in Argentina (Paganzo basin) and one in Brazil (Paraná basin). The Argentinean areas, located in the eastern Andean Precordillera, near San Juan city, comprise a palaeofjord (Las Lajas) and a more open valley system (Quebrada Grande), both incised into Early Paleozoic shelf carbonates (San Juan Formation) and olistostrome (Rinconada Formation). The Brazilian case study in the roughly coeval

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Paraná basin is situated around Vidal Ramos where basement depressions produced by glacial carving into upper Proterozoic schists (secondly marbles) were firstly filled with deep marine shales and turbidites. Long mapping and logging field seasons in all areas, with the help of laser range finders, to delineate surfaces in inaccessible sites, and 3D photomosaics, to interpret facies geometry, comprised the adopted approach in terms of data acquisition. From mapping and logging, the extent of confinement (morphological models) was characterized and conceptual models integrating sequence stratigraphy, sedimentology and valley morphology were proposed. The deglacial sedimentary fill of glacial palaeovalleys, in special the associated turbidite sand sheets, includes features both independent and dependent of the confinement extent. The independent features occurs in all studied sites and include high frequency climate changes controlling vertical facies successions, turbidity currents flowing mostly along the valley axis following a unidirectional trend and an assumed association with fjord- or valley-head deltas prograding into relatively steep valley floor, hence favouring recurrent sand remobilization from delta front to deep marine settings as turbidity, depletive surges. On the other hand, the higher is the confinement, more complex are paleocurrent trend and facies association due to current rebound of the flows and more common large-scale remobilization near the confining walls. Besides, facies succession suggest that the larger the confinement, the higher the possibility of the existence of effective longitudinal barriers trapping mud in more proximal depositional settings.

Key words: Confined turbidites; turbidite sand sheet; deglacial strata; Paraná basin; Paganzo basin; Gondwana glacial record.

1. INTRODUCTION

Deglacial marine sedimentary sequences record the melting retreat of glaciers. Along the Late Palaeozoic, glaciation that affected Gondwana over one hundred million years, glaciogenic deposits can be found in all present day continents formerly assembled in the Gondwanaland (Crowell, 1983).

Fjords (long, narrow inlets with steep sides or cliffs) and valleys can provide especially detailed records of environmental changes associated with deglaciation due to the large amount of accommodation space available within these glacial, overdeepened features. Their typical high deposition rates provide a high-resolution record (e.g. Eyles *et al.*, 1997; Kneller *et al.*, 2004). These local basins were overdeepened during glacial excavation, and commonly left under hundreds of meters of seawater following glacial retreat and relative sea level rise (e.g. Powell and Molnia, 1989; Eyles *et al.*, 1990; DaSilva *et al.*, 1997; Benn and Evans, 1998). Also, high sedimentation rates during and following deglaciation can emplace hundreds of meters of stratigraphy in relatively short time spans (hundreds to thousands of years), therefore enhancing resolution of their sedimentary records (e.g., Cowan *et al.*, 1988; Powell and Molnia, 1989; Eyles *et al.*, 1990; Lysa *et al.*, 2004).

Deposition tends to respond to seasonal cycles, as snowmelt, rainy seasons or seasonal thawing of a valley glacier (Dykstra *et al.*, 2007). Spring or summer melts periods can be responsible for the greatest sediment pulses. Sediment failures may also contribute significantly accumulations in these confined settings. Commonly produced from delta-fronts, usually related to Gilbert-type deltas, or rock fall off the steep fjord walls, most gravity flows can evolve into turbidity currents upon entrainment of ambient water and flow for very long distances down to local basins (Prior *et al.*, 1987; Benn and Evans, 1998; Kneller *et al.*, 2004; Dykstra, 2005).

Besides, most of the early conceptual and experimental models of turbidity systems have pictured essentially unconfined, radial lobes and fan-shape depositional bodies (Lomas & Joseph, 2004). However, Andel & Komar (1969) proposed the term ponding to describe turbidity currents fully contained by an area of enclosed bathymetry and unable to surmount the bounding slopes. Lomas & Joseph (2004) suggested the use of a more general term (confinement) to describe all cases where sediment gravity flows and their deposits are strongly affected by the presence of significant basin floor topography, not necessarily implying in an entire containment. Closely associated with this concept (confinement) is the idea of flow reflection and deflection, usually deduced from reversals on paleocurrent indicators assumed as a border effect (e.g. Pickering & Hiscott, 1985; Kneller et al. 1991). Onlap relationship with steeper surfaces is also an indicative of confinement. Moreover, the recognition of stratal onlap against a confining frontal slope is understood as a diagnostic feature of the fill of a true silled sub-basin. In terms of confined settings, the most frequent causes include large-scale structural features, slope collapse masses, halokinesis, mud diapirism and differential compaction (Lomas & Joseph 2004).

Within this context, this paper aims to describe and to compare three roughly coeval deglacial turbidite sand sheets deposited in a broad range of confined valleys carved by glaciers. The idea is to test the possible influence of different degree of confinement on turbidite facies, facies association, succession and geometry, pattern of stratal onlap, sand connectivity and sand-to-mud ratio as well as to compare the deglacial record of two large South American basins (Paraná and Paganzo).

2. GEOLOGICAL SETTING

This study took place in two correlative basins (Parana and Paganzo basins). The Paganzo basin is a pericratonic basin that received glacial-derived sediments during and just after the Late Carboniferous glaciation of Gondwana (Salfity and Gorustovich, 1983; Fernandez Seveso and Tankard, 1995; Kneller *et al.*, 2004; Eyeles *et al.*, 1995). It was formed close to the active western margin of the supercontinent near paleolatitude 60ºS. An area of recent and constant deformation resultant from the convergence of the Nazca and South American plate, reflected in modern GPS velocity data and constant seismicity in the region, including large-magnitude, catastrophic seismic events (Brooks *et al.*, 2003; Costa and Finzi, 1996; Perucca and Paredes, 2003; Ramos *et al.*, 2002; Siame *et al.*, 2002). Recent uplift and arid climate favour excellent rock exposure in the Paganzo Basin, one of the largest (140,000 Km^2) upper Palaeozoic depositional areas documented along the western margin of Gondwana (Saltify and Gorustovich, 1983; López Gamundí at al., 1994). The Precordillera was apparently connected to the open ocean during the Carboniferous transgression, which is recorded in the lower portion of Jejenes and correlative Formations and characterized by gravity-flow deposits and shales with marine fossils (López Gamundí, 1983, 1989).

The Paraná basin, one of the largest intracratonic basins of South America, is situated in the central-eastern part of the continent. It is present from the Mato Grosso do Sul to the Rio Grande do Sul state, in Brazil, and also in eastern Paraguay, Argentina and northern Uruguay (Milani and Zalán, 2003). The shape of the depression is elliptical and covers about 1,500,000 km^2 (Souza *et al.*, 2007; Milani *et al.*, 2007), with 2/3 of its surface covered by Mesozoic basaltic lavas of the Serra Geral Formation (Melo, 1988; Zalán *et al.*, 1991; Souza *et al.*, 2007). The stratigraphic interval directly related to this work is the Gondwana I, a Late Carboniferous to Early Triassic Megasequence that includes the glaciogenic Itararé Group (Milani 1997).

The data set comes from the well exposed Paganzo Basin and the densely covered Paraná Basin in order to use the first as a possible analogue to the second one. As both study areas in the Paganzo Basin represent highly (Quebrada de Las Lajas - QLL) to moderately (Quebrada Grande - QG) confined depositional settings, the area chosen in Brazil (Vidal Ramos - VR) was also supposed to represent some sort of depositional confinement. The Argentinean areas include a narrow palaeofjord (QLL) and a palaeovalley (QG), both located in the western sector of San Juan city (Fig. 1), which were rapidly filled subsequently to Pennsylvanian Gondwana glaciation (Dykstra, 2005). They show excellent exposures, minimal structural overprint, and three-dimensional view that allow a detailed observation of the evolution of their thick turbidite fill. On the other hand, the Brazilian counterpart comprises a previously mentioned palaeovalley (Santos, 1996) situated near the south-eastern border of the Paraná Basin in the Vidal Ramos (Santa Catarina State) surroundings (Fig 2). Its geological record is poorly exposed due to a dense plant cover. Therefore, a representation of the turbidite facies geometry, architecture and stratigraphy can only be assessed from the correlation of scarce, usually small outcrops.
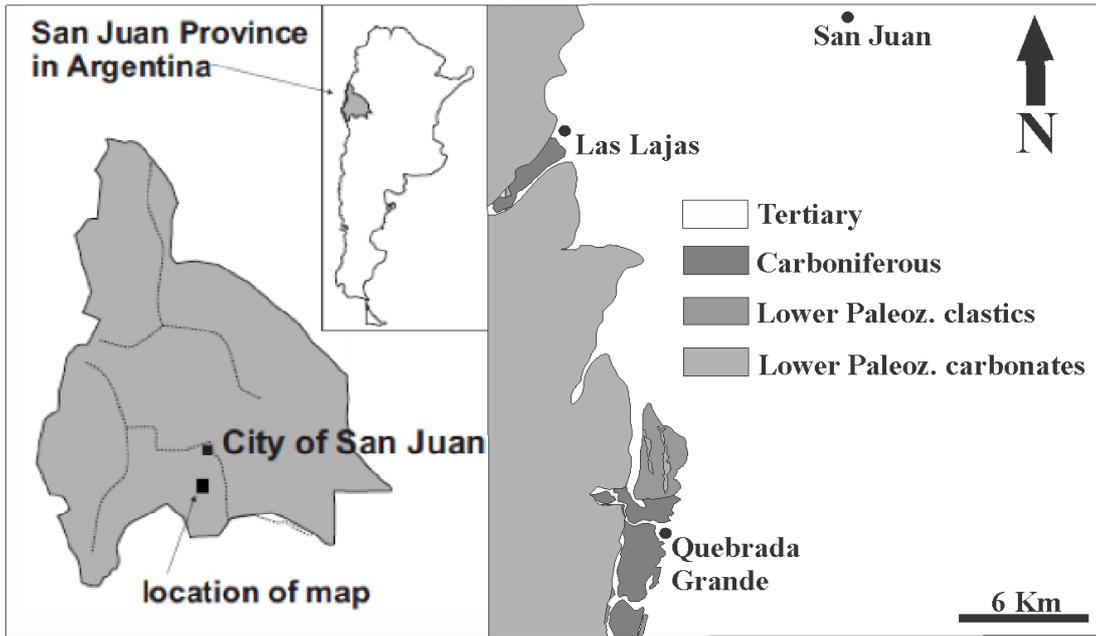
Figure 1: Location of the Quebrada de las Lajas palaeofjord and Quebrada Grande palaeovalley.
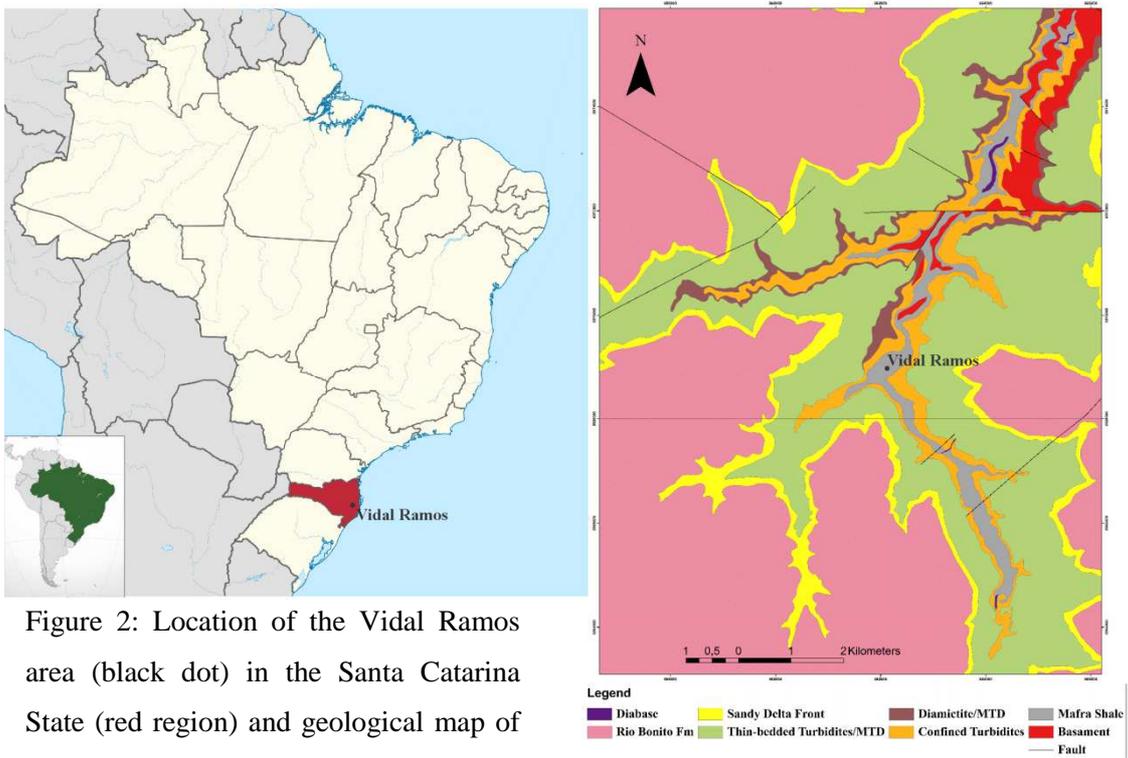
Figure 2: Location of the Vidal Ramos area (black dot) in the Santa Catarina State (red region) and geological map of the studied area.

3. METHODS

Data collection was taken from the Paganzo and Paraná basins and later processed in lab. A geological map of each study area was produced in order to understand and organize individual sedimentary successions.

The lack of detailed topographic maps as well as the arid climate in the Paganzo Basin lead to the use of satellite and aerial images, GPS marker-points and outcrop description to produce the QLL and QG geological maps (1:10,000). Their incorporation into ArcGis project has enabled later input of data acquired from laser-shots used to locate unreachable points and delineate inaccessible stratigraphic surfaces.

In the Paraná Basin (VR), topographic-map, GPS points, laser-shots and outcrop features were used to produce a larger-scale (1:100,000) geological map in an area characterized by dense vegetation and usually largely weathered.

Both in Argentina and Brazil, several sedimentary logs were measured in strategic points during the mapping to describe facies and facies succession and later to define turbidite stratigraphic architecture. To improve depositional elements characterization, the analysis of photo-mosaic and 3D images (supported by laser-shooting) analysis was performed in both QLL and QG.

To demonstrate the interplay between-confinement shape and sediment transport paths within these confined local basins, paleocurrent readings were taken from all mapped units, usually presented as rose diagrams and / or mean vectors, and from the basal surface scoured by ice into Precambrian basement rocks. To better differentiate distinct turbidite intervals and to bring into light some pieces of evidence related to source areas, some samples were taken to petrographic analysis. A few, dark grey to black shales were sampled for TOC analysis.

4. CONCEPTUAL BACKGROUD

The interplay of the dynamic properties of the eroding glacier and the mass strength of the subglacial bedrock control glacial valley morphology. When ice discharge reaches its maximum at the equilibrium line altitude (ELA), the pattern of erosion ought to mimic this trend (Anderson *et al.*, 2006), depending on bedrock strength as there is a strong positive correlation between rock mass strength and Fr number (Fr = D/2w), where D is the valley depth and w is the valleys half width (Sugden and John, 1976).

For the interpretation of the turbidite sand-sheets its was used the concept in which the motion of the fluid behind the turbidite head can be approached with the modified formula of the Chezy equation for flow in open channels, using reduced gravity (Middleton, 1993), and it is slope-dependent. Thus, the buoyance flux (Turner, 1973) into the head increases with increasing slope, with a concomitant effect on mixing. Finite-volume releases (surge-type currents) may be dominated by the properties of the head in contrast to sustained underflows (Hacker *et al.*, 1996). The deposits of turbidity current mostly depend of the behaviour of the particles in terms of spacial and temporal acceleration within the flow. Accepting that when both accelerations are zero, the fluid and particles will be moving in a constant flow, if it has a substantive acceleration the particles will not fall from flow suspension (bypass/erosion). When the fluid starts decelerating, it will eventually reach a suspension threshold forcing the sediment to fall-out (Bouma, 1962).

4.1. Confinement

In many cases, the development of turbidite systems is moderately to strongly modified by pre-existing bounding slopes (Lomas and Joseph, 2004). Van Andel & Komar (1969) used the term *ponding* to refer to a situation in which turbidity currents of

sufficiently large volume are fully contained in an area of subdue bathymetry. Similarly, Pickering & Hiscott (1985) used the expression "contained turbidites" to describe beds deposited from turbidity currents that were confined within a basin too small to allow sustained unidirectional flow. These terms describe situations in which gravity flow sediments fill the floor of an enclosed depression and are unable to surmount the bounding sloped which bound that depositional area. Closely associated with the concept of confined deposition are the notions of flow reflection and deflection and fill-and-spill. According to Pickering and Hiscott (1985) some common features related to turbidity currents confinement include (1) the presence of thick siltstone/mudstone caps overlying or interbedded with sandstone beds; (2) complex multidirectional palaeocurrent directions; (3) dominance of thick beds; (4) complex grading patterns; (5) soft-sediment-deformation; and (6) onlap of sheet-like beds onto confining-basement.

The upper part of confined sedimentary successions tends to present an increase in the sand-to-mud ratio due to flow stripping (Fig. 3), *i.e.* the denser, sand-rich flow travels along the basin floor eroding and depositing amalgamated sandstones, while the finer-grained material stays in suspension, therefore escaping from confinement (Sinclair and Tomasso, 2002).
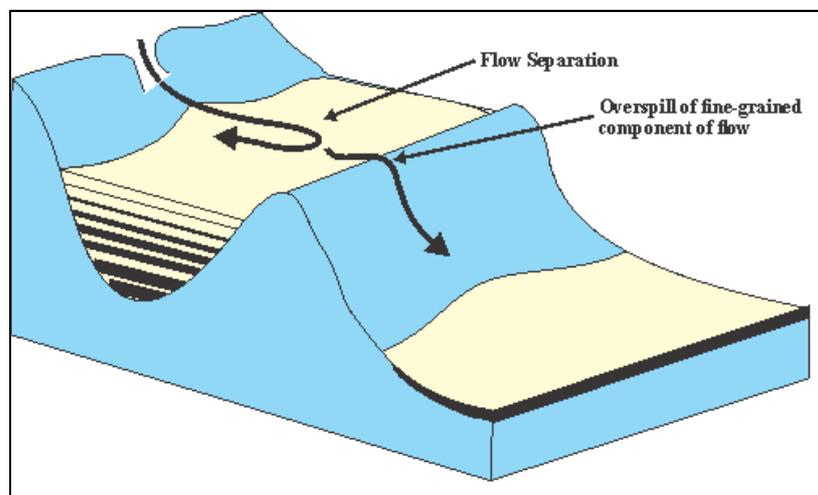
Figure 3: Depositional model for the infill of a confined turbidite basin at the flow stripping stage (modified from Sinclair and Tomasso 2002)

4.2. Onlap

The development of stratigraphic traps, a key feature of many hydrocarbon reservoirs, is favoured at the margin of confined turbidites sheet systems (*sensu* Haughton, 1994a). According to McCaffrey & Kneller (2001) there are two main types of pinch out (types A and B). In the first case (Type A), turbidites thin onto the confining surface – although the final sandstone pinch-out is commonly abrupt – and individual beds tend not to erode into earlier deposits. In type B, turbidite sandstones commonly thicken toward the confining slope, and bed may incise into earlier deposits. These two types may occur in combination, so producing a wide spectrum of pinch out characteristics (Fig. 4).

Figure 4: Classification of sand body terminations after (a) McCaffrey & Kneller (2001); and (b) Smith &Josep (2004). *Apud* Gardiner (2006).

Sandbody pinchout occur in a range of depositional environments and can take a variety of forms. In the case of glacial-valley, individual beds may preserve their thickness, but usually as they approach the onlap surface sandbody thickness tends to decrease. (Gardiner, 2006). In some cases, as sand beds thin at the onlap surface, they continue up the surface for some distance (McCaffrey & Kneller, 2006; Smith & Joseph, 2004), often with a thickness of only a few cm.

5. RESULTS

The results were firstly based on the description of the fjord/valley morphology. Acknowledging the shape of each glacial valley, their fill was described, with emphasis on the turbidite sand sheets. Turbidite sand-sheets were interpreted in terms of the main transport (long- and short-distance) and depositional mechanisms as well as in terms of temporal and spatial acceleration of the turbidity currents and the type of lateral onlap in each studied area. At last, possible relationships between valley morphology and sedimentary fill were then endeavored.

However, before describing their morphology, it is important to highlight why they are treated as glacial-carved features. First of all, the general glacial setting of the Jejenes Formation in the Paganzo Basin and the Mafra and Rio do Sul formations in the Paraná Basin is a well-known aspect for a long time (e.g. Schneider et al. 1974; Eyles et al 1985; Canuto 1993; Eyles and França, 1993; Limarino 1987; Milana and Bercowski 1993; Limarino et al 2002). In terms of the specific areas of study the Vidal Ramos palaeovalley was previously ascribed to glacial carving by Santos (1996) whereas a palaeofjord setting for the succession exposed in the Quebradas Grande and Las Lajas was presented by Kneller et al (2004) and Dykstra et al. (2006, 2007). Besides, the Vidal Ramos palaeovalley presents glacial striations (Fig. 5a) on the basement rocks (Santos 1996) above which the glacio-marine deposits rest (Mafra Formation). The Quebrada de Las Lajas palaeofjord is an entirely exhumed and exposed feature that is elongate, narrow and bounded by almost vertical walls whose initial record includes glacio-lacustrine mudstones with dropstones (Dykstra, 2007). At last, the Quebrada Grande palaeovalley, although not wholly exposed as the Las Lajas palaeofjord, also comprises an elongate and narrow feature, confined by a steep and high wall at least to the west , that includes ice-related features in its sedimentary record (Fig. 5b).

Fig.5. Feature associated with the glacial nature of the studied palaeovalleys, glacial striation at the base of the Vidal Ramos palaeovalley fill (a), iceberg-related scour features in the Quebrada Grande palaeovalley fill (b).

5.1. Glacial Valley Morphology

Three glacial valleys are here presented. They range from a very narrow and deep palaeofjord (QLL) through a moderately narrow and deep palaeovalley (QG) to a much broader and shallower palaeovalley (VR). All of them include deep marine turbidite successions. The QLL palaeovalley presents the higher depth to width ratio (Fr number = 0.55) whereas the VR one shows the lowest Fr value (0.006). Although the QG palaeovalley presents an intermediate ratio, it is much closer to the QLL Fr number (Tab. 1). This range of valley configuration, and their relative differences, are better seen the figure 6, which display a typical cross-section of each example, from the narrower to the wider one, hence enhancing their differences.

Glacial-Valley	Depth	Width	Fr Number
Quebrada de Las Lajas (Argentina)	297	270	0.55
Quebrada Grande (Argentina)	258	741.8	0.17
Vidal Ramos (Brazil)	80	6000	0.007

Table 1: It presents the parameters that define glacial valleys morphology. Note the wide range of depth to width ratios (Fr number).
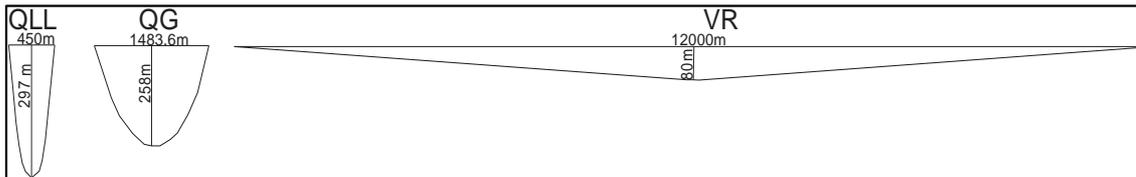
Figure 6: General picture of the glacial palaeovalley morphology.

The QLL and QG palaeovalleys share a similar basement. The former was carved into Ordovician limestone (San Juan Formation) whereas the latter was scoured into debris-flow strata encompassing a large amount of huge carbonate blocks derived from the same San Juan Formation. The expected uniaxial compressive strength (Marinos and Hoek 2000) of intact basement of these two valleys is strong. The QLL basement is likely to have a uniaxial compressive strength between 50 and 100 MPa and a point load index of 2 to 4 MPa. The QG palaeovalley was carved into a slighter weaker mass strength rock, as suggested by its composition and indicated by its slighter wider and shallower geometry relative to the QLL. Although both have been carved by eastward moving glaciers sourced from the southeast in the proto-precordillera Sierras Pampeanas, their positions and directions are different, the QLL oriented to SW-NE and the QG to N-S. The VR glacial palaeovalley is the widest and shallowest one. It was scoured into a metasedimentary basement (mica-schist and scarcer meta-conglomerate and marble). It is possible to estimate a uniaxial compressive strength of 25 to 50 MPa

and a point load index from 1 to 2 MPa for this sort of metamorphic rock, therefore explaining its wider and shallower nature.

5.2. Glacial Valley Fill

5.2.1. Quebrada de Las Lajas palaeofjord

The Quebrada de las Lajas (QLL) comprises a 5km long and up to 600m wide fjord-type valley that records a maximum sedimentary pile of 297 m at its centre. Its sedimentary fill was subdivided into five main stages (Figs. 7, 8 and 9) being the III and IV ones those that encompass the more expressive sandy turbidite intervals (Figs. 8 and 9). These later intervals comprise a pile of about 200m that is better exposed in the centre of the palaeofjord. Successive stages are gradually exposed from both margins to the centre and from SW to NE, the later due to the differential uplift and resulting exhumation caused by Andean tectonics. As a result, older stages are exposed at the head of the palaeofjord and younger stages at the mouth (Fig. 7).
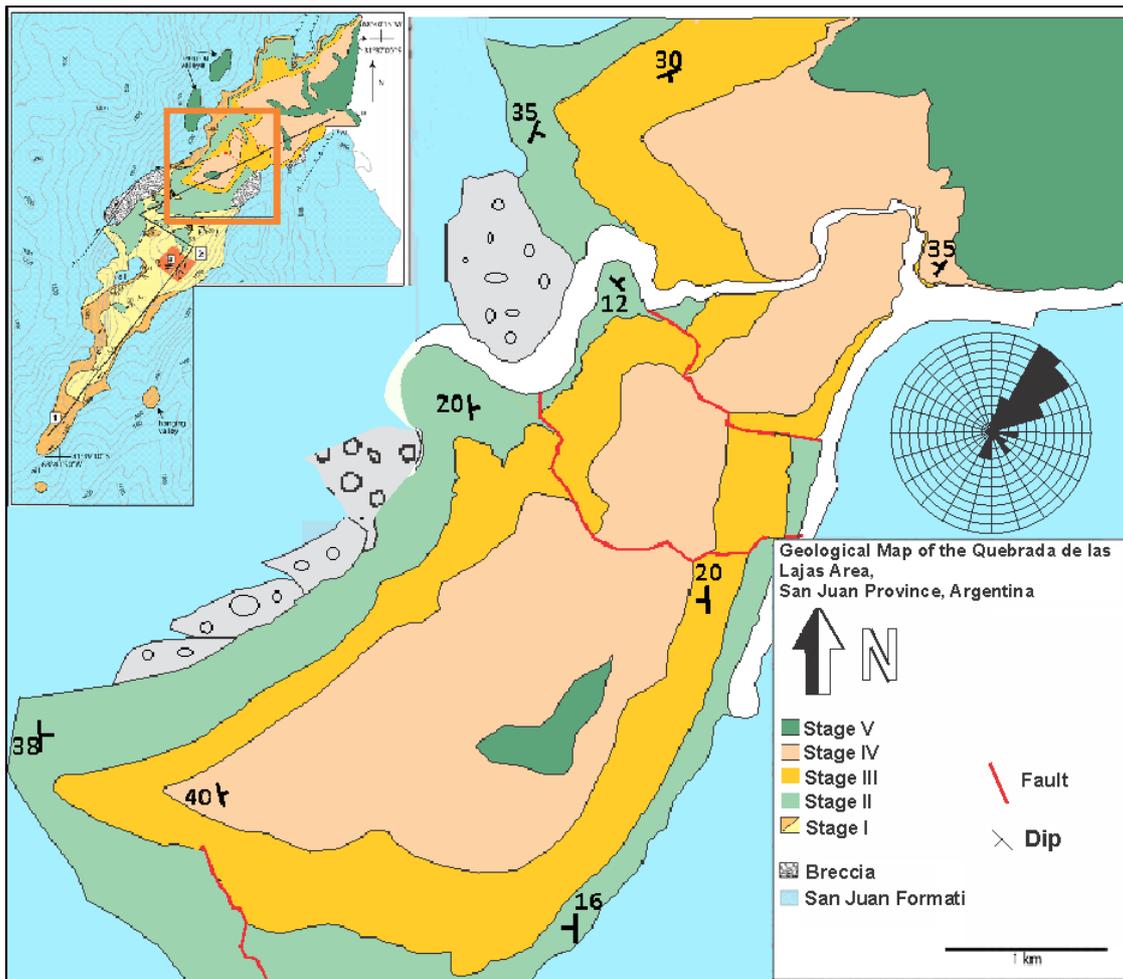
Figure 7: Quebrada de Las Lajas geological map (modified from Dykstra, 2007) and detailed geological map of the central sector of the palaeofjord.

Paleocurrent readings taken from sole-marks (flute / groove casts and tool marks), current ripple cross-lamination and parting lineation show a very consistent north-eastward trend, *i.e.* parallel to the fjord orientation, more specifically from head to mouth. This trend suggests a strong control of the flows due to the confined setting. A few transversal paleocurrent data were recorded and supposed to represent current rebound along confining walls, occasional sediment discharge from hanging-valleys and / or turbidity currents derived from wall failures.

Figure 8: Large scale photomosaic of stages I-V displaying almost the entire fill of the Quebrada de las Lajas palaeofjord.

The QLL palaeovalley fill records a transgressive-regressive cycle (Fig. 8). Initially it includes breccias, conglomerates, interbedded sandstones and conglomerates, and debris flow and slump deposits related to mass-transport (MTD), which are overlain by thin bedded siltstones and greenish mudstones (rhythmites) with dropstones. This facies association represents proglacial deltas and lateral talus deposits overlain by lacustrine facies of Stage I (Dykstra, 2007).

Underlying the turbidite sand-sheets (stages III and IV) there is an interval (Stage II) that comprises organic-rich black shales and thin-bedded, very fine-grained sandstone (massive, parallel- and ripple cross-laminated) and mudstone. Plant remains, such as large trunks, leaves and plant debris, are abundant. These facies are interpreted as basin plain, mostly low-density turbidites and hemi-pelagic deposits. Black shales are supposed to record the maximum flooding in the palaeofjord. Paleocurrent readings indicate north-eastward turbidity currents, i.e. from palaeofjord head to its mouth.
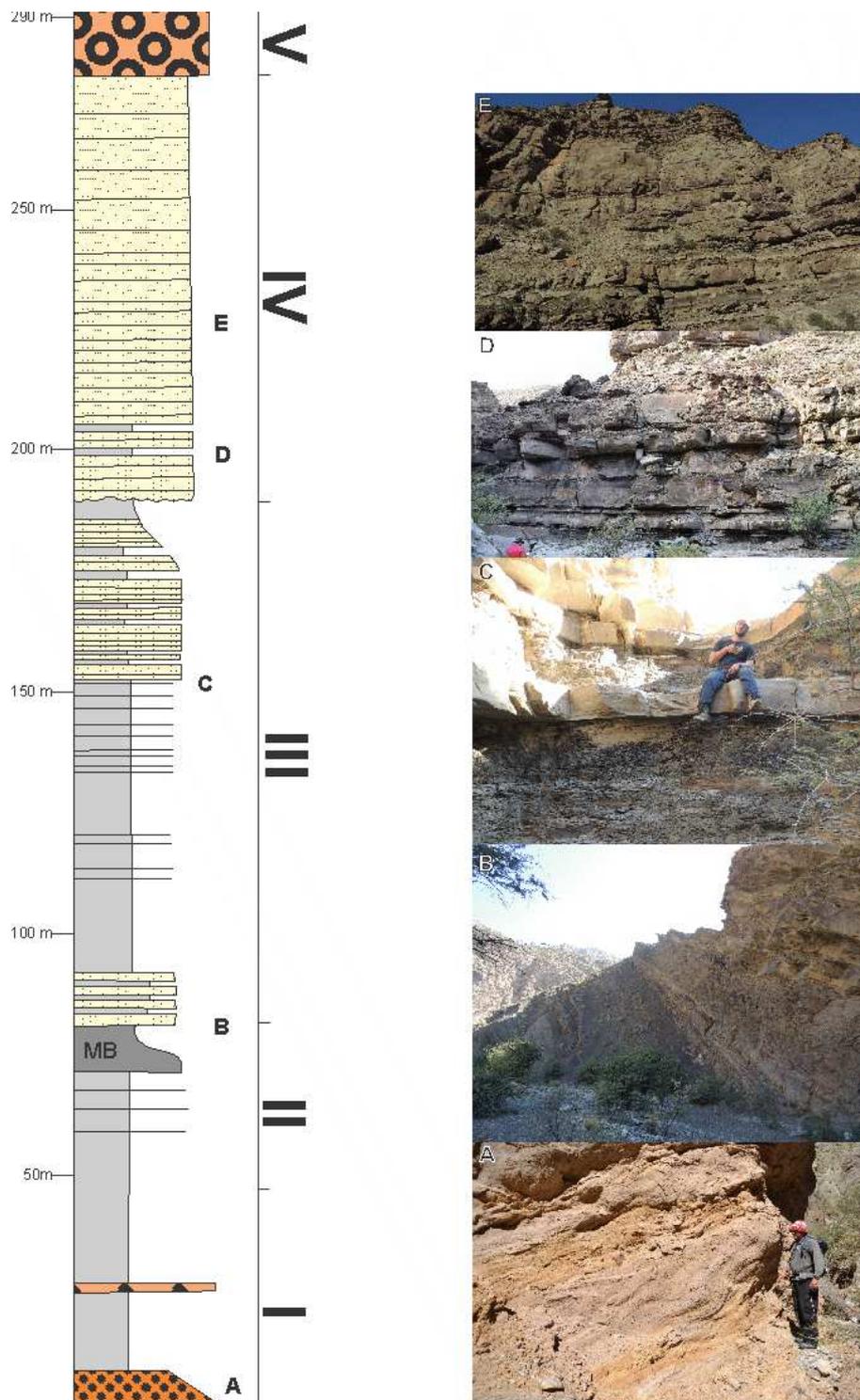
Figure 9: Composite stratigraphic log of the QLL palaeovalley (photos located in the log):
(A) slumped sandstone and conglomerate derived from stage I coarse-grained delta deposits;
(B) contact between stages II and III (notice the black mega bed underlying the turbidite
sandstone sheet); (C) typical Stage III sandstone grading to shale; (D) base of Stage IV
characterized by low mud/shale ratio and thick sandstone beds; (E) mid to top of Stage IV
represented by amalgamated sandstone beds with almost no intervening fine-grained strata.

The upper part of Stage II also includes eventual, 5 cm to 1 m thick graded beds (Fig. 10a). Close to the borders, these beds comprise conglomerate with angular to sub-rounded pebbles (granite, quartz, chert and limestone) that grade upwards to coarse-grained sandstone (Fig. 10b) with cross-stratification (paleocurrent to SE, *i.e.* transverse to the palaeofjord axis). Towards the fjord centre, conglomerates disappear and sandstones are finer-grained, but still display transverse transport trend. These features suggest turbulent, high-energy density flows derived from the hanging valleys.

A 4 m thick, disorganized, clast-supported conglomerate similar to the previous facies, but including very large, protruding limestone blocks (>2.5m), also occurs in this same context (Fig. 11b). Laterally, this bed is replaced by contorted, conglomerate and coarse-grained sandstone also including large limestone blocks (Fig. 11a). These features suggest cohesive debris flows related to slumping and derived from two sources, a local (limestone blocks) and an external and far more distant (granite, chert, quartz grains) source. Contorted conglomerate beds were ascribed to the remobilization of formerly deposited gravels derived from coarse–grained deltas or subaqueous outwash formed at the mouth of the hanging-valleys (Fig. 12). The outsized limestone blocks incorporated within the gravel came from the fjord or hanging-valleys walls.
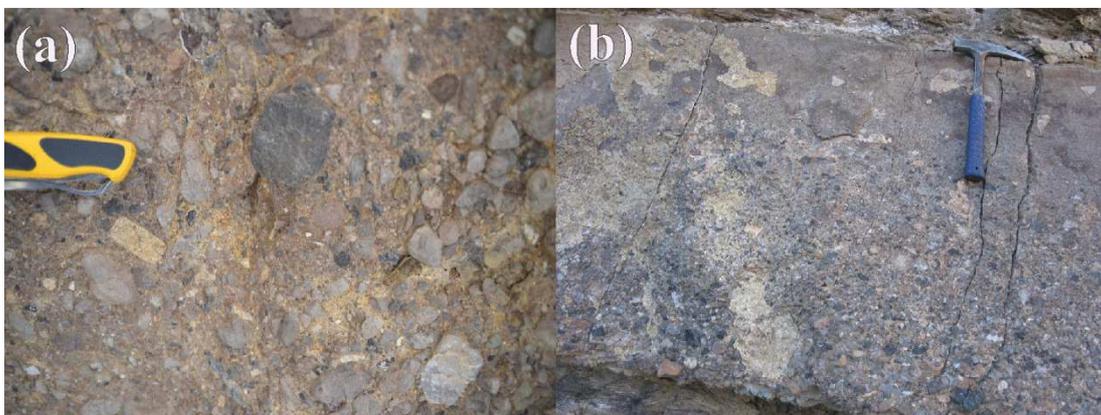
Fig 10: (a) Clast-supported conglomerate (b) grading to coarse-grained sandstone.
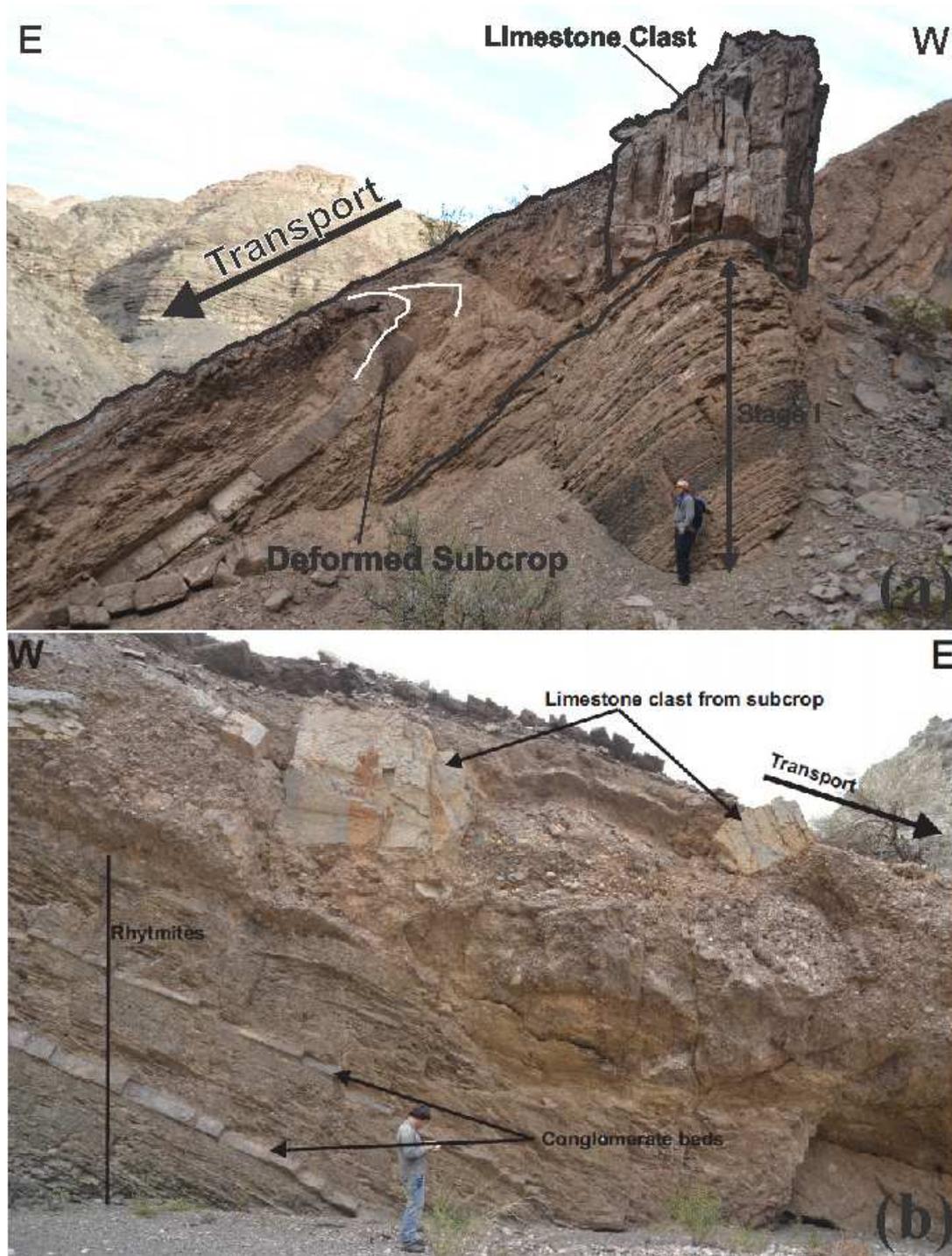
Figure 11: (a) Contorted conglomerate and coarse-grained sandstone strata, enclosing outsized limestone blocks, and (b) clast-supported conglomerate with protruding, limestone outsized clasts.
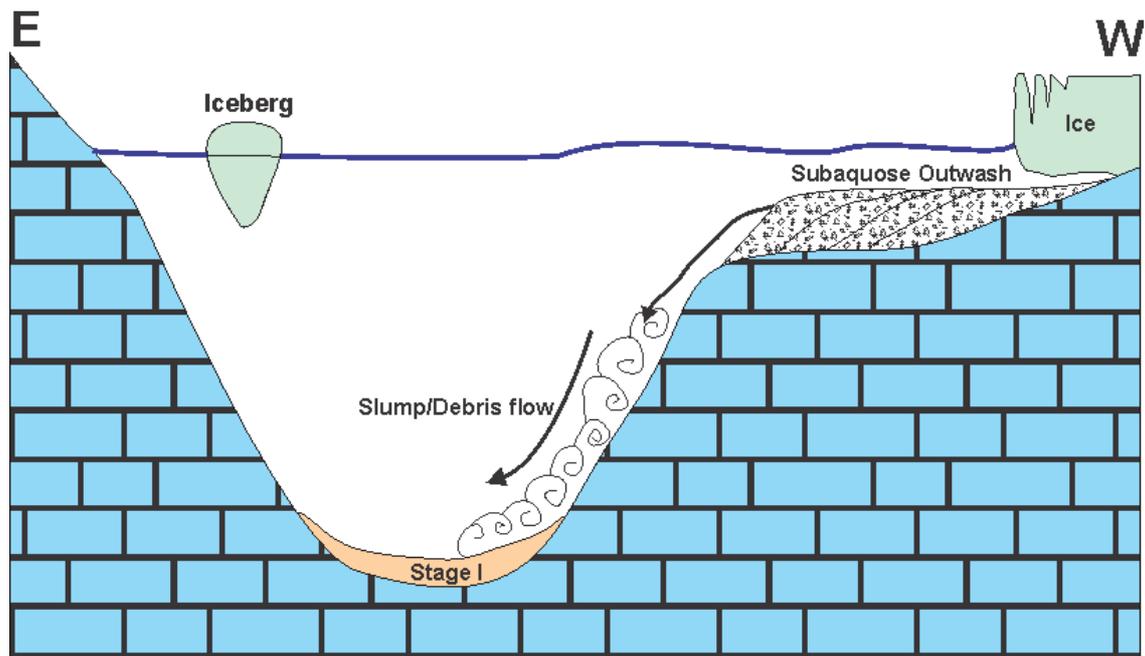
Fig. 12: Conceptual model for the thick, chaotic and/or contorted, conglomerate and coarse-grained sandstone strata with outsized protruding limestone blocks.

Stage II is capped by an up to 10 m thick mega-bed in the palaeofjord centre. This graded, massive bed ranges from sandstone to shale, is particularly rich in plant-debris and has recently been interpreted as the record of a bipartite turbidity flow (Rodrigues, in prep).

The Quebrada de las Lajas turbidite succession represent the progradation of a turbidite system that includes distal (stage II and III) to proximal (stage III and IV) lobe facies. Acknowledging that sediment input was related to a deglacial episode, the gradual progradation of the system (the regressive phase of the T-R cycle) is supposed to represent the ever growing increase of ice melting, liberating more and more sediments to the fjord as a result of progressively warmer conditions.

Besides the strong lateral confinement of the palaeofjord, a longitudinal sill could have trapped the fine-grained, suspended load within the fjord, leaving behind thick mudstone packages interbedded with the sandstone beds in the stages II, III and IV. As the sediment pile grew, the diluted plume of the turbidity currents could achieve the spill point, therefore enabling the fine material in suspension to outflow (Fig. 13). This feature is well recorded at the upper part of stage III, when suddenly mudstone intervals almost disappear. However, a simple progradation of a turbidite lobe could also explain the lack of mud deposits along its proximal portions.
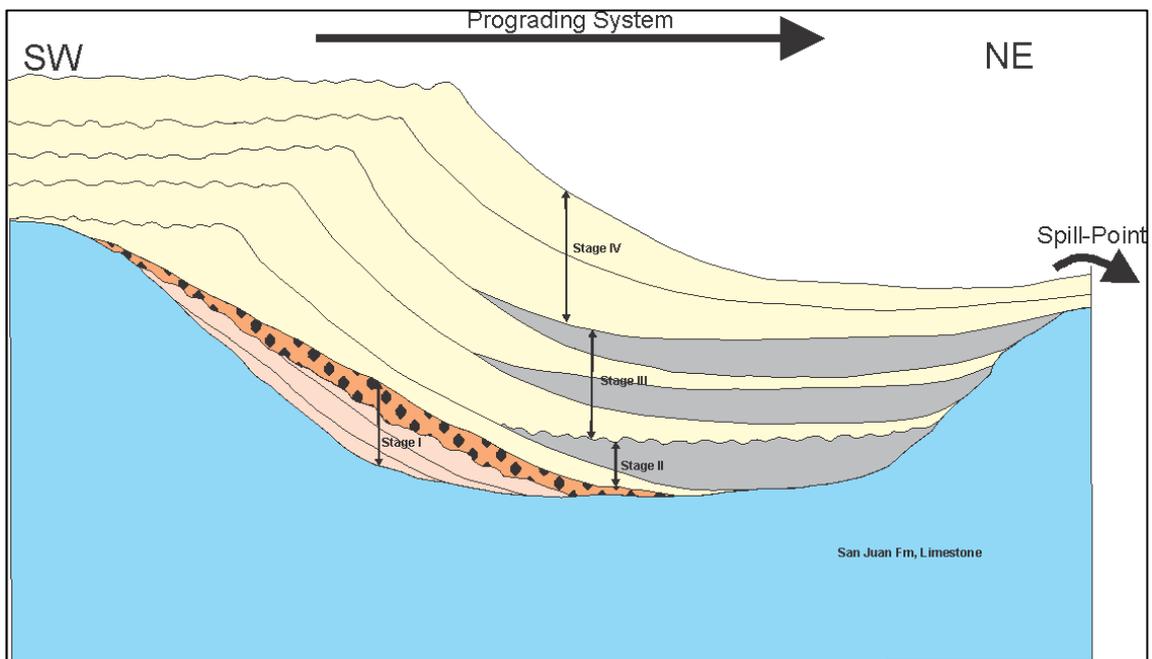
Figure 13: Conceptual model of the QLL along its axis showing the longitudinal confinement and spill point.

Due to the strong lateral confinement and resulting lateral changes in terms of thickness and lithology, the stratigraphic pile tends to replicate the palaeofjord morphology becoming synclinal shaped, although stratigraphy is homogeneous and displays no major changes along the palaeofjord axis (Fig. 14).
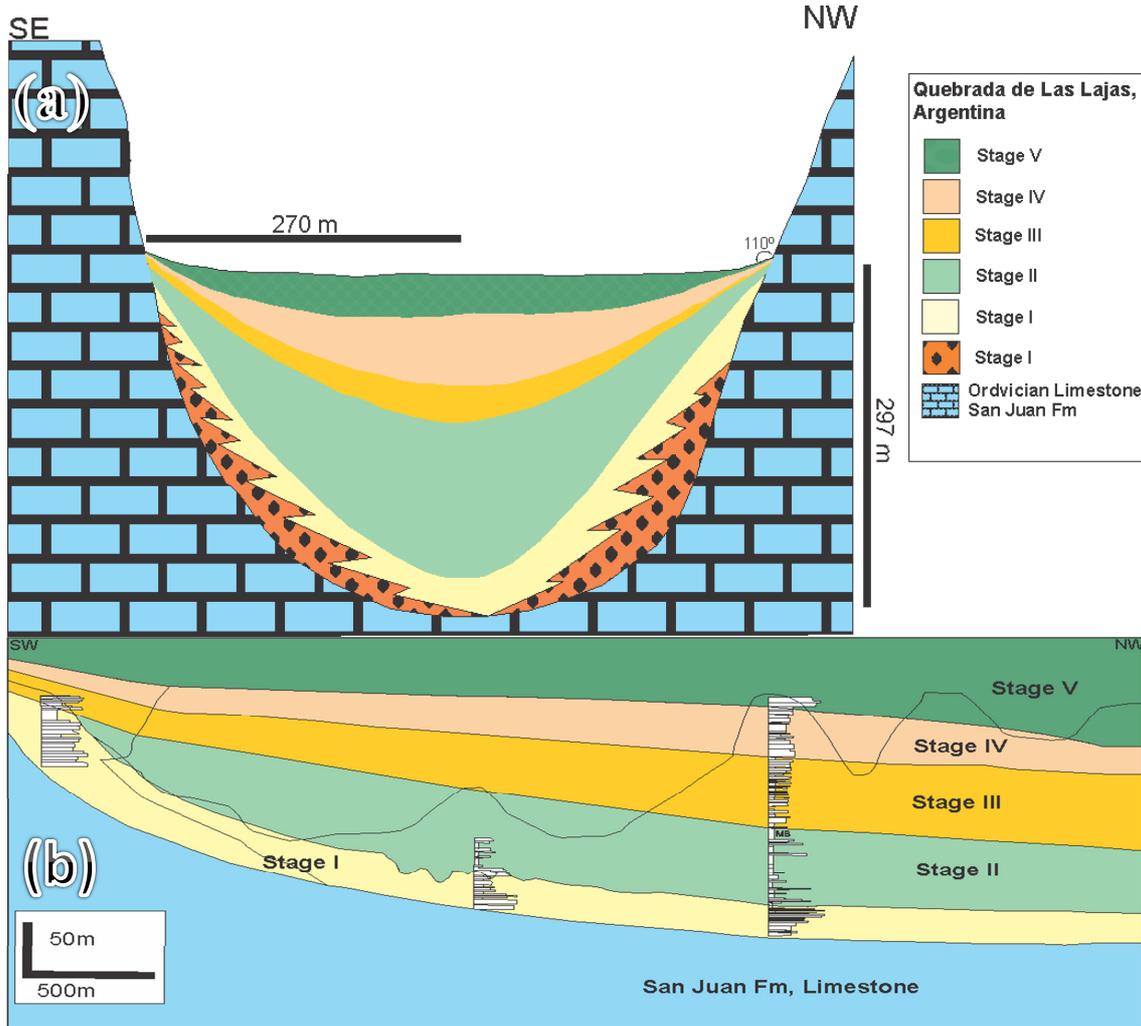
Fig.14: The sedimentary succession that fills the QLL palaeofjord changes laterally due to the pinchout of the sandstone beds (a) whereas is relatively easy to follow along the fjord axis (b).

5.2.2. Quebrada Grande Palaeovalley

Quebrada Grande (QG) is a large palaeovalley located 15 km south of Quebrada de las Lajas. It is about 6 km long and 1.48 km wide, and comprises a maximum sedimentary pile of 250m at its centre. To the west, the sedimentary strata in the Quebrada Grande palaeovalley is confined by a steep wall almost 2 km high composed of Ordovician limestones of the San Juan Formation. To the north and east, however, the confining walls are relatively subdue and related to an olistostrome of the San Juan formation. Valley entrance was located on its southeast portion.

The sedimentary fill was divided into five main stages (Figs. 15 and 16). Stage IV comprises the main focus of this work. The successive stages are gradually exposed from the western to the eastern margin due to the differential uplift and resulting exhumation caused by Andean tectonics. Palaeocurrent readings taken from sole-marks (flute / groove casts and tool marks), current ripples cross-lamination and parting lineation show a very consistent trend toward NW. Only near the western wall palaeocurrents readings of distal turbidite deposits of stage III regularly present an inverse, southeastward trend here interpreted as the result of the turbidity currents rebound (Fig. 15). Stage III is composed of black shale and sporadic, thin bedded, medium- to fine-grained sandstone that increase toward the top of this stage. These sandstones are interpreted as distal, thin-bedded turbidites. Stage III deposits are gradually overlain by a 160m thick, mostly sandy succession (Stage IV turbidites) that are in turn disconformably overlain by the Stage V conglomerates (Fig. 16).

A general cross-section of the Quebrada Grande palaeovalley fill was erected from the correlation of all available sedimentary logs (Fig. 17).

5.2.3. Vidal Ramos Palaeovalley

In the Vidal Ramos area, in the eastern outcrop belt of the Paraná Basin, an up to 400 m thick, Upper Carboniferous to Lower Permian succession (Fig. 18) related to the Upper Mafra and Rio do Sul formations of the Itararé Group is poorly exposed. The older strata of this succession (Mafra Formation) overlie and are confined by Upper Proterozoic, glacial-striated metamorphic rocks (mica-schist and occasional marble lenses) whereas the Rio do Sul Formation onlap and overlies both Mafra Formation and basement rocks.
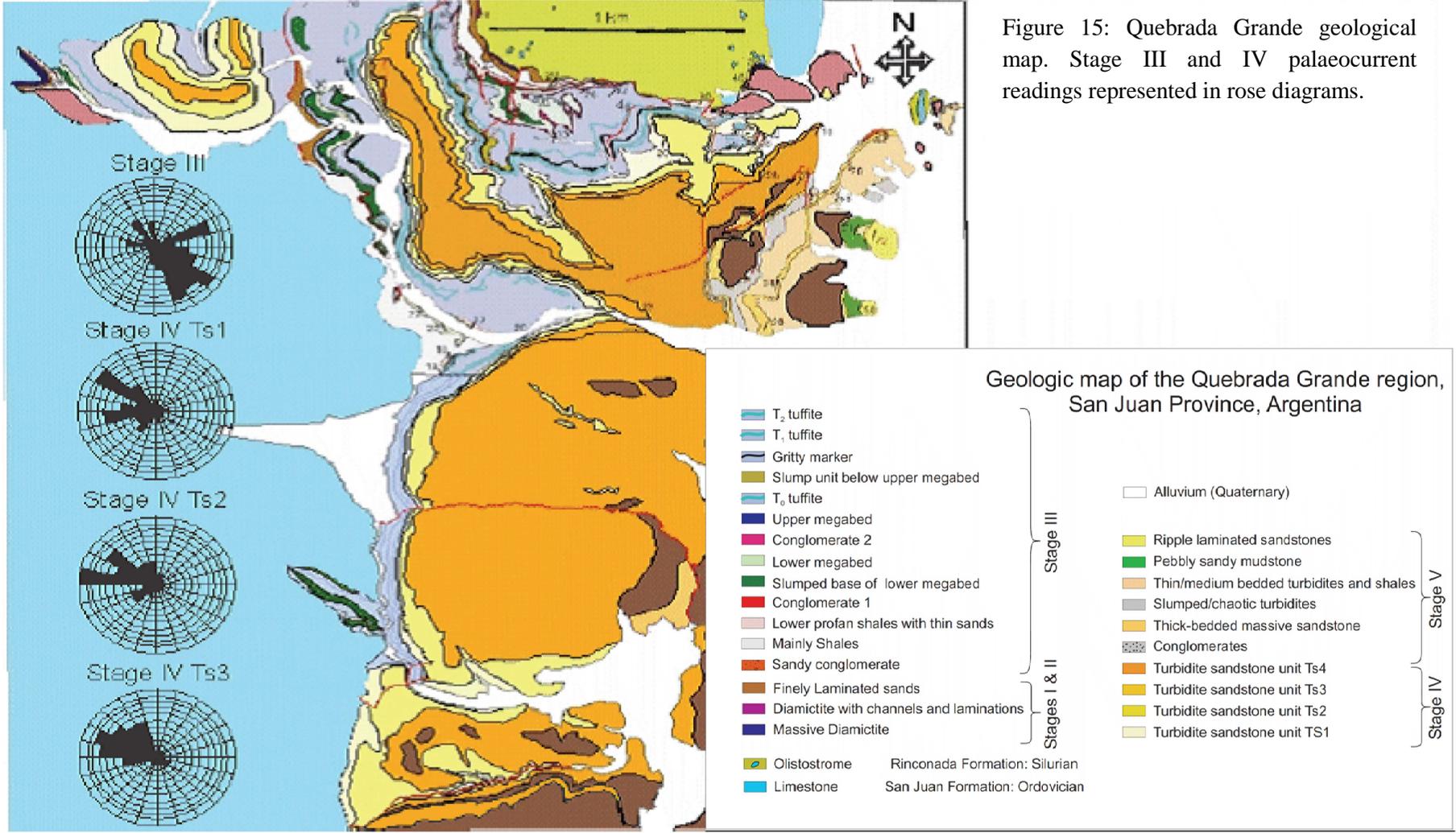
Figure 15: Quebrada Grande geological map. Stage III and IV palaeocurrent readings represented in rose diagrams.
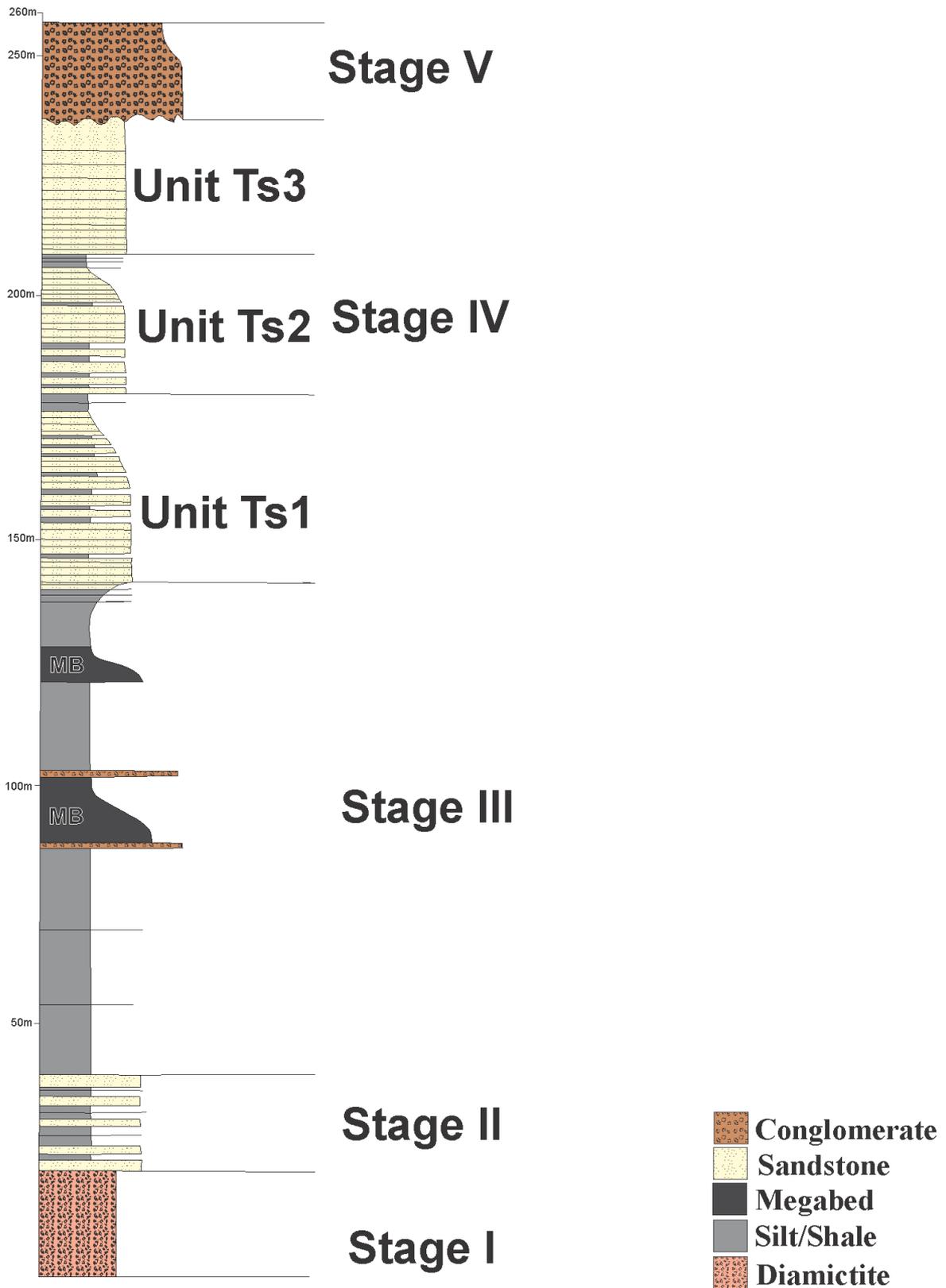
Figure 16: General log illustrative of the Quebrada Grande palaeovalley fill.
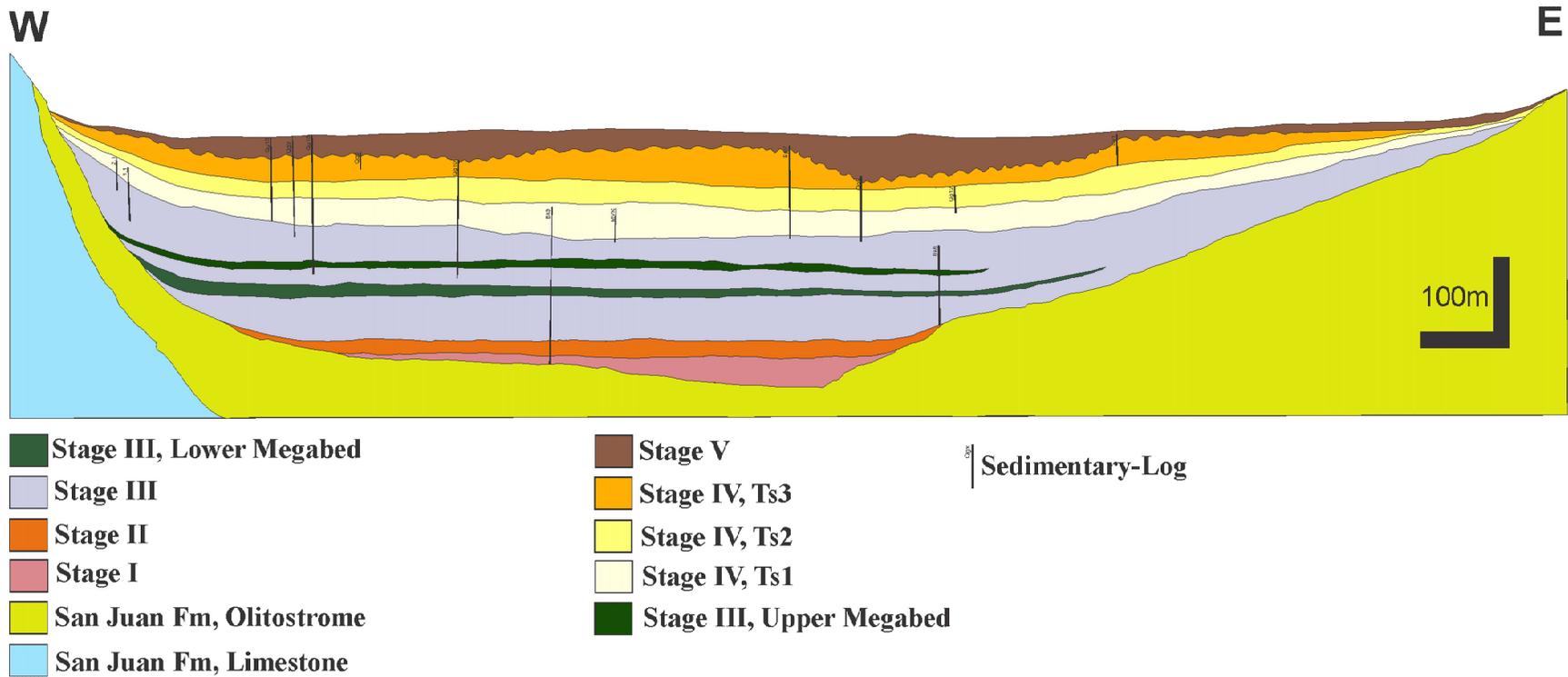
Figure 17: Quebrada Grande cross-section and associated logs (vertical numbered lines).
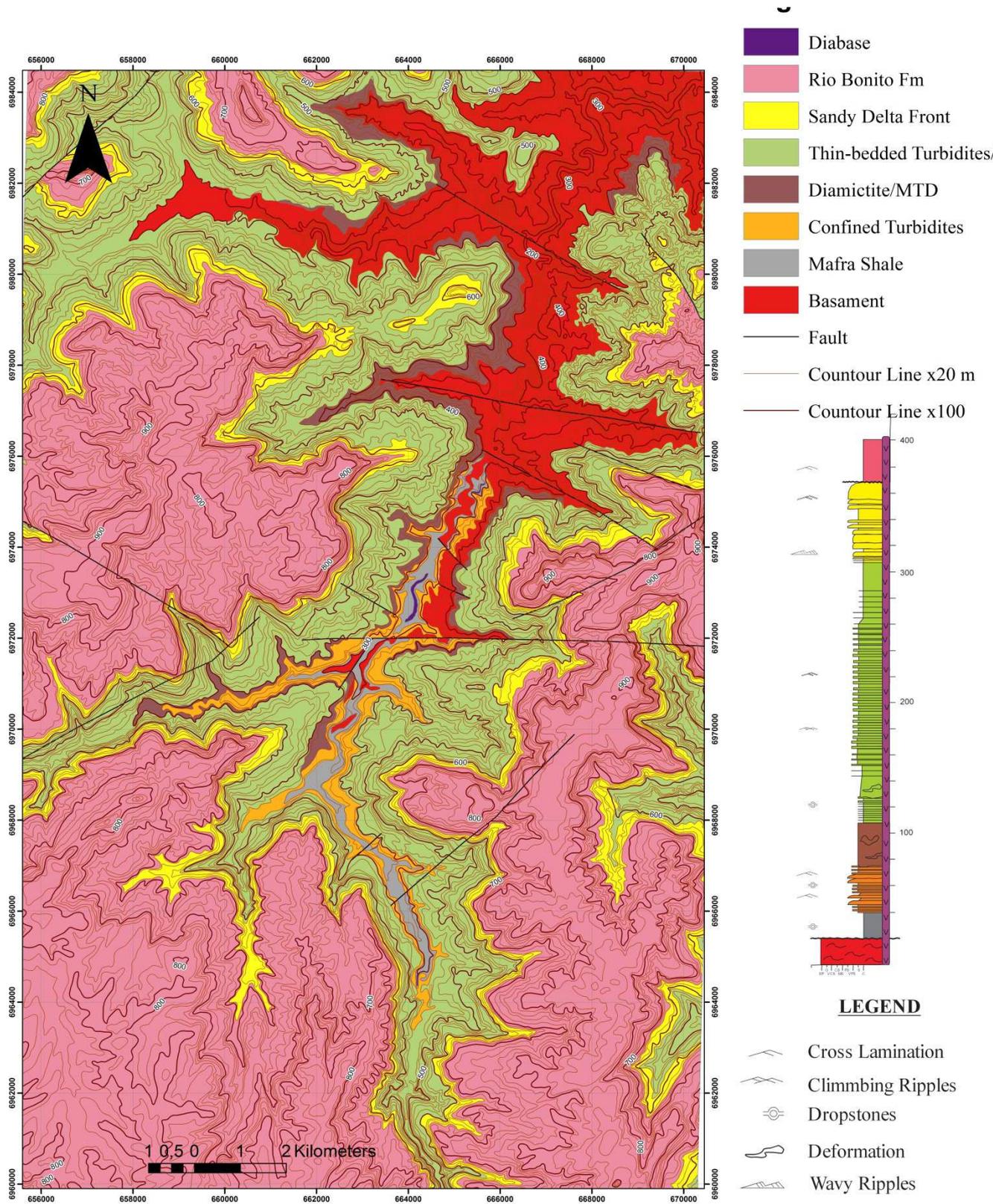
Figure 18: Geological map and vertical section of Vidal Ramos.

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This early confinement is apparent by the north-eastward disappearance of the Formation (Mafra Shale and overlying turbidite sand sheets - Figs. 18 and 19) and by the south-westward and north-eastward thinning of the turbidite strata (Fig. 20). It is also suggested by the NW / SE oriented glacial striation and north-westward trend of paleocurrent readings on turbidite strata (Fig. 18). The Vidal Ramos palaeovalley is 12km wide, at least 13km long and its infill encompasses an 80 m thick sedimentary pile (Figs. 19 and 20).

The VR palaeovalley fill that includes the studied turbidite interval was subdivided into six main stages: the Mafra shale (MS) and the successive R1, T1, R2, T2 and R3 turbidite intervals (Fig. 19). They include basin plain, fine-grained marine deposits (MS), thin-bedded (R1, R2 and R3) and sand sheets (T1 and T2) turbidites, all of them ascribed to a deep (below storm wave base level) marine setting. Stratigraphy is locally disturbed by dolerite dikes and sills of the Serra Geral Formation. Palaeocurrent readings taken from current ripple cross-lamination and few flute casts show a very consistent trend toward NW, parallel to the preceding glacial movement that carved the palaeovalley as pointed by glacial striations scoured into local basement rocks (Fig. 18)

The lowermost unit comprises hemi-pelagic, shaly mudstone known as the Mafra shale (Fig. 21). It is composed of very thin, normal graded lamina that are discernible due to the colour change from dark grey (siltstone) to black (shale). In the centre of the palaeovalley the Mafra shale includes dropstones in its upper portion (Fig. 21), which become rarer next to the confining walls. Total organic carbon content is low (1.2 to 1.4%) and preliminary palinofácies studies show a dominance of wood and amorphous organic matter (probable type 4 kerogen). The above mentioned features suggest a deep marine proglacial setting (black shales with dropstones) subject to the inflow of distal, low concentration turbidite plumes (graded, siltstone to claystone couplets) responsible

for bringing highly crushed, oxidized plant debris into a deep marine setting. The Mafra shale is abruptly overlain by thin-bedded turbidites of the R1 interval, locally displaying angular relationships (Figs. 21 and 22b). This abrupt, sometimes even angular boundary indicates an important base level drop and the formation of a correlative conformity above which the turbidite systems have been deposited as a lowstand system tract.
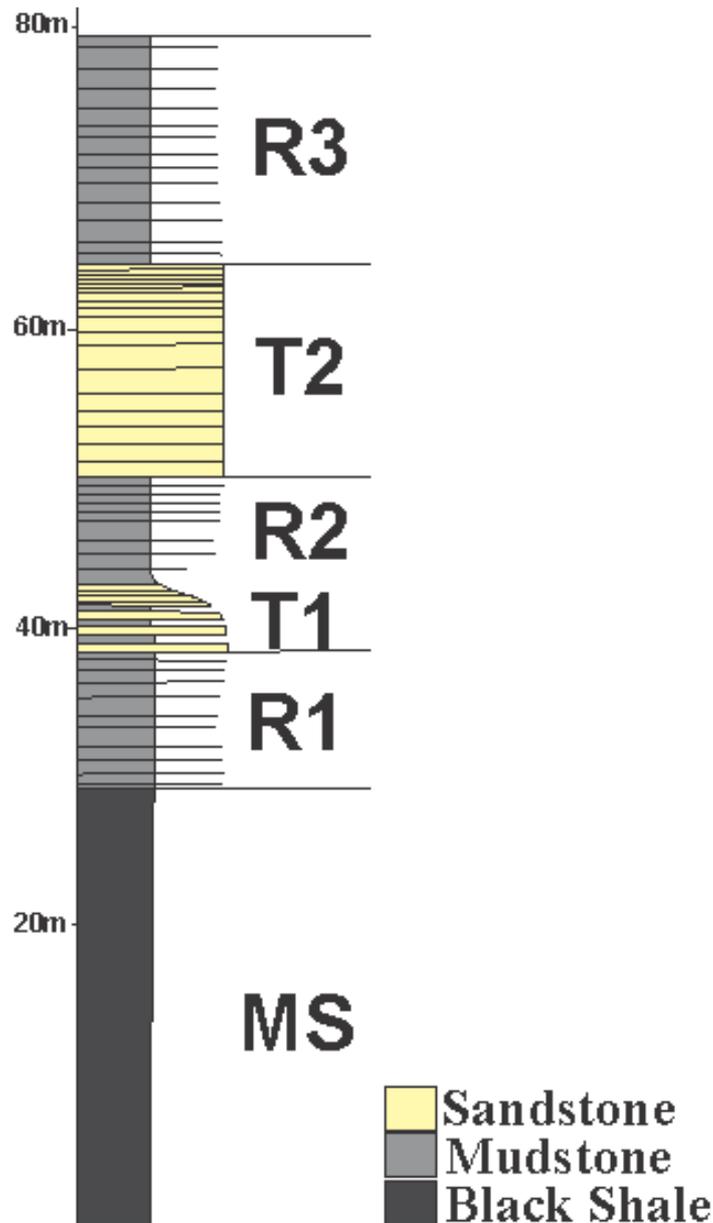
Figure 19: Sedimentary log of the lowermost interval (Mafra Formation) that fills the Vidal Ramos palaeovalley.

The R1 Stage comprises thin-bedded intercalation of sandstone and mudstone. The 1 to 5cm thick beds are composed of fine- to very fine-grained sandstones and siltstones with disperse dropstones. Sandstone display abundant ripple cross-laminations. These rhythmites have been ascribed to seasonally-induced, low density turbidity currents related to waning depletive surges. This stage is supposed to represent a colder interval relative to T1 and T2 stages. During this colder interval, nearby glacial melting rate was low, and so sediment supply. A gradual climate amelioration, increasing melting rates and resulting sediment input, produced the coarsening-upward trend that characterizes its transition to T1 Stage (Fig. 22a).

Transitionally overlying the first sand-sheet (T1 Stage) occur another interval of thin-bedded, very fine-grained sandstone interbedded with mudstone (R2 Stage). Sandstone beds present ripple cross-lamination and grade to mudstone. Mudstones intervals (2 to 5 cm thick) are laminated. Sandstone to mudstone couplets were ascribed to low-density turbidity currents (waning depletive surges) whereas laminated mudstones were related to fine-grained hypopicnal plumes.

Dropstones are present throughout R2 stage (Fig. 23) strengthening the interpretation that from stage T1 to R2 climate got gradually cooler, therefore ice melt rate slowed down, and possibly has even ceased during the advance of the glaciers toward the palaeovalley. This advance has later (deglacial phase) released icebergs to float over the valley and deposit dropstones. R2 interval finishes in a sharp contact with T2 stage, a second turbidite sand-sheet.
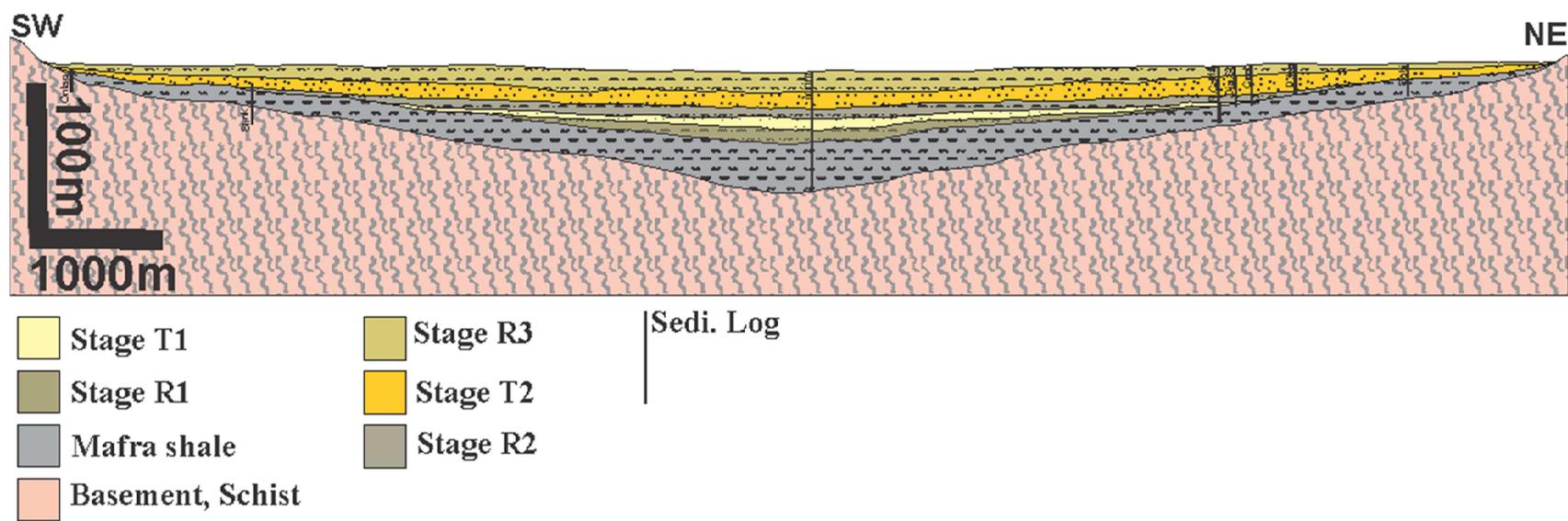
Figure 20: Vidal Ramos geological cross-section derived from the correlation of several sedimentary logs acquired in the southern sector of the area.
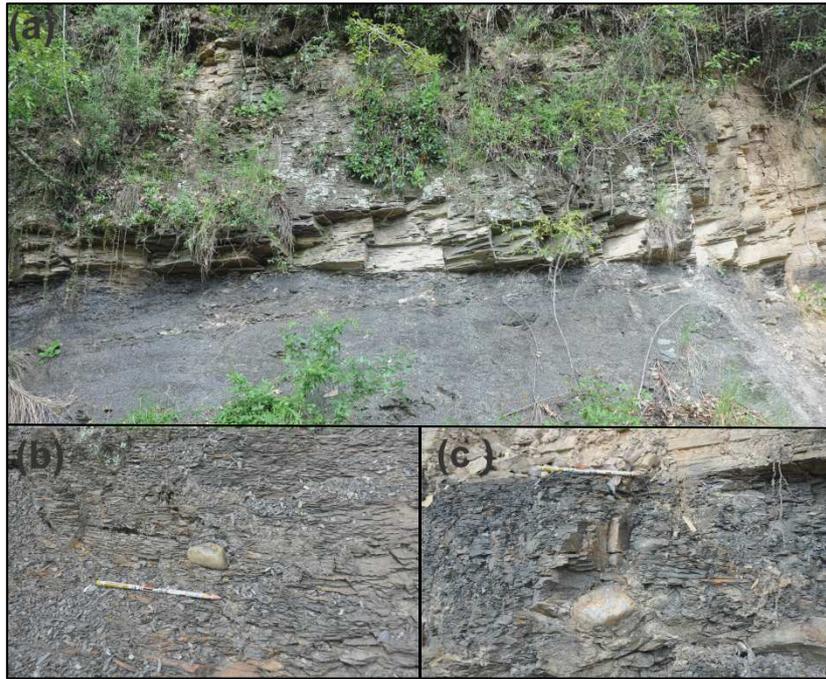
Figure 21: Mafra Shale with dropstones. Notice its sharp contact with the overlying R1 Stage.

Figure 22: (A) Transition from R1 to T1 stages and (B) sharp contact between Mafra Shale and R1 Stage.

Figure 23: Dropstone in the middle portion of R2 stage.

The last turbidite stage (R3 stage) presents a transitional contact with the previous T2 Stage. It comprises another thin-bedded succession composed of very fine-grained sandstone and mudstone. Sandstones present current ripple cross-lamination and dropstones are present along the entire R3 stage strata. These deposits were ascribed to low-density turbidite currents (waning depletive surges) and the usual presence of dropstone as an indicative of another cooler interval, what is compatible with a smaller sediment supply into a proglacial marine setting. R3 strata were suddenly covered by a mass-transport-deposit (MTD) that extrapolates the previously confined depositional setting (Fig. 18) and eroded the top of this stage.

In general, is possible to deduce that due to its shallow and wide nature, the Vidal Ramos palaeovalley was possibly entirely submerged during the inter-glacial marine flooding represented by the Mafra Shale when a large amount of fine-grained material covered the valley floor. Turbidite deposition within the valley took place after an abrupt sea level drop (correlative conformity) related to a new glacial episode. During subsequent deglaciation, the alternation of cooler and warmer intervals produced the intercalation of thin-bedded, dropstone-rich turbidites and sand-rich turbidite

systems, respectively. During the cooler periods, glacial melting rate was low, and the glacier advanced along the valley, liberating icebergs to float over the valley and deposit dropstones. On the other hand, during the warmer intervals thicker sandstone beds related to turbidite currents resulted from higher ice melting rates, and so higher sediment input into the valley.

5.3. Turbidite Sand Sheets: Facies and Facies Association.

The turbidite systems were described and interpreted taking into account the nature of their facies, facies associations and successions, and depositional setting. As a general rule, field data indicate that most of the turbidite beds resulted from short-lived depositional events. They display part or even all of the Bouma's divisions, according to its position from source (proximal/distal) and closeness to the confining walls. Besides, individual turbidites are supposed to represent the product of surge-like, bipartite turbidity currents (high- and low-density) related to high frequency (bedsets and sand sheets) depositional events. Below, turbidite sand sheets of each area are described and interpreted.

5.3.1. Quebrada de Las Lajas Palaeofjord

The turbidite sand sheets in the Quebrada de las Lajas were subdivided into two stages (III and IV). They represent the upper part of the regressive portion of the T-R succession preserved in the palaeovalley (Fig. 9).

The Stage III turbidites displays a thinning- to thickening and associated finning- to coarsening-upward trend (Fig. 9). Its lower portion is composed of medium- to fine-grained sandstones grading to mudstones. Only a few of the sandstone intervals display inverse grading. Sandstone beds exhibits flute and groove casts (Fig. 24a) and common soft-sediment deformation, such as convolute bedding, flame and loading structures

(Fig. 24b). Upwards, sandstones become finer (fine- to very-fine grained) and thinner, display a few ripple cross-lamination at their top and are interbedded with thicker mudstone caps. Towards to the top, sandstone become thicker and coarser (medium- to fine-grained) again and the mud content decreases fast. Sandstone beds are massive at their base, ripple-cross laminated at their top and grade to mudstone. All sandstone / mudstone couplets were related to waning depletive surges (normal grading) ascribed to bi-partite turbidity currents (massive and laminated sandstones). Thicker mudstone packages were ascribed to fine-grained plumes trapped within the fjord due to some sort of longitudinal confinement.
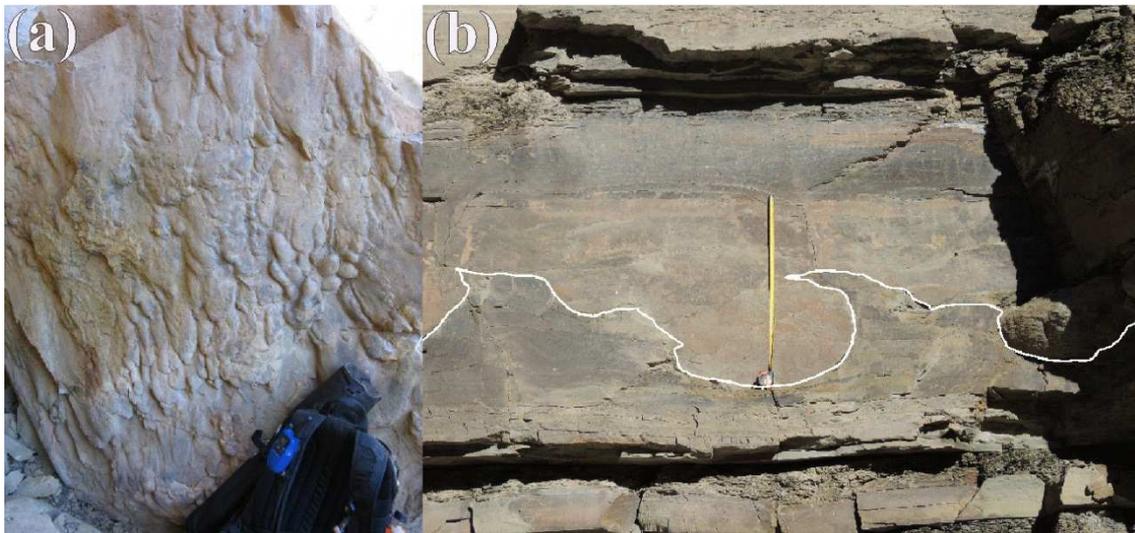
Fig. 24: (a) closely-spaced and overlapping, elongated flute casts and (b) bottom irregularities (loading and injection) at the base of Stage III turbidites.

Stage IV (Fig. 9) turbidite sand sheets are composed of amalgamated, coarse- to medium-grained sandstones beds. Sandstones are massive, include rare mud clasts and eventually display a normal graded top to fine-grained sandstone. These beds were related to waning depletive flows.

A graph derived from the logs shows the upward trend of the sand to mud ratio (Fig. 25). In spite of high frequency changes, this graph displays a general net gross increase

from 0.2 to 0.4 from the base to approximately 100 metres above it whet this ratio becomes stable until the contact between stage III and IV. Just after this boundary, the mud content drops drastically (net gross from 0.8 to 1) This abrupt change can result either from the progradation of the turbidite system or be the result of the overspill and resulting fine-grained plume escape from a confined setting created by a sill (frontal moraine?) at the fjord mouth (Fig. 13).
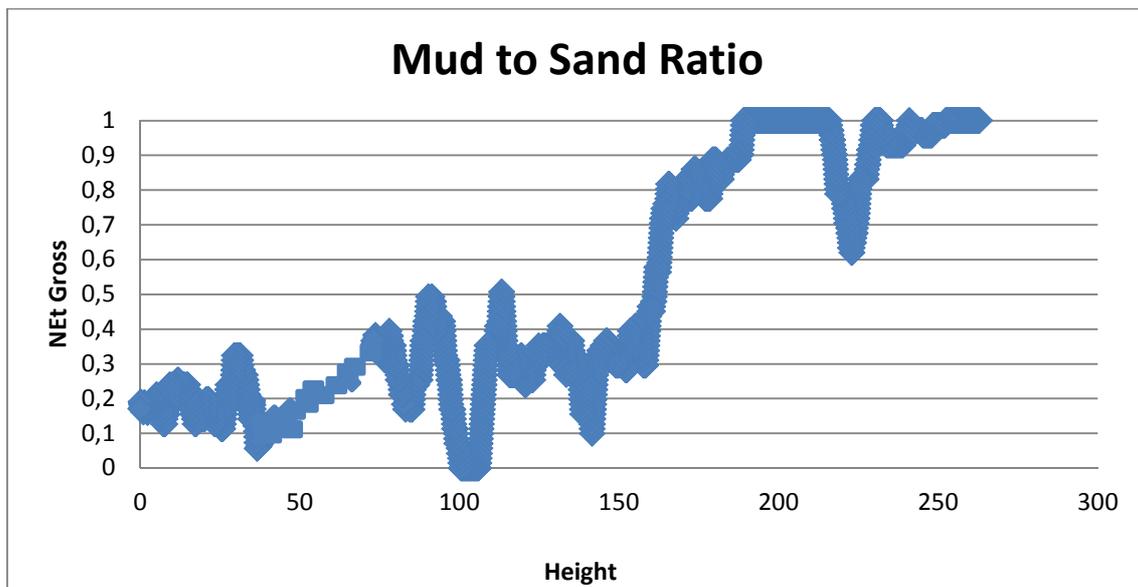
Figure 25: Mud to sand ratio at Quebrada de las Lajas. Notice the sudden drop of mud just above at 150 m above the base

5.3.2. Quebrada Grande Palaeovalley

At Quebrada Grande, the turbidite sand sheet of Stage IV comprises the main focus of this work. This stage was subdivided into three thinner intervals (turbidite systems 1, 2 and 3, in short Ts1, Ts2 and Ts3) bounded by relatively thick mudstone intervals (Fig. 16).

The **Ts1** is composed of 0.2 to 1.2m thick, very fine- to medium-grained sandstone that grades to shale. A few sandstone beds display inverse grading. At its lower portion, the sandstone beds are massive and include a large amount of mud clasts. Sandstone is

made of grains of quartz, muscovite and biotite and presents almost no permeability (Fig. 26). Its usual normal grading indicates waning depletive flows in which fluid turbulence has played a major role as a long-distance transport mechanism. A few, fine- to medium-grained, inversely-graded sandstone suggest eventual waxing depletive flows. On the other hand, the massive nature of the sandstone beds points to late stage flow bipartition and ensuing frictional freezing of a basal, high-density laminar flow (short-distance transport mechanism).
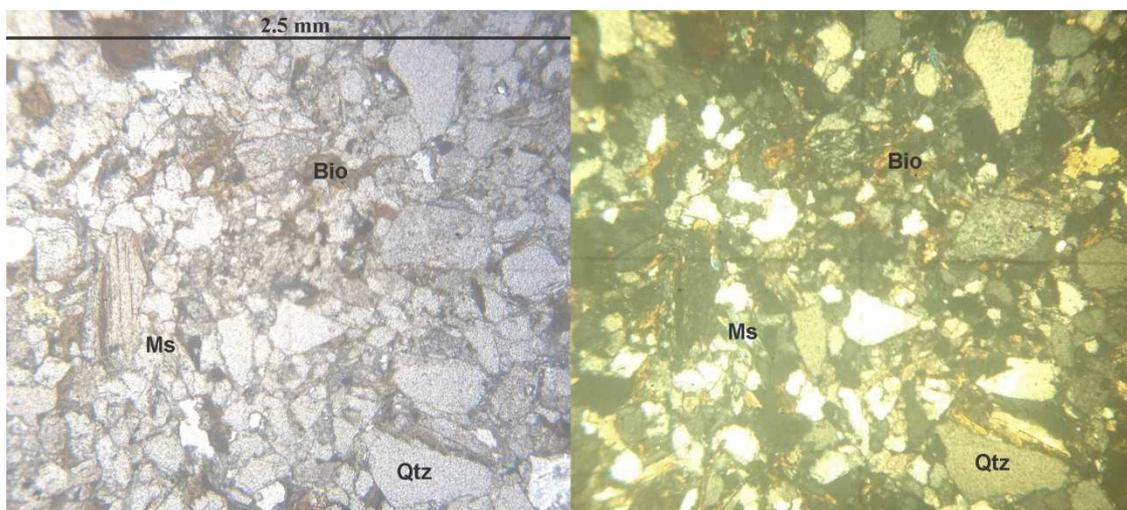
Figure 26: Sandstone of the lower portion of Ts1 composed of biotite, muscovite and quartz. No to very low permeability is suggested by the lack of the impregnating blue resin.

To the top of Ts1, sandstone beds become slightly thinner and grade rapidly into thick mudstone caps. Common features comprise abundant flutes and groove casts, plant material, including impressions of *lepidodendron*, a common primitive, vascular, arborescent plant related to *lycopsids* (Fig. 27c), and frequent ripple cross-lamination at the top of each bed. Similar to the base of this interval, sandstone beds are also composed of quartz, biotite and muscovite and display low porosity (Fig. 28). These couplets were also associated with waning depletive flows, but more turbulent (low-density turbidity currents) than before as attested by the larger amount of ripple cross-

lamination and flute casts. Its turbulent nature was responsible for a better grain size segregation and resulting larger amount of mud caps.

Figure 27: (a) and (b) Well preserved groove cast at the base of Ts2 sandstone; and (c) impression of *lepidodendron* within Ts1 unit sandstone.
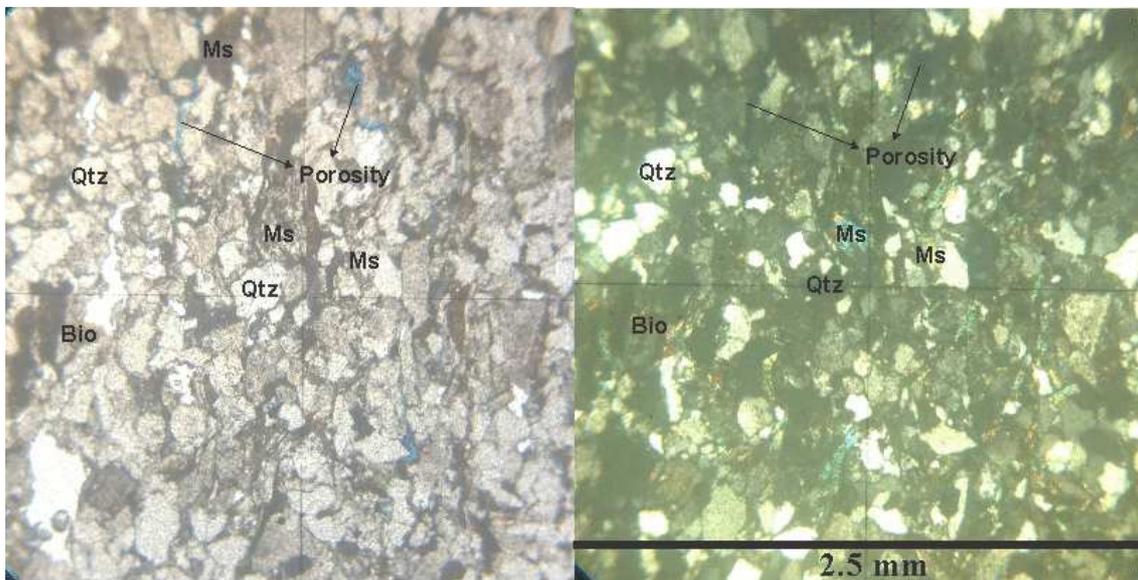
Figure 28: Thin-section of the upper portion of Ts1 unit. A higher permeability and porosity relative to the lower part of this interval is apparent and at least partially a result of the dissolution of calcite veins.

The top of Ts1 is delineated by a 6m thick mudstone interval that includes thin bedded, fine-grained sandstones. This facies association could be ascribed either to a pause in the sediment inflow (allogenic, probably due to climate changes) or to some sort of lobe avulsion (autogenic origin).

The Ts2 sharply overlies the Ts1. Its basal portion is composed of 0.3 to 1m thick sandstone beds that grade upward to thin caps of mudstone. Sandstones present groove casts (Fig. 27a, b) and display a massive base and a ripple cross-laminated top. Sandstone grades upward to thin, about 0.1m thick mudstone. These pieces of evidence indicate long-distance transport by turbulent currents (normal grading and sole marks) and downslope gravity flow transformation into bipartite currents during the depositional phase, as attested by the basal, massive (high density turbidity current) and the upper, ripple cross laminated sandstone and its mud cap (low density turbidity current). Together, these features point to waning, depletive flows.

The middle portion of the Ts2 is also composed of normal (seldom inverse) graded, medium- to very-fine grained sandstones, massive at their base and ripple cross-laminated toward their top, which grade upwards to mudstone. Sandstone beds are rich in plant material (Fig. 29a) and wrinkle structures interpreted as MISS (microbial induced sedimentary structures), which may be related to bacteria, fungi, protozoans, archaea and/or algae (Fig. 29b). Near the palaeovalley walls some palaeocurrent reading indicate reflective currents and small scale folds (slumped beds) become common.

The large amount of plant material within the sandstones could represent either a catastrophic event or a seasonal period in which plants were largely washed into the valley. After these high energy episodes, a low energy interval could allow the wood

pieces to float until a saturation point, forcing it to sink into the bottom of the valley floor. Also during these quiet periods, burrowing and MISS formation took place indicating that the environmental energy was low, but enough to transport nutrients.
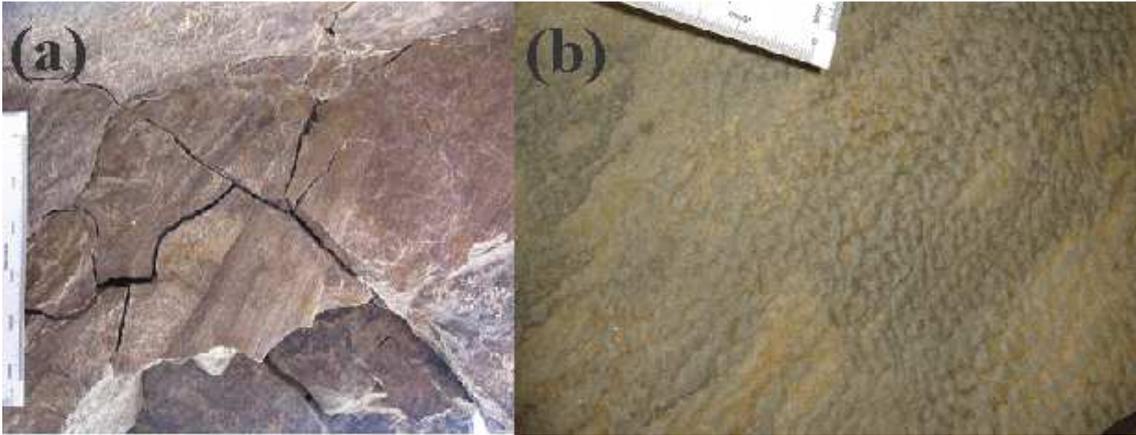
Figure 29: Fossils found at QG Ts2 sandstones, (a) large and abundant trunk impressions at the base of Ts2 sandstones, (b) Microbial induced sedimentary structures (MISS) at the base of Ts2 sandstone bed.

Once again, sedimentary features indicate long-distance transport by turbulent currents (normal grading) that become bipartite downslope (depositional phase) as indicated by the basal, massive sandstone (high density turbidity current) and the upper, rippled interval and its mud cap (low density plume). These pieces of evidence indicate waning, depletive flows related to surge-type turbidity currents. Long, calm intervals are supposed to have been common and enabled the deposition of large wood pieces on the valley floor after their saturation. This plant material was not scattered by high energy bottom currents, but environmental energy was enough to transport nutrients and allow MISS to evolve on the bottom surface, creating primary structures due the interaction of microbes with the sediment. Scarcer inversely graded sandstone beds suggest eventual waxing depletive flows possibly ascribed to hyperpycnal flows.

Along the upper part of the Ts2, sandstones become thinner (0.2 to 0.6m), medium- to very fine-grained and grade upwards into thick mudstone caps (0.1 to 0.2m). Sandstones

are plane-bedded and ripple cross-laminated in their lower and upper portions, respectively. Similar to the top of Ts1, the uppermost interval of Ts2 is topped by 4m of mudstone that includes thin beds of cross-laminated, fine-grained sandstone. These graded beds (laminated sandstone to mud) were interpreted as turbulent, low density turbidity currents ascribed to waning depletive flow.

In general, Ts2 has less mud and gradually thicker sandstone beds than Ts1, suggesting a progradation of the turbidite complex, However, Ts2 turbidite strata thin upwards, indicating a later, progressive decrease of the sediment input and flow energy.

The Ts3 encompasses 0.3 to 1.5m thick, medium- to fine-grained sandstone beds that grade into thin caps of mudstone. Sandstone beds are amalgamated in the palaeovalley axis and include progressively thicker intervening muddy caps toward the western wall. Sandstones are massive and present a large amount of soft-sediment deformation (small scale slumping, sand injections, flame structures and convolute lamination; Fig. 30). The abundance of massive beds suggest that the low density portion of the turbidity currents went further downstream leaving behind them only the concentrated part of the bi-partite flow in a more proximal reach. These sandstone beds were possibly deposited by waning depletive flow. The upper part of Ts3 is missing throughout most of the palaeovalley due the erosional contact between stages IV and V.

Figure 30: Soft-sediment deformation (injections), a common feature of the Ts3.

5.3.2. Vidal Ramos Palaeovalley

The turbidite strata that occur at the top of the Vidal Ramos palaeovalley fill, just above the Mafra Shale, include two stages characterized by turbidite sand-sheets (T1 and T2).

T1 Stage comprises a thickening- to thinning-upward succession with the thickest beds deposited along the palaeovalley axis (Fig. 31). Both lower and upper contacts with R1 and R2 stages, respectively, are gradual. In this context, the basal coarsening-upward trend is interpreted as a result of a climate warming, gradually increasing ice melt rate and therefore sediment discharge from the glaciers into the valley. Conversely, the

upper thinning-upward trend is interpreted as a subsequent climate cooling, glacial melting rate gradually slowed down, resulting in less sediment input and consequently thinner sandstone beds.

Figure 31: Thick sandstones at the mid portion of the T1 Stage.

T1 sandstones are fine-grained, include mud clasts and are usually massive, although rare plane bedding and/or ripple cross-lamination are sometimes present near the top of the beds. Sandstones are composed of quartz, muscovite and biotite, quartz overgrowth is common and they present almost no permeability and porosity (Fig. 32).
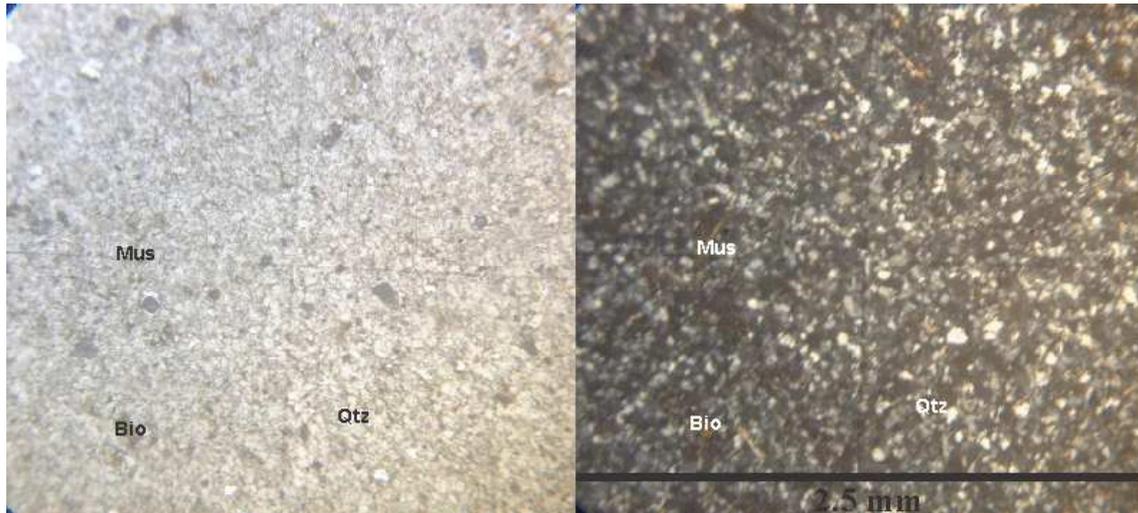
Figure 32: Fine-grained sandstone composed of muscovite, biotite and quartz grains with no pores within them (T1 stage)

T2 Stage lay directly onto an erosional contact with R2 Stage. It is composed of 0.2 to 0.8 m thick, medium- to fine-grained sandstone that may grade to thin caps of mudstone. Sandstones tend to be amalgamated, and therefore difficult to be discriminated, along the palaeovalley axis (Fig. 33), but better defined towards the palaeovalley margins (Fig. 34). Upwards, sandstones become gradually thinner and intervening mudstone beds thicken.

Each bed tends to display the following succession of features: massive, including few mudclasts, at the base; plane-bedding and ripple cross-lamination toward the top (Fig. 33b, c). These beds were interpreted as the product of bi-partite turbidity currents related to waning-depletive surges. The centre of the palaeovalley concentrated the largest part of the turbidity currents therefore producing amalgamated sandstone beds due to the erosion of previously deposited, fine-grained caps.
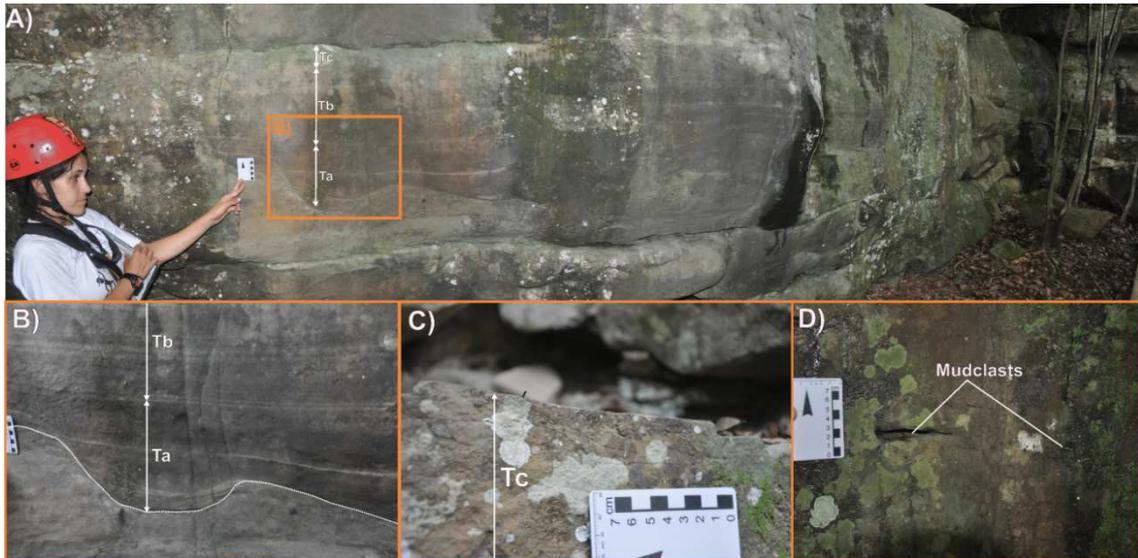
Figure 33: (A) T2 stage amalgamated sandstone beds at the centre of the valley. These sandstones may include (B) a massive and plane-bedded base as well as (C) a ripple cross-laminated top. (D) Mud clasts are common within the beds.

Figure 34: Well defined T2 stage beds due to presence of thin mud caps near the confining southern wall.

Sandstone beds related to this stage are coarser-grained compared to previous stage. Its composition includes quartz, muscovite, biotite and k-feldspar (Fig. 46). Besides their relative stratigraphic position, its coarser-grained nature and the presence of K-feldspar provides the clues to differentiate T1 to T2 strata.

The thicker and coarser-grained sandstone beds of the T2 Stage suggest a warmer climate relative to previous stages as climate improvement could increase melting rates and sediment input into the palaeovalley.
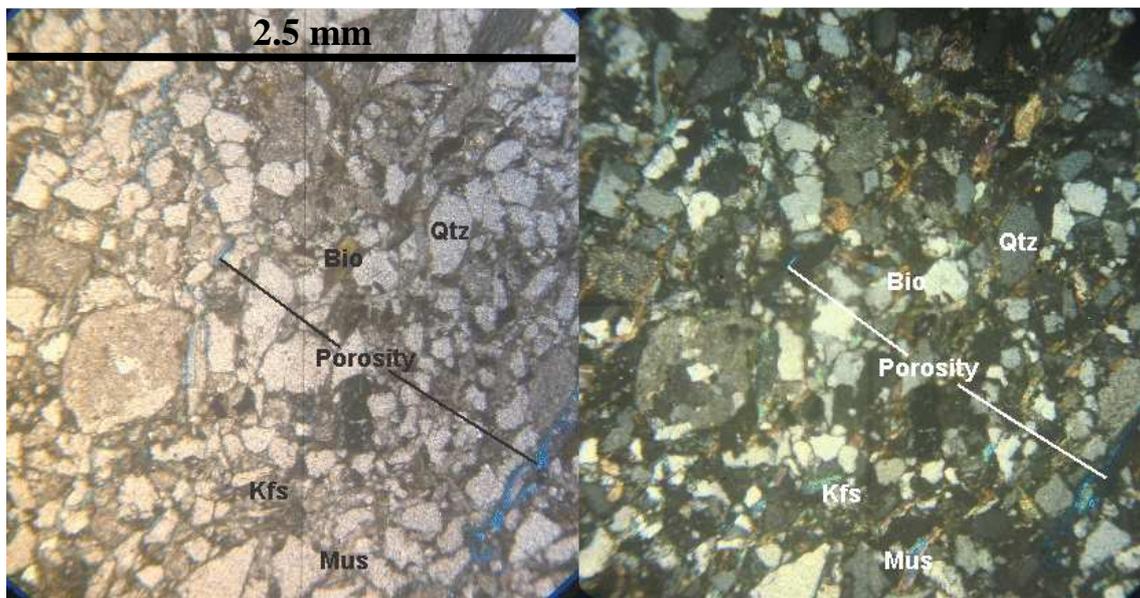
Figure 35: T2 stage sandstone. Note its coarser-grained nature and higher permeability relative to previous stages as well as the presence of K-feldspar.

5.4. Onlap

The Quebrada de las Lajas palaeofjord is the only studied site exposed enough to allow the analysis of the stratigraphy from wall to wall, hence enabling a detailed overall view of the sand-sheets. Using some marker beds it was possible to realize that at the centre of the palaeofjord sandstone beds tend to be thicker and amalgamated while mudstone tends to be thinner, and the reverse towards the walls (Fig. 36). This sort of behaviour indicates a type A onlap onto both fjord walls. Besides, sandstones tend to onlap

upwards for some distance parallel to the confining wall before pinching out. That is a common feature of a draping-onlap.
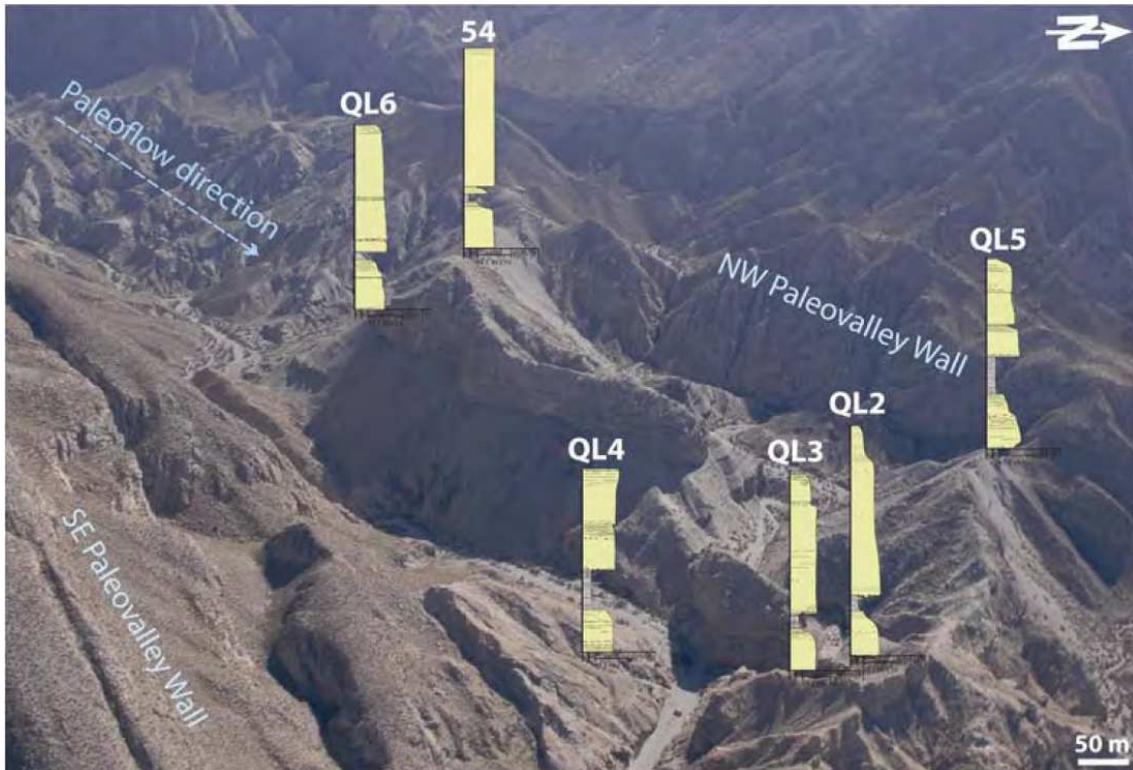
Figure 36: Aerial view and location of the sedimentary logs at the Quebrada de Las Lajas palaeofjord. Logs 54, QL6, QL2 and QL3 are located near the palaeofjord axis whereas logs QL5 and QL4 are closer of the palaeofjord walls.

Toward the walls sandstone becomes richer in structures, suggesting denser axial flows relative to their lateral flow counterpart. The influence of the walls on sedimentation is apparent by the eventual presence of bidirectional structures and more common small scale slumped beds.

At Quebrada Grande, sandstone beds get thinner to lastly pinchout toward the western wall of the palaeovalley whereas mudstone packages become thicker. Sandstone extends some distance up the onlaping surface. These features suggest a type 2 draping-onlap and a class 2 slope base trajectory (aggradational onlap).

It is important to notice that the basement rock is a debris-flow package that includes both turbidite-derived, fine-grained strata and very large limestone blocks. This heterogeneous composition might impose an accidental basin floor, instead of a perfect U-shaped valley.

In the Vidal Ramos palaeovalley turbidite sandstone beds tend to pinchout toward the confining walls. However, a gradual enlargement of the depositional area of the subsequent stages is also apparent (Fig.20). A differential compression of the Mafra shale due to its own weight and to the deposition of the overlying sandstone strata, in both cases thicker towards the palaeovalley axis, could explain the angular relationship between the MS and R1 stages and the increasingly wider onlap of the T1, R2 and T2 strata onto this deformed, upper surface of the Mafra shale, leading therefore to a draping-offlap geometry. In this sort of geometry, sand bodies extend gradually further than below, always thinning toward its closure, so leading to less sharp sandbody termination (featheredge pinchout) with the beds turning parallel to the onlap surface.

6. GENERAL DISCUSSION

As previously commented, the turbidite sand sheets have been deposited in a wide range of confinement, from extremely (Quebrada de las Lajas) through moderately (Quebrada Grande) to loosely (Vidal Ramos) confined depositional settings. However, the turbidite sand sheets share some common features, but also present striking differences. Below, these resemblances and dissimilarities are highlighted.

6.1. Similarities

As suggested by their proglacial fill (dropstones and iceberg-related scours) and attested by direct evidence, such as glacial striation on basement rocks (Vidal Ramos) and fjord-

like morphology (Quebrada de Las Lajas), all valleys are supposed to represent glacier carving features formed during a preceding glacier incursion. And in all areas, the succession is preserved as more or less open synclines due lateral facies changes and basement glacial erosional morphology.

In both Paraná and Paganzo basins, the intercalation of sand-prone and thin-bedded turbidites was ascribed to high-frequency climate changes. In this context, the coarse-grained inflow was reduced during cooler (muddy intervals) and increased during warmer (sandier intervals) periods, when ice melting rate was higher, therefore enhancing the expansion of the turbidite sandstone sheets.

Sandy turbidites pinch out towards the confining walls where they are sealed by thicker mudstone caps. Similarly, massive and laminated sandstones decrease and increase towards the borders, respectively. And all turbidite systems display a narrow, unidirectional paleocurrent trend. These pieces of evidence are a direct response to the confined setting that forced turbidity currents to flow along the valley axis, therefore producing a higher proportion of massive and thick sandstones along it and laminated and thin sandstone towards the borders, as well as general unidirectional sense of sand transport.

The usual normal grading of the turbidites (sand to mud couplets) points to surges rather than long-lived flows. The longitudinal change from sandier and thicker (proximal) to heterolithic and thinner (distal) turbidite beds indicates waning-depletive turbidity currents. These features are compatible with fjord- or valley-head delta prograding into relatively steep valley floor and triggering recurrent sand remobilization from delta front to deep marine settings as depletive, sand rich surges.

6.2. Differences

The studied sand-sheet turbidites of the Paganzo Basin occur after the maximum, deglacial transgressive system tract and ensuing maximum flooding, hence pointing to turbidity flows related to highstand systems tract. On the other hand, the turbidite succession in the studied section of the Paraná Basin overlies a relative conformity, is positioned below the maximum flooding surface, hence represents an early lowstand system tract. Therefore, although all examples present regressive trends (coarsening- and thickening-upward successions), they are related to distinct stratigraphic positions, including the lowstand, forced regressive turbidite strata of the Vidal Ramos palaeovalley, and the highstand, normal regressive turbidite packages of the Las Lajas palaeofjord and Quebrada Grande palaeovalley.

The degree of confinement played its role in terms of palaeocurrent dispersal and facies associations. For instance, in the Quebrada de Las Lajas and Quebrada Grande areas the usually narrow unidirectional paleocurrent trend of all studied is more often affected near the walls due to current rebound. Besides, in these largely confined setting, and near their steep walls and associated hanging valleys, more frequent remobilization produced rock fall, and slump and debris flows deposits.

Distinctly of the Quebrada Grande and Vidal Ramos wider palaeovalleys, the thick mudstone intervals present in the Quebrada de las Lajas palaeofjord and the upward change in the sand-to-mud content of its fill were tentatively interpreted as a result of the existence of some sort of depositional barrier along the palaeofjord axis. This longitudinal confinement would trap the fine-grained sediments brought by the turbidity currents until the spill-point was achieved. Therefore, it is here considered that the

larger the confinement, the higher the possibility of the existence of effective transversal barriers.

The type of onlap also seems to be dependent on the degree on confinement. Higher confinement favours draping-onlap. On the other hand, low confinement favours draping-offlap.

7. CONCLUSIONS

The degree of confinement seems to have controlled some features related to the deglacial sedimentary fill of glacial palaeovalleys. The overall geometry of their fill is controlled by the U-shaped morphology and resembles synclines that are as tighter as more confined the valley. In terms of the associated turbidite sand sheets, high frequency climate changes have driven vertical facies trends (alternation of sandy and thin bedded turbidites). Through warmer intervals higher ice melting rate produced a larger growth of the turbidite sandstone sheets, and the opposite during cooler intervals. In general, confinement forced turbidity currents to flow along the valley axis depositing mostly massive and thick sandstones along it, and laminated and thin bedded sandstone towards the borders. Confinement has also produced unidirectional trend of sand transport. As a result, these confined turbidite sandy sheets show good vertical connectivity at the centre of the valleys, decreasing toward the borders. Turbidites main features support an association with fjord- or valley-head deltas prograding into relatively steep valley floor that favoured recurrent sand remobilization from delta front to deep marine settings as turbidity, depletive surges.

The studied examples of sand sheet turbidites of the Paganzo and Paraná basin are related to distinct stratigraphic positions, including the lowstand turbidite strata of the Vidal Ramos palaeovalley and the highstand turbidite packages of the Las Lajas

palaeofjord and Quebrada Grande palaeovalley. The higher degree of confinement of the Las Lajas Palaeofjord, and secondly of the Quebrada Grande palaeovalley, altered the unidirectional paleocurrent trend due to more often current rebound of the turbidity currents and large-scale remobilization near the confining walls. Besides, the facies succession suggest that the larger the confinement, the higher the possibility of the existence of effective transversal barriers trapping mud in more proximal depositional settings.

8. ACKNOWLEDGMENTS

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