



The late Paleozoic Ice Age along the southwestern margin of Gondwana: Facies models, age constraints, correlation and sequence stratigraphic framework

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ABSTRACT

The southwestern margin of South America offers a complete record of the Late Paleozoic Ice Age (LPIA) that affected the Gondwana supercontinent. The tripartite division of LPIA glacial episodes has been refined with the help of new radiometric dates and biostratigraphic (flora and fauna) zonations in recent years to five shorter-lived discrete events: 1. Latest Devonian-earliest Tournaisian, 2. Tournaisian, 3. Visean, 4. Serpukhovian-Early Bashkirian, 5. late Pennsylvanian-earliest Permian. These glacial events are capped by postglacial transgressive deposits with marine fauna. The unbalanced preservation potential of the glacial deposits, skewed toward the glaciomarine sediments, provides an uneven stratigraphic record with few cases of continental glacial sedimentation, confined to the Serpukhovian-Early Bashkirian event, and numerous examples of glacial sedimentation in marine environments. Glacial sedimentation in marine settings has been grouped in two main facies associations: a valley-glacier-retreat (fjord) facies association and a submarine-retreat (glaciomarine apron) facies association in open-marine areas.

Transitional facies, correspondent to those formed by the flooding of valleys during postglacial transgressions, are widely distributed along the ProtopreCORDillera in western Argentina, where paleofjord successions are well exposed particularly in western Paganzo Basin, and mapped in subsurface in the Tarija basin. A general paleofjord model includes (from base to top) the following stages: (i) Incision of paleovalley and deposition of subglacial diamictites in ice contact deltas, (ii) Early Transgressive stage characterized by resedimentation of subglacial material by subaqueous sediment gravity flows and slumps in proglacial settings, (iii) Maximum flooding (late transgressive stage) dominated by black shales or laminated mudstones related to a marine incursion that flooded valleys; normal marine or brackish conditions may dominate this stage and (iv) Highstand: progradation of a fluvial-deltaic system including in some cases Gilbert-type deltas. In glaciomarine apron environments, the basal facies includes massive clast-supported conglomerates, with few striated and polished clasts, followed by fining-upward successions including thinly bedded diamictites with ice-rafted debris (IRD) and locally contorted sandstone masses in diamictite beds, indicative of mass-emplacment mechanisms. The presence of inter- and intratill pavements suggests glacial advance/retreat fluctuations along the basin margins.

Deglaciation sequences, reflecting deposition mainly during the retreat of ice sheets, ice caps and alpine glaciers and successive deglaciation, can be used as operational tools for the analysis of glacial successions in southwestern Gondwana. They are characterized as rather simple upward-fining successions in open marine settings as exemplified in most of the Calingasta-Uspallata Basin, Claromecó Basin (and adjacent Ventana foldbelt, VFB) and central portions of the Paraná and Karoo basins. In more proximal areas (i.e., paleofjords) this vertical trend is commonly punctuated with deltaic wedges fed by nearby provenance areas.

The late Paleozoic glacial-related successions of southwestern Gondwana exhibit a common tripartite motif, equivalent to second-order sequences with estimated durations of 10–80 Myr. The lower section corresponds to glacial and glacially-influenced diamictites; the middle interval is initiated with postglacial transgressions. The lower and middle intervals correspond to the deglaciation sequence as described and identified in several basins of Gondwana. Finally, the upper term includes coastal progradation, followed in some places by continentalization, accompanied in many sectors by increasing aridization. Examples of second-order sequences can be identified in the thick late Paleozoic successions of the Paraná and Karoo basins and in the VFB. Thinner second order sequences can be identified in the Calingasta-Uspallata, Rio Blanco, Paganzo and San Rafael basins. In the Paganzo and San Rafael basins the middle interval is also punctuated by short lived marine incursions. The basal sequence boundary is commonly an abrasion surface (glacial erosion surface, GES) developed on bedrock. Deglaciation sequences are assigned to third order sequences made up, when present, of a thin lowstand system tract (LST) of subglacial deposits followed upward by thick glaciomarine and glacially influenced sediments. These facies are part of a thick transgressive systems tract (TST) that culminates with marine shales that reflect interglacial or postglacial conditions during ice retreat. Thus, the deglaciation sequences are proposed to be third order sequences made up of LST-TST or exclusively TST.

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1. Introduction

The fluctuations in ice extent across Gondwana during the late Paleozoic are recorded in many basins by thick deposits of glacial and glaciomarine diamictites followed by postglacial mudstones. Glacial deposits of late Paleozoic age occur in all the subcontinents of Gondwana: South America, Africa, Antarctica, Falkland/Malvinas Islands, India and Australia (Crowell and Frakes, 1975; Crowell, 1978; 1999; Hambrey and Harland, 1981; Eyles, 1993; Deynoux et al., 1994; Isbell et al., 2003; 2012; Fielding et al., 2008; López-Gamundí and Buatois, 2010; Rosa and Isbell, 2020). A long-term control on deglaciation was climatic warming as the Gondwana supercontinent drifted over the South Pole (Powell and Li, 1994; Opdyke et al., 2001) resulting in a diachronous shifting of glacial centers across the supercontinent (Crowell, 1978; Caputo and Crowell, 1985). Other factors include variable concentration of atmospheric CO₂ (Horton et al., 2007; Frank et al., 2008; Montañez et al., 2007; 2016; Myers, 2016), and/or elevation of mountain belts (Isbell et al., 2012; Goddérís et al., 2017). Moreover, the waxing and waning of each glacial episode of the late Paleozoic Ice Age (LPIA), as well as the shorter duration glacial/interglacial cycles, resulted from a complex interplay of local, regional, and global factors.

In the last two decades significant progress has been made in the study of the late Paleozoic glacial episodes present along the southwestern margin of Gondwana (Starck and del Papa, 2006; López-Gamundí, 2010; Limarino et al., 2006; 2014a, b; Henry et al., 2008, 2014; Bache et al., 2012; Vergel et al., 2015; Rousse et al., 2018; Ezpeleta et al., 2020; Pauls, 2020; Valdez Buso et al., 2020). Particularly major advances in chronostratigraphy with new and precise radiometric ages in early postglacial sediments help to cap the age of these episodes and their subcycles or events. This wealth of new absolute ages, coupled with facies and sequence stratigraphic studies, and more refined biozonations help to provide a much better understanding of the type, extent and duration of the late Paleozoic glacial episodes and events. Particularly insightful has been the use of absolute ages to constrain invertebrate, macro- and microfloral biozones around the glaciation - postglacial transition (Césari et al., 2011) in the context of the icehouse-to- greenhouse turnover. Also the discrimination of global and regional driving forces in the formation and demise of the late Paleozoic glaciers, and the subsequent paleoclimatic evolution along the southwestern Gondwana margin, has been an active field of research (Fielding et al., 2008; Isbell et al., 2012; Montañez and Poulsen, 2013; Limarino et al., 2014b; Pauls, 2020).

Widespread glaciation occurred throughout much of western Gondwana (Fig. 1) including five, possibly six, synchronous glacial events separated by synchronous ice-minimal or non-glacial intervals in the Río

Blanco, Calingasta-Uspallata, San Rafael, western Paganzo, Tarija, Sauce Grande (VFB), Chaco-Paraná and Paraná basins (cf. Limarino et al., 2014b). Although glaciation was originally reported for the eastern Paganzo Basin and the eastern portion of the Rio Grande do Sul Shield in the southeastern portion of the Paraná Basin, Brazil (Rocha Campos et al., 2008; Tedesco et al., 2016), no glacial signatures occur in strata from those regions (Moxness et al., 2018; Pauls et al., 2019, this volume; Fedorchuk et al., 2019; Pauls, 2020). Regionally, a stepwise deglaciation began in the Bashkirian in the western basins (Río Blanco, Calingasta-Uspallata, San Rafael, western Paganzo) and proceeded eastward reaching the Chaco-Paraná and Paraná basins in the latest Pennsylvanian (López-Gamundí et al., 1994; López-Gamundí 1997; Limarino et al., 2002, 2006, 2014; Henry et al., 2008, 2010; Gulbranson et al., 2010; Isbell et al., 2012; Cagliari et al., 2016; Moxness et al., 2018; Griffis et al., 2018; Pauls et al., 2019; Pauls, 2020). Following Bashkirian deglaciation, the western Argentine basins transitioned into an early to middle Pennsylvanian humid temperate climate and then into a late Pennsylvanian to early Permian semiarid/arid climate as indicated in some areas by the occurrence of coal-bearing strata (López-Gamundí et al., 1993) that transitioned upward into eolian erg deposits, and/or by changing mudrock geochemistry (chemical index of alteration; Gulbranson et al., 2010; Limarino et al., 2014b; Pauls et al., 2019; Pauls, 2020). During this western climate transition, glaciation continued farther to the east at similar paleolatitudes in the Claromecó, Chaco-Paraná and Paraná basins and in the Tarija Basin to the north (cf. Limarino et al., 2014b; Isbell et al., 2012; Moxness et al., 2018; Pauls, 2020). In the central and northeastern portion of the Paraná Basin, glaciation continued into the early Permian (Asselian-Sakmarian; cf. Mottin et al., 2018; Jurigan et al., 2019).

The record of the LPIA in southwestern Gondwana seems to cluster around three distinct episodes (Veevers and Powell, 1987; López-Gamundí, 1997; Isbell et al., 2003). These episodes were constrained by Isbell et al. (2003) who propose that the three LPIA episodes were time constrained as follows: Episode I, ~375–350 Ma; Episode II, ~315–307 Ma; and Episode III, ~299–276 Ma. Each episode is loosely defined not as much for its duration but rather by its demise. Also each episode contemplates the possibility of including two or more shorter-lived glacial events.

Paleogeographically these glacial episodes are broadly distributed, but not limited to, specific area in southwestern Gondwana (Fig. 3):

- Episode I (latest Devonian - Tournasian) confined to a broad SW-NE belt across the central and northern parts of South America (Caputo, 1984; Caputo and Crowell, 1985; Díaz Martínez and Isaacson, 1994; Eiras et al., 1994; Góes and Feijó, 1994; Díaz Martínez et al., 1999; Carlotto et al., 2004; Cerpa et al., 2004; Isaacson et al., 2008; Caputo et al., 2008; Borghi and Lobato, 2012; Playford et al., 2012).

- Episode II (Tournaisian and Namurian-Bashkirian) present mostly along the southwestern margin of Gondwana in southern Bolivia, west central and south Argentina and eastern Paraná Basin in Brazil (López-Gamundí et al., 1987; Limarino et al., 2002; Henry et al., 2008; Rosa et al., 2019).

- Episode III (late Pennsylvanian-earliest Cisularian) with a significant stratigraphic record in the Paraná basin, Falkland/Malvinas Islands, Ventana Foldbelt (VFB)/Claromecó basin in South America and Karoo and Kalahari basins in Southern Africa (Rocha Campos and dos Santos, 1981; França, 1994; Rocha-Campos et al., 2008; Visser, 1987; 1989; von Brunn, 1996; Trewin et al., 2002; Andreis and Torres Ribeiro, 2003; Isbell et al., 2008; Horan et al., 2018).

This contribution summarizes the latest advances on characteristics, distribution, and timing of glaciation across southwestern Gondwana, with particular emphasis on the recent progress on facies models, absolute dating and sequence stratigraphy. We emphasize the paleofjord depositional model proposed principally for the basins located along and adjacent to the paleo Pacific margin (i.e., Paganzo and Tarija basins). New radiometric dates, refined biozones and sequence stratigraphic tools, allowed to expand and improve our understanding of the

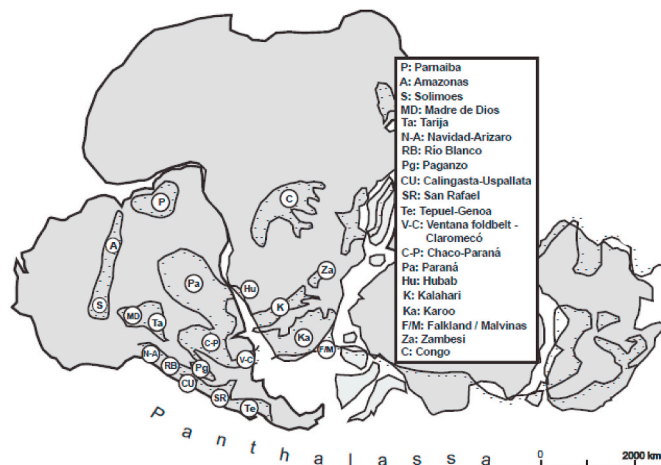


Fig. 1. Late Paleozoic SW Gondwana reconstruction with main sedimentary basins.

relationships between the glacial record present in basins along the southwestern Gondwanan margin (Calingasta-Uspallata, Río Blanco, Paganzo and Tarija basins) and their counterparts in the interior of the continent (e.g., Paraná Basin and contiguous African basins). Thus, more robust intra- and interbasinal correlations are proposed.

2. Late Paleozoic ice age: spatial and temporal distribution

Although basically considered a rather continuous and diachronous glacial period (Frakes et al., 1992), progress in the knowledge of the late Paleozoic glaciation in southwestern Gondwana demonstrated that the LPIA was formed by a series of glacial episodes punctuated by interglacial intermissions. The LPIA deposits were distributed in several basins that showed diverse paleogeographic contexts and associated basin margin configurations which conditioned the characteristics of the glacial and subsequent postglacial sediments. We have subdivided these basins subjected to glacial conditions according to their position with respect to the active proto-Pacific margin of the Gondwana supercontinent (Figs. 1 and 2) (Limarino and Spalletti (2006): 1) The western marginal basins behind the margin, 2) interior basins developed between the marginal basins and the Pampean basement arch, 3) the Chaco-Paraná and Ventana fold belt (VFB) - Claromecó basin and 4) the cratonic Paraná basin (Fig. 2).

The western marginal basins developed along the proto-Pacific margin include the San Rafael, Calingasta-Uspallata, Río Blanco, Navidad-Arizaro, Tarija and Madre de Dios basins (Figs. 1 and 2).

In South America, paleogeographically located in the southwestern portion of Gondwana, the LPIA has been subdivided in three glacial episodes (Episode I, II and III; López-Gamundí, 1997; Isbell et al., 2003). Five shorter-lived glacial events have been discriminated in the stratigraphic record in this paper based on the information published in recent years (Caputo et al., 2008; Pérez Loinaze et al., 2010; Anderson, 2011; López-Gamundí et al., 2013; di Pasquo et al., 2019). Each of these events (Fig. 3) is terminated by a postglacial marine flooding that can be recognized as a general mechanism whose stratigraphic record is correlatable at basinal and even interbasinal scale. Each glacial episode tentatively encompasses two or more glacial events. In the case of South America five main glacial events are recognized during the: 1. Latest Devonian - earliest Tournaisian, 2. Tournaisian, 3. Viséan, 4. Serpukhovian - early Bashkirian 5. late Pennsylvanian - earliest Permian [Fig. 3].

In this chapter we revise the new age constraints as evidenced by recent studies in the key basins with glacial deposits. Although there is abundance of non-age diagnostic cold resistant marine invertebrates and floras in interglacial and postglacial fine-grained sediments, some of the biozones defined (Fig. 4) have helped to provide a better

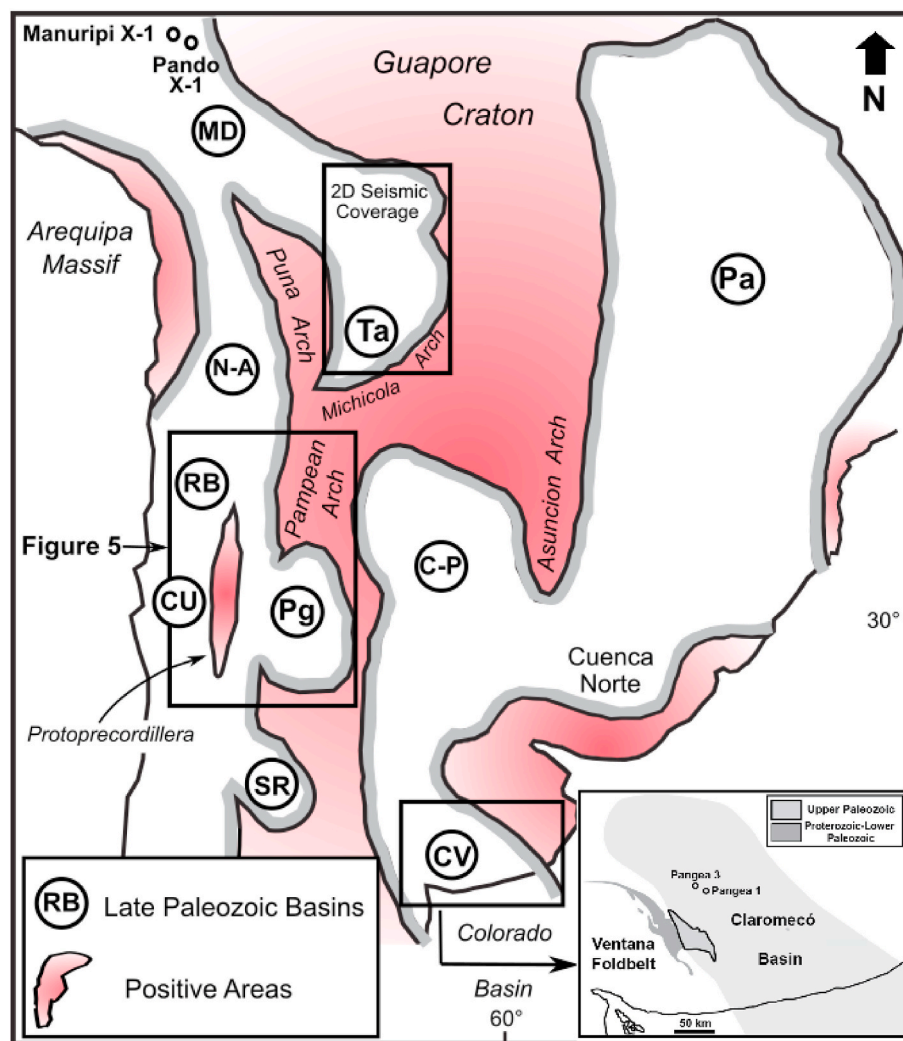


Fig. 2. Generalized Late Paleozoic paleogeography of SW Gondwana, modified from Limarino and Spalletti (2006). C-P: Chaco-Paraná Basin, CU: Calingasta-Uspallata Basin, CV: Claromecó Basin (and adjacent Ventana Foldbelt, VFB), MD: Madre de Dios Basin, N-A: Navidad-Arizaro Basin, Pg: Paganzo Basin, RB: Río Blanco Basin, SR: San Rafael Basin, Ta: Tarija Basin.

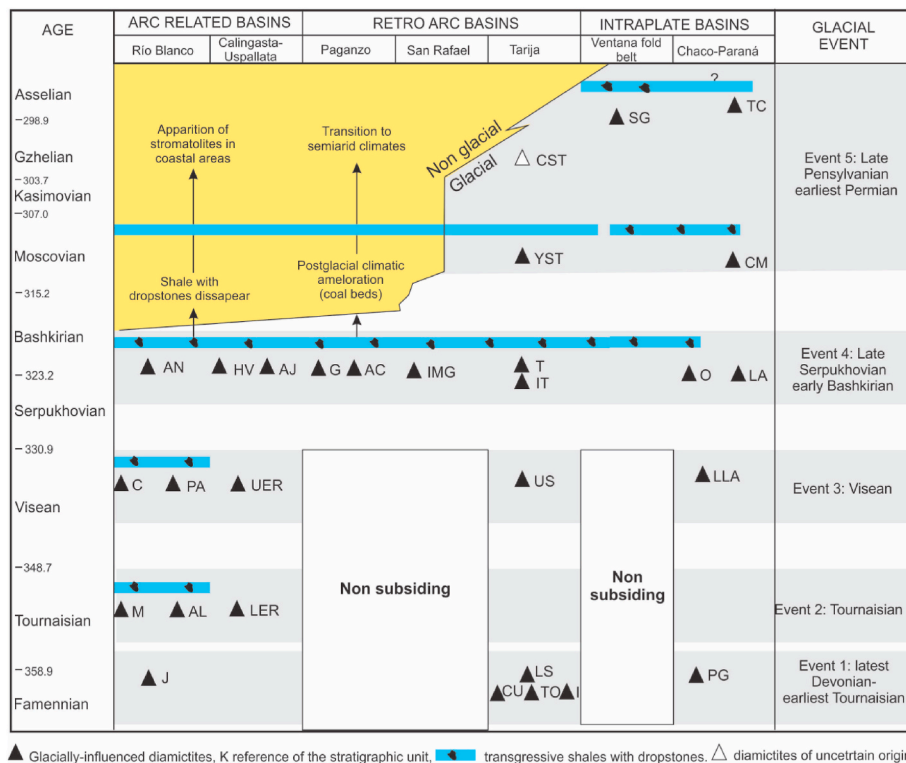


Fig. 3. Stratigraphic chart with late Paleozoic glacial events (black triangles) in SW Gondwana (Chaco-Paraná Basin includes also the Paraná Basin). Stratigraphic units in alphabetic order: AC: Agua Colorada, AJ: Agua de Jagüel, AL: Agua de Lucho, AN: Agua Negra, C: Cortaderas, CM: Campo Murão, CST: Cai-guamí Member (San Telmo Formation), CU: Cumaná Formation, G: Guandacol, HV: Hoyada Verde, I: Itacua, IMG: Glacigénico Member (El Imperial Formation), J: Jagüel, O: Ordóñez, LA: Lagoa Azul Formation, LER: lower El Ratón Formation, PA: Punta del Agua, PG: Punta Grossa, M: Malimán, SG: Sauce Grande, T: Tarija, TC: Taciba Formation, UER: upper El Ratón Formation, US: upper Saipurú Formation, YST: Yaguacua Member (San Telmo Formation).

chronostratigraphic range for the end of each of the glacial episodes. In isolation and without the help of radiometric dating, it is exceedingly difficult to accurately estimate the age of the glacial deposits and the glacial-postglacial transition.

2.1. Event 1 (Latest Devonian-earliest Tournaisian)

In southern South America, the latest Devonian-earliest Tournaisian glaciation was recognized in the Madre de Dios, Tarija and Paraná basins, and potentially in the Río Blanco Basin (Fauqué and Limarino, 1991). Further north, this glacial event is recorded in the Amazonas (Cunha et al., 1994) Solimões (Eiras et al., 1994) and Parnaíba (Góes and Feijó, 1994; Borghi and Lobato, 2012; Playford et al., 2012; Caputo and dos Santos, 2019) basins.

Probably one of the best characterized outcrops of the latest Devonian-earliest Tournaisian event occurs in the Madre de Dios Basin. In the eastern part of the basin the Cumaná Formation consists of glacially-influenced diamictites with faceted and striated clasts, along with shales with dropstones (Díaz Martínez and Isaacson, 1994). There, Díaz Martínez and Isaacson (1994), Díaz Martínez et al. (1999) and Isaacson et al. (2008) reported a palynological association belonging to the Late Famennian *Retispora lepidophyta*-*Indotiradites explanatus* palynozone (Fig. 4). This assemblage is associated with marine elements like *Umbellaspheeridium saharicum* Jardine et al. (Vavrdová et al., 1993; Díaz-Martínez et al., 1999; Lakin et al., 2016).

The lower part of the Retama Group contains two diamictite-bearing intervals. The lower diamictite section corresponds to the Toregua Formation (Madre de Dios Basin), which has been correlated to the diamictites found in the Cumaná Formation. The occurrence of *Retispora lepidophyta* and *Umbellaspheeridium saharicum* in the Pando X-1 and Manuripi X-1 wells (Fig. 2) allowed the strata to be assigned a latest Devonian-earliest Mississippian age (Isaacson et al., 1995; Lakin et al., 2016). Diamictites containing striated and faceted clasts together with shales bearing dropstones were reported from the Saipurú Formation in the Tarija Basin (Suárez-Soruco and López-Pugliesi, 1983; Caputo et al., 2008), the presence of palynomorphs belonging to the *Retispora*

lepidophyta zone is indicative of a Famennian age for this glacial interval (Suárez-Soruco and López-Pugliesi, 1983).

Diamictites and dropstone-bearing shales found in the Itacua Formation of the Tarija basin (Fig. 1) were also assigned to the latest Devonian-early Mississippian glaciation (Lakin et al., 2016). However, palynological studies by di Pasquo (2007) suggest an early Visean age for the diamictites in southern Bolivia. A detailed analysis carried out by Wicander et al. (2011) 500 km to the north, proposed a Late Devonian age for the Itacua Formation in the Tarija Basin recognizing the *Retispora lepidophyta*/*Knoxisporites literatus* (LL), *Retispora lepidophyta*-*Indotiradites explanatus* (LE) and *Retispora lepidophyta*-*Verrucosporites nitidus* (LN) miospore biozones of Western Europe (Wicander et al., 2011). Streel et al. (2013) confirmed a latest Devonian to the diamictites of the Itacua Formation although constraining its age to the LE and LN biozones.

Moreover, taking into account the presence of several levels of diamictites and the vertical distributions of the biostratigraphic zones, several glacial and interglacial events could be represented in the Itacua Formation (Wicander et al., 2011; Streel et al., 2013).

In the Paraná Basin, Itararé Group diamictites unconformably resting on the Devonian Ponta Grossa Formation were originally referred to the Pennsylvanian. However, the presence of palynomorphs of the LN Zone (Loboziak et al., 1995) in the lower diamictite interval allowed for identification of a probable latest Devonian-Early Mississippian glaciation in the Paraná area (Caputo et al., 2008; Lakin et al., 2016). However, it is uncertain whether these are contemporaneous palynomorphs or reworked from the underlying Devonian strata (cf. Streel et al., 2013; Lakin et al., 2016).

Latest Devonian glaciation was also reported in the Jandiatuba Formation of the Solimões basin, Curiri Formation in the Amazon basin and Cabeças Formation of the Parnaíba - Marajo basins (Caputo, 1985; Caputo et al., 2008; Isaacson et al., 2008; Caputo and dos Santos, 2018, 2019).

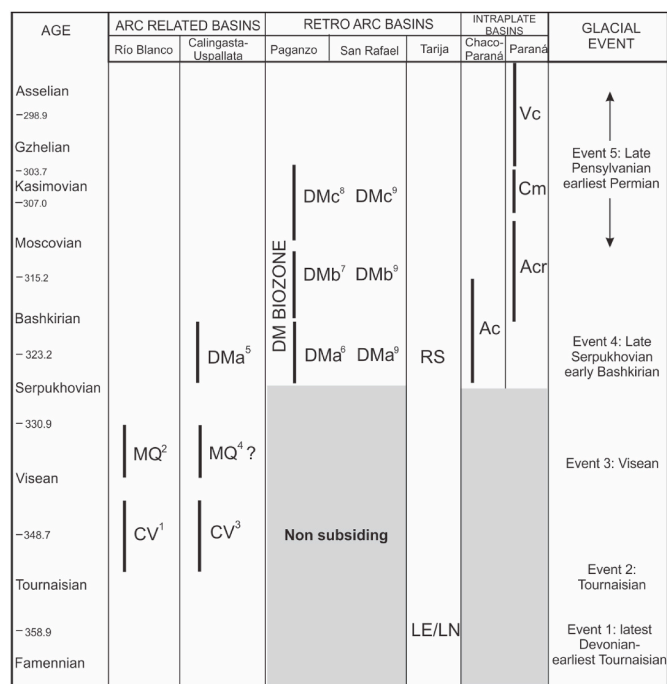


Fig. 4. Chart of palynological biozones associated with the late Paleozoic glacial events of southwestern Gondwana. CV: *Cordylsporites-Verrucosporites* p.p. Biozone; MQ: *Reticulatisporites magnidictyus-Verrucosporites quasigobettii* Biozone; DM: *Raistrickia densa-Convolutispora muriornata* Biozone; DMA: Subzone A; DMb: Subzone B; DMc: Subzone C; LE/LN: *Retispora lepidophyta-Indotriradites explanatus/Retispora lepidophyta-Verrucosporites nitidus* biozones, Itacua Formation; RS: *Raistrickia radiosa-Apiculatisporites spinulistratus*, Itacuami Formation; Ac: *Anapiculatisporites concinnus* Biozone; Ordoñez Formation; Acr: *Ahrensisporites cristatus* Biozone; Cm: *Crucisaccites monoletus* Biozone; Vc: *Vittatina costabilis* Biozone. 1: Malimán Formation; 2: Cortaderas Formation; 3: El Ratón Formation; 4. El Planchón Formation; 5: El Paso Formation; 6. Guandacol Formation and equivalent units; 7–8: Tupe Formation and equivalent units; 9. El Imperial Formation.

2.2. Event 2 (Tournaisian)

In western Argentina (Río Blanco Basin, Fig. 1), the Agua de Lucho Formation exhibits several levels of shales with dropstones and diamictites with large (up to boulder size) clasts, some of them faceted and striated (Ezpeleta et al., 2020). This diamictite facies has been interpreted as deposited in glaciomarine proximal to distal environments. Ezpeleta et al. (2020) identified proglacial accumulations linked to fan deltas, grounding-line fans and ice rafting in more distal areas. Although the succession was considered Late Devonian (Prestianni et al., 2011), a mid to late Tournaisian age for the glaciation was proposed by the presence of the miospore *Waltzisporea lanzonii* Daemon 50 m above the base of the uppermost diamictite bed (Prestianni et al., 2015). The rare specimens of *W. lanzonii* were identified in a palynological assemblage composed of up to 99% of Devonian taxa, which were considered all reworked. The brachiopod *Azurduya chavelensis* (Amos) Cisterna & Isaacson was found above and below the levels with palynomorphs (Sterren et al., 2013; Ezpeleta et al., 2020).

A second Tournaisian unit bearing putative glaciomarine deposits occurs in the Malimán Formation in which Pazos (2007) identified two intervals of diamictites associated to shales with dropstones interpreted as glaciomarine deposits. Two levels of diamictites were also identified in the Malimán Formation by Limarino and Césari (1993), but in this case diamictites were interpreted as formed by debris flows in shelf settings. However, the presence of some isolated clasts (up to 9 cm in diameter) in laminated shales may indicate ice-rafted material. Based on palynological content, Césari and Limarino (1995) suggested a

Tournaisian age for this section. Later, a late Tournaisian age for the lower part of the Malimán Formation was proposed by Amenábar et al. (2009) based on new palynological assemblages. The presence of invertebrates also supports a Tournaisian age (Sabattini et al., 2001; Taboada and Shi, 2009; Sterren et al., 2013).

In the Calingasta-Uspallata Basin (Fig. 1) the presence of a relatively thin (about 50 m) stratigraphic interval of matrix-supported conglomerates, diamictites and shales with dropstones was reported by Colombo et al. (2014) in the lower member of the El Ratón Formation. Although tillite levels were not recognized, the presence of dropstones, diamictites and synsedimentary deformation led Colombo et al. (2014) to interpret the lower member as deposited in deltaic environments very probably developed in a fjord environment. Later, studies by Milana and di Pasquo (2019) reported glacial striation of clasts included in the conglomerates of the lower part of the El Ratón Formation, although suggesting that probably the striated clasts were derived from a previous glacial cycle not preserved in the El Ratón Formation. Milana and di Pasquo (2019) include this glacial event in the Tournaisian (see Fig. 11, Milana and di Pasquo, 2019). The age of the Lower Member of the El Ratón Formation is inferred from palynological studies carried out by Amenábar and di Pasquo (2009) who reported the presence of *Anapiculatisporites amplus* Playford & Powis, *Anapiculatisporites hystricosus* Playford, *Colatisporites decorus* (Bharadwaj & Venkatachala) Williams et al., *Verrucosporites morulatus* Potonié & Kremp emend. Smith & Butterworth and *Verrucosporites papulosus* Hacquebard. This association was ascribed to the early Visean and correlated with the middle part of the Malimán Formation (Sessarego and Césari, 1989; Amenábar and di Pasquo, 2009). On the other hand, a U–Pb zircon age of 348 ± 2 Ma obtained from granitic clasts from the conglomerates of the Upper Member of El Ratón Formation indicates a maximum Tournaisian age for the diamictite interval (Gallastegui Suárez et al., 2014).

In the western part of the Madre de Dios Basin (Fig. 1) the presence of Tournaisian-Visean diamictites could suggest glacial conditions during deposition. However, the diamictites were interpreted as gravity-flow deposits in delta-front environments (Díaz Martínez et al., 1993; Díaz Martínez and Isaacson, 1994), likely connected to glacial conditions inland.

Tournaisian glacial deposits were reported from the Longá Formation in the Parnaíba Basin (Lobato and Borghi 2007, 2014; Caputo and dos Santos, 2018, 2019), from the Oriximiná and upper Jaraquí Formations in the Solimões and Amazon basins.

2.3. Event 3 (Visean)

The existence of a Visean glaciation along the southwestern margin of Gondwana was largely neglected until the finding of glacial diamictites in the upper member of the Cortaderas Formation (Río Blanco Basin, Limarino et al., 1992; Limarino and Césari, 1993). This unit forms the upper unit of the Angualasto Group and rest over the previously cited Malimán Formation (Event 2, Fig. 3). The upper member of the Cortaderas Formation comprises laminated shales, diamictites, shales with dropstones, and interbeds of shales and fine-grained sandstones. Limarino et al. (1992) distinguished two different types of diamictites: i) reworked stratified diamictites interpreted as deposited by subaqueous gravity flows, and, ii) massive diamictites, some of which resemble iceberg dump structures like those described by Thomas and Connell (1985). Originally, Limarino et al. (1992) correlated these diamictites with those found in the Hoyada Verde Formation (Event 4, Late Serpukhovian-Bashkirian, Fig. 3) but the later finding of palynomorphs in the associated laminated shales and shales with dropstones allowed their reassignment to the Visean (Césari and Limarino, 1992; Limarino and Césari, 1993; Pérez Loinaze et al., 2010). The upper glacial deposits of the Cortaderas Formation yielded spores of late Visean age together with younger species such as *Apiculatisporis variornatus* di Pasquo et al., *Cristatisporites inordinatus* (Menéndez and Azcuy) Playford, *Indotriradites volkheimeri* (Azcuy) Pérez Loinaze, *Raistrickia paganziana* Azcuy,

Raistrickia rotunda Azcuy, *Retusotriletes anfractus* Menéndez and Azcuy and *Spinozonotriletes hirsutus* Azcuy. A late Visean age is assigned to this assemblage (Limarino and Césari, 1993; Pérez Loinaze et al., 2010) and referred to the *Reticulatisporites magnidictyus-Verrucosisorites quasi-gobbettii* (MQ) Interval Biozone (Fig. 4), defined for western Argentina (Pérez Loinaze, 2007) in partial replacement of the former *Cordylisorites-Verrucosisorites* Biozone (Césari and Gutiérrez, 2001).

In the northernmost part of the Argentinian Precordillera (Fig. 5) Fauqué and Limarino (1991) described thick beds of diamictites, pebbly shales and shales with dropstones up to 50 cm in diameter in the Jagüel Formation. Considering that these rocks are unconformably overlain by the Río del Peñón Formation (Serpukhovian-Late Permian) the glacial interval was referred to the early Carboniferous *sensu lato* (Fauqué and Limarino, 1991). Based on their similar stratigraphic positions, it is probable that the diamictite outcrops described by Baez et al. (2014) in the Punta del Agua Formation are time-equivalent deposits to those included in the Jagüel Formation. The age of the glacial deposits

included in the Punta del Agua Formation is constrained by the radiometric age of 335.99 ± 0.06 Ma (U–Pb zircon) from an andesitic level emplaced several meters above the diamictite interval close to the top of the unit, (Gulbranson et al., 2010). Therefore, the glacial interval described by Baez et al. (2014) is pre-Serpukhovian and very probably Visean.

The glacial event 3 is also recorded in the Calingasta-Uspallata Basin; Colombo et al. (2014) described in the El Planchón Formation a thick sequence of shales followed by diamictites with soft sediment deformation and shales with dropstones bearing faceted and striated clasts in the diamictites. Milana and di Pasquo (2019) also identified a glacial episode in the lower part of the El Planchón Formation. According to Milana and di Pasquo (2019), the age of the El Planchón Formation is estimated by the presence of palynological species that range from Tournaisian to Visean (e.g., *Anapiculatisporites hystricosus*, *Apiculir-etusispora semisenta*, *Cristatisporites aculeatus*, *Cristatisporites echinatus*, *Cristatisporites peruvianus*, *Crassispora scrupulosa*, *Densosporites*

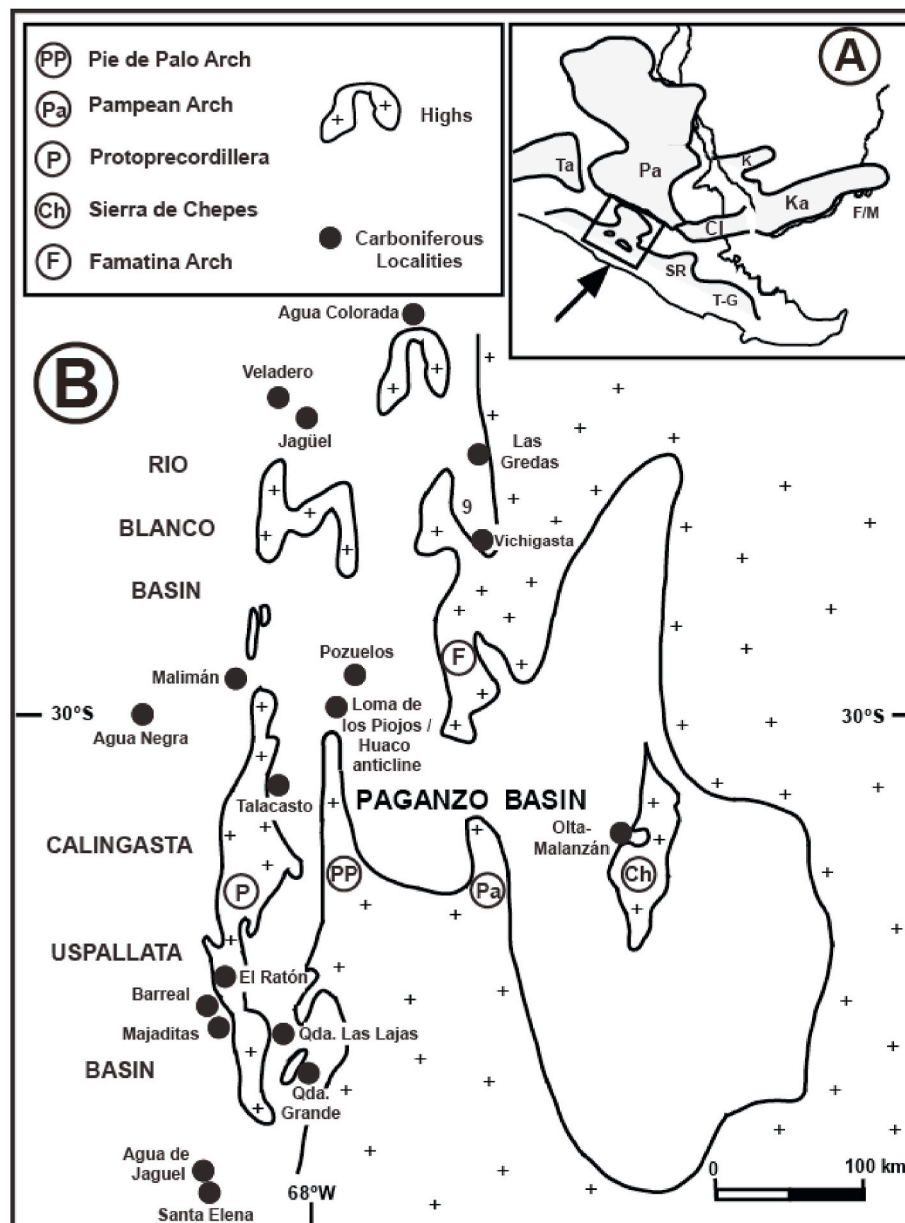


Fig. 5. A. Generalized late Paleozoic paleogeography of western Gondwana with main basins, C: Claromecó Basin, F/M: Falkland/Malvinas, K: Kalahari Basin, Ka: Karoo Basin, Pa: Paraná Basin, SR: San Rafael Basin, Ta: Tarija Basin, T-G: Tepuel-Genoa Basin. B. Paganzo, Calingasta-Uspallata and Río Blanco basins with localities mentioned in the text.

secundus, among others) together with species characteristic of the Visean (*Cristatisporites inconstans*, *Cristatisporites indignabundus*, *Cristatisporites indolatus*, *Cristatisporites scabiosus*, *Cristatisporites stellatus*, *Densosporites steinii*, *Dibolisporites microspicatus*, *Verrucosporites morulatus*). The U–Pb 348 ± 2 Ma age obtained from a clast in the underlying El Ratón Formation also supports a Visean age for these glacial deposits.

In the Paraná Basin, new palynological assemblages from the basal-most part (unit 1-B by Vergel et al., 2015) of the Itararé Group allowed Rosa et al. (2019) to suggest a middle Visean-early Serpukhovian age for the glacial interval. The palynofloras are characterized by spores usual in the MQ Biozone of western Argentina in the Paganzo and Río Blanco Basins and the Mag Biozone of northern Brazil in the Amazonas Basin (Melo and Playford, 2012).

2.4. Event 4 (late Serpukhovian – early Bashkirian)

This glacial event corresponds to the classic and best-known Carboniferous glacial deposits in the western basins of Argentina (González, 1981, 1990; Limarino and Gutiérrez, 1990; López-Gamundí et al., 1992; Taboada and Carrizo, 1992; López-Gamundí, 1997; Pazos, 2002; Marensi et al., 2005; Henry et al., 2008; Isbell et al., 2012). Unlike in the Paganzo basin, the absence of tuffaceous levels in the Late Paleozoic glacial and early postglacial successions of the Calingasta-Uspallata Basin prevents a precise chronostratigraphic framework for the glaciomarine deposits present in the basin. Instead, recent work has emphasized the study of faunal and floral content and the use of sequence stratigraphic analysis for correlation.

Glaciomarine facies are represented along the Calingasta-Uspallata (Hoyada Verde, El Paso, Majaditas, Leoncito, Yalguez and Agua de Jagüel Formations), Río Blanco (Río del Peñón) and San Rafael (lower Imperial Formation) basins. According to the paleogeographic position glacial deposits vary from distal glaciomarine to fjord-like accumulations exposing tillites, boulder striated pavements, faceted and striated clasts, and shales with dropstones (Mésigos, 1953; López-Gamundí, 1991, 1997; Taboada and Carrizo, 1992; Henry et al., 2010, 2014).

The age of the Hoyada Verde and El Leoncito Formations was first established by the presence of marine invertebrates belonging to the *Levipustula levis* fauna assigned to the late Carboniferous by Amos and Rolleri (1965). Initially, brachiopods of the *Levipustula* Biozone were found in postglacial shales that cover the diamictite horizons (González, 1981, 1990; Taboada, 1997) but later Cisterna and Sterren (2008, 2010) and López-Gamundí et al. (2016) reported a “post-glacial” and an “intra-glacial” *Levipustula* Fauna. A new palynological assemblage described from the glacial middle part and the early postglacial upper part of the El Paso Formation, contiguous to the Hoyada Verde exposures, and correlatable to the Subzone A of the *Raistrickia densa-Convolutispora muriornata* (DM) Biozone has been assigned to the late Serpukhovian–Bashkirian (Vergel et al., 2015). This correlation confirms previous interpretation of López-Gamundí and Martínez (2003) who, based on facies and sequence stratigraphic analysis, suggest that the Hoyada Verde and El Paso Formations, part of the San Eduardo Group of Mésigos (1953), are time equivalent units in agreement with the original stratigraphic scheme proposed by Amos and López-Gamundí (1981a). This is in sharp contrast with the proposal by González (1990) who, based on studies of faunal assemblages and facies, suggests that the El Paso Formation is older than the Hoyada Verde Formation (“The present author’s (C.R. González) interpretation differs from that of Amos and López-Gamundí (1981a) in that those formations are not equivalent in part, but they occur one above the other, as demonstrated by the sequence of faunal assemblages and lithofacies.”). Consequently, González (1990) defines “three different glacial episodes, presumably the oldest that can be clearly recognized in Gondwana, are present in the Barreal Hills of the southern Precordilleran area. They are the First El Paso, the Second El Paso and the Hoyada Verde glaciations. The San Eduardo Group is thus an outstanding glacial sequence that was deposited during latest Visean plus part, or perhaps all of the Namurian time”.

The succession of the Majaditas Formation (Amos and Rolleri, 1965; López-Gamundí et al., 1992; López-Gamundí, 2001), exposed a few kilometers south of the Hoyada Verde/El Paso outcrops, has provided palynological data from its middle part suggesting a correlation with the Subzone A of the DM Biozone (Vergel et al., 2016). Additionally, brachiopods found in lenses of siltstones with dropstones associated with diamictites have been assigned to the *Levipustula levis* Zone (González, 2002). Some bivalves and gastropods identified by Taboada (1997), González (2002), and Pinilla and Taboada (2018) were assigned to this zone. In addition to the *Levipustula* fauna, Cisterna et al. (2017) defined the *Aseptella-Tuberculatella/Rhipidomella-Micraphelia* (AT/RM) fauna which characterizes the postglacial record during the Serpukhovian–Bashkirian interval. Sterren et al. (2020) extended the record of the AT/RM fauna to the southernmost outcrops of the Calingasta-Uspallata Basin in the Agua de Jagüel Formation, establishing a Serpukhovian–Bashkirian age for the glacial retreat and postglacial flooding of the western Andean basins of Argentina.

The presence of an early Permian glacial episode for the Calingasta-Uspallata Basin, postulated originally by Keidel (1940), has been further reaffirmed primarily on the basis of findings of faunal assemblages characterized by bivalves found in the Agua del Jagüel Formation and the Santa Elena Formation (González, 1993; González and Díaz Saravia, 2007, 2010, 2020). This view acknowledges the presence in the basin of the widespread mid Carboniferous glacial episode and considers that “the ‘middle’ Carboniferous and the early Permian glacial deposits are separated by a thick non-glacial sequence of Late Pennsylvanian age” (González and Díaz Saravia, 2020). This school of thought further proposes the correlation between the Agua del Jagüel and the Santa Elena Formations exposed in the southern part of the basin and assigns an early Permian age to the glacial deposits of both units. This is in contrast to the opinion of authors (López-Gamundí, 1984; Henry et al., 2010) who assigned the glacial interval in these units to the Serpukhovian – early Bashkirian glaciation. The glacial section of the Santa Elena Formation is overlain by littoral deposits with plant remains of the *Nothorhacopteris argentinica* - *Botrychiopsis weissiana* - *Ginkgophyllum diazii* (NBG) flora and invertebrates of the late Pennsylvanian *Tivertonia jachalensis* – *Streptorhynchus inaequioratus* (T-S) Zone, both in the lower Member (= Jarillal Member), which in turn is overlain by wave-dominated and river-influenced deltaic deposits (Buatois et al., 2013) with invertebrates of the early Permian *Costatumulus amosi* Zone (Taboada, 1998; 1999).

In the Agua de Jagüel Formation the assignment to the early Permian of the glacial section is based on a misinterpretation of the local structure. An interval of sandstones and conglomerates (“sección inferior” Taboada, 1987) with invertebrates of the *Costatumulus amosi* (Taboada, 1987, 2006) and palynological content (Pérez Loinaze et al., 2010) both of early Permian age is in faulted contact with the overlying mid Carboniferous glacial section. This section was then incorrectly considered the basal interval of the Agua de Jagüel Formation and therefore the glacial section and the associated postglacial shales, in this interpretation, constituted evidence of an early Permian episode in the basin. An alternative structural interpretation (Limarino et al., 2013) re-confirms the mid Carboniferous age for the glacial section of the Agua de Jagüel as proposed by López-Gamundí (1984) and López-Gamundí et al. (1987), consistent with the presence of glacial deposits of such age in the rest of the basin. This age assignment is also corroborated by a mid Pennsylvanian (Moscovian–Kasimovian) K/Ar biotite age of 307.2 ± 5.2 Ma obtained from a dacitic lava found in the overlying middle section of the formation (Lech, 2002). More recent interpretations suggest that the upper part of the Agua de Jagüel Formation can be assigned to the late Carboniferous T-S Biozone (Taboada, 2010) rather than to the early Permian.

In transitional and continental areas (Paganzo, easternmost Río Blanco and San Rafael basins, Figs. 1 and 2) this glacial event was recorded in the Río del Peñón, Guandacol, Agua Colorada and El Imperial Formations (Limarino and Gutiérrez, 1990; Espejo, 1990; Espejo,

1996; Marenssi et al., 2005; Henry et al., 2014; Vázquez et al., 2019). U–Pb zircon radiometric ages from tuffs allow constraining the age of the glaciation in the Paganzo and Río Blanco Basins to the Serpukhovian - early Bashkirian. This is consistent with the 335.99 ± 0.06 Ma (late Visean) age obtained from an andesitic lava flow emplaced some meters below of the glacial horizon. The end of the glaciation can be estimated by the age of 319.57 ± 0.086 Ma (middle Bashkirian) reported by Gulbranson et al. (2010) from a tuff intercalation in the postglacial transgressive shales. Valdez Buso et al. (2017) and Valdez Buso et al. (2020) identified U–Pb ages of 320, 321 and 321 Ma in the Bashkirian transgressive deposits in the Guandacol and Jejenes Formations of the western Paganzo basin. These radiometric ages are consistent with palynological information obtained from the Guandacol Formation (Pérez Loinaze et al., 2011; Pérez Loinaze and Césari, 2012; Valdez Buso et al., 2020), Agua Colorada (Limarino and Gutiérrez, 1990; Césari et al., 2019) Quebrada Larga Formation (Limarino et al., 2014a), Tuminico Formation (Colombi et al., 2018), El Imperial Formation (Vázquez et al., 2019), El Paso Formation (Vergel et al., 2015) and La Deheza Formation (Balarino et al., 2015) that allowed to refer the palynofloras to the Subzone A of the *Raistrickia densa/Convolutispora muriornata* Biozone.

Glacial diamictites of event 4 are also exposed in the Tarija and in Chaco-Paraná basins Figs. 1 and 2). The 800 m thick Tarija Formation comprises diamictites, interpreted as glacial in origin, which are present at different stratigraphic levels (Frakes et al., 1969; Helwig, 1972; López-Gamundí, 1984, 1986b; del Papa and Martínez, 2001; Starck and del Papa, 2006; Bache et al., 2012; Rousse et al., 2018). Palynological data by di Pasquo et al. (2019) reported two glacial intervals, the older one in the Itacuami Formation and assigned to the mid Bashkirian (GI 3 by Anderson, 2011) and the upper one (lower part of the Tarija Formation) to the late Bashkirian (GI 4 of Anderson, 2011). Both glacial intervals in the Itacuami Formation and the lower part of the Tarija Formation (Intervals I and II of López Gamundí, 1986a, b) could correspond to event 4. The palynological associations identified by di Pasquo et al. (2017, 2019) may be older. The composition of the RS Biozone and the Subzone A of the DM Biozone (central western Argentina) is equivalent and constrained by radiometric dating to the late Serpukhovian-early Bashkirian (Césari and Pérez Loinaze, this volume).

Glacial deposits have been recorded in two areas of the Chaco-Paraná Basin. In the southern part of the basin diamictites have been described in lower part of the Ordoñez Formation (Fernández Garrasino, 1996; Winn and Steinmetz, 1998). Based on palynological content from the diamictite-bearing interval (lower Ordoñez), Gutiérrez and Balarino (2019) defined the *Anapiculatisporites concinnus* Assemblage Zone of late Serpukhovian–Bashkirian age. Other diamictite deposits of probable glacial origin were reported in the northern part of the of the Chaco-Paraná basin (Charata Formation) and in eastern Paraguay along the eastern margin of the Paraná basin in the Cabrera Formation and Coronel Oviedo Group but the precise age of these intervals in the Carboniferous is yet uncertain (Fúlfaro, 1996; Winn and Steinmetz, 1998; Limarino et al., 2014b).

In the intraplate region, Paraná Basin, this glacial event is identified in the Lagoa Azul Formation, the basal unit of the Itararé Group (Bigarella et al., 1967; dos Santos et al., 1996; Rocha-Campos et al., 2008; Vergel et al., 2015). Lagoa Azul, originally defined in the subsurface (França and Potter, 1991), is made up by different types of diamictites with striated and faceted clasts, shales with dropstones, shales, and fine-grained sandstones (Vesely and Assine, 2006; Rocha-Campos et al., 2008; Vergel et al., 2015). The base of the Itararé Group exhibit striated pavements and other features indicating glacial erosion in different areas of the basin (Rocha Campos et al., 2008; Rosa et al. 2016; Fallgatter and Paim, 2019). Palynological assemblages recovered from the Unit 3 of the Lagoa Azul Formation (Vergel et al., 2015) at approximately 90 m from the basal unconformity were analyzed by Kipper et al. (2017). The palynofloras are characterized as monosaccate pollen and spores that were referred to the *Ahrensispores cristatus* (Ac) Biozone,

considered of late Bashkirian-Kasimovian age. However, the species recorded in the sampled interval are also identified in the Subzone A of the DM Biozone in central western Argentina, constrained by radiometric dates to the late Serpukhovian-early Bashkirian (Césari et al., 2011).

2.5. Event 5 (late Pennsylvanian-earliest Permian)

This is the youngest late Paleozoic glacial event recorded in southern South American basins, the one that was recognized in the Paraná and Tarija Basins. In the Paraná Basin this glaciation is represented in the middle and upper part of the Itararé Group in the Campo Mourão and Taciba formations (França and Potter, 1991; Vesely and Assine, 2006; Mottin et al., 2018). Glacial evidence include tillites, reworked diamictites bearing faceted and striated clasts, shales with dropstones, striated pavements, boulder striated pavements and varve-like rhythmites (Bigarella et al., 1967; Rocha-Campos and dos Santos, 1981; Gravenor and Rocha-Campos, 1983; Trosdorf et al., 2003; Vesely and Assine, 2006; 2008; Franco and Hinnov, 2012; Aquino et al., 2016; Mottin et al., 2018; Vesely et al., 2018; Fallgatter and Paim, 2019; Fedorchuk et al., 2019). The age of this glacial event in the Paraná basin has been basically established by the study of palynological associations on shales at the top of deglaciation sequences (i.e., Campo Murão - Lontras Shale, Taciba - Passinho) and some radiometric ages. Palynological assemblages from the Lontras Shale and Passinho Shale have been referred to the *Vittatina costabilis* Biozone (Valdez Buso et al., 2020) formerly considered of early Permian age (Souza, 2006). The Lontras Shale is a basin-scale marker related to a maximum marine flooding of the Campo Murao-Lontras deglaciation sequence, located in the upper third of the Itararé Group, now considered by Valdez Buso et al. (2020) of latest Pennsylvanian age. An age of 307 ± 3.1 Ma was obtained by Cagliari et al. (2016) for the topmost glacial deposits of the Itararé Group (Taciba Formation) in the southern portion of the Paraná Basin. This was considered also by Valdez Buso et al. (2020) to constrain the Itararé Group to the Carboniferous (Late Pennsylvanian). Mollusks and brachiopods from shales in the Taciba Formation (Capivari Marine Bed) have been assigned to the *Eurydesma-Lyonia* fauna (Simões et al., 2019), which is further correlated with the *Eurydesma* fauna of Asselian-earliest Sakmarian age identified in the Bonete Formation in the VFB and the Hardap shale in the Kakahari basin (Neves et al., 2014; Simões et al., 2012, 2019).

Radiometric ages from ash and tonstein levels from the overlying Río Bonito Formation have helped to constrain the end of the glaciation in the Paraná basin (Guerra-Sommer et al. 2006, 2008a, 2008b; Rocha-Campos et al. 2006, 2019; Mori et al., 2012; Simas et al., 2012; Cagliari et al., 2014; Griffis et al., 2019). A tonstein intercalated in the lower Candiota coal seam provided an age of 304.0 ± 5.6 Ma while an ash level in the Faxinal-Leão upper coal seam give and age of 303.0 ± 5.7 Ma (Rocha-Campos et al., 2019). Other ages for the Río Bonito Formation were obtained by Matos et al. (2001), Guerra-Sommer et al. (2008a, 2008b) and Mori et al. (2012). Using higher precision CA-TIMS dating, Griffis et al. (2018) has returned dates for the oldest coals in the Río Bonito Formation in the 296–298 Ma range. Thus, the analysis of all these ages suggests that glacial conditions finished before the beginning of the Permian probably during the mid Gzhelian. However, palynology in Mottin et al. (2018) and a U–Pb age of Jurigan et al. (2019) place the top of the Itararé Group in Paraná State (Eastern Paraná Basin) in the early Permian (Asselian-Sakmarian), which suggest that ice feeding into the Paraná Basin came from different ice spreading centers, and that final deglaciation was not synchronous among the various glaciers, or glacial lobes.

In the Tarija Basin of southern Bolivia and northern Argentina (Fig. 1), glacial episodes 5 and 6 (Anderson, 2011) are recorded in the Yaguacua and Caiguamí members of the San Telmo Formation (Ahlfeld, 1946; Starck and del Papa 2006; di Pasquo et al., 2019; Gallo et al., this volume). The lower Yaguacua Member is composed of shales and

diamictites containing small polished, faceted, and striated clasts that upward pass to medium- and coarse-grained fluvial and eolian sandstones (Chimeo Member). The upper glacial event is represented in the Caiguamí member that exhibits massive, coarse- and fine-grained diamictites, shales with dropstones and cross-bedded sandstones. The San Telmo Formation was considered Kasimovian - earliest Asselian by [di Pasquo et al. \(2019\)](#) and referred to the MR and TB Biozones. The MR Biozone ([di Pasquo, 2002](#)) includes the key spore *Converrucosporites confluentis* considered Ghzelian in age ([Stephenson, 2008](#)).

This glacial event in the San Telmo Formation is particularly problematic since it is in partial conflict with a new radiometric date from an ash bed in the Copacabana Formation in Manuripi X-1 well ([Fig. 2](#)) in northern Bolivia, Madre de Dios basin ([Hamilton et al., 2016](#)). A massive, siliceous air-fall ash/tuffaceous interval surrounded above and below by carbonates yielded a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 316.0 ± 0.4 Ma (CA-ID-TIMS single grain analyses). The presence of red algae, crinoids and bryozoans in the limestones ([Carvajal et al., 2018](#)) suggests deposition at mid- and high-latitudes. This Bashkirian age confirms that the Copacabana carbonates formed contemporaneous with glacial deposits in the nearby Tarija basin. U-Pb [ID-TIMS] radiometric ages from zircon-bearing interbedded tuffs in the Copacabana Formation (central Bolivia) have a range 298 - 291.3 Ma ([di Pasquo et al., 2014](#)). A younger age (Ghzelian-earliest Asselian) assigned to the San Telmo Formation in Tarija basin allows a correlation with the younger part of the Copacabana Formation, considered to have been deposited in subtropical to tropical oceans (cf. Gallo et al., this volume).

New dates on the Tunas Formation exposed along the VFB of central Argentina ([Fig. 2](#)) help capping the minimum age of the glacial deposits (grouped under the Sauce Grande Formation) and the transition to postglacial marine mudstones (Piedra Azul Formation). Recent palynological studies of the middle member of the Sauce Grande Formation indicate a rather broad age range from late Pennsylvanian to Cisuralian ([di Pasquo et al., 2008](#)). The Piedra Azul Formation consists of mudstones and sparse cross-bedded, fine-grained sandstones and grades upward to the Bonete Formation. The latter is characterized by bioturbated shales and rippled and cross-bedded sandstones with marine fauna (*Eurydesma* fauna, [Harrington, 1955](#)) and flora (*Glossopteris* flora, [Archangelsky and Cúneo, 1984](#)) indicative of an early Permian (Asselian-Sakmarian) age. Tuffaceous levels found in the *Eurydesma*-bearing beds found in the Hardship Shale in the Kalahari Basin has provided a U-Pb SHRIMP age of 297 ± 1.8 Ma ([Werner, 2006](#)). Recently this age has been confirmed by a new more precise age of $296.41\text{--}0.27\text{--}0.35$ Ma ([Griffis et al., 2019](#)). This date indirectly provides additional evidence of the late Asselian age of the Bonete postglacial transgression. More recently, a palynological assemblage of Cisuralian-Guadalupian age has been recently described from an interval assigned to the Piedra Azul and Bonete Formations from two offshore wells ([Balarino, 2012, 2014](#)).

SHRIMP radiogenic isotope dating on zircons in vitric tuffs (280.8 ± 1.9 Ma) indicates an early Permian (Artinskian) age for the uppermost section of the Tunas Formation exposed in the VBF ([López-Gamundí et al., 2013](#)). This age is complemented with the SHRIMP zircon age of 295.7 ± 8.0 Ma obtained at the base of the Tunas Formation by [Arzadún et al. \(2018\)](#). An association of mudstones and coal interbeds with thin tuff levels was identified in the Tunas section of the Pangea 1 and Pangea 3 wells in the Claromecó basin ([Fig. 2, Arzadún et al., 2016](#)). An age of 291.7 ± 2.9 Ma was obtained from these tuffaceous levels in the Pangea 1 well ([Arzadún et al., 2018](#)). In summary, radiometric ages show the Tunas Formation was deposited approximately between 295 and 280 Ma, suggesting a pre-Sakmarian age for the Sauce Grande glacial deposits. Furthermore, dating of the *Eurydesma* fauna as late Asselian constrains the glacial-postglacial transition in this basin to the latest Pennsylvanian-Asselian.

3. Facies models

The unbalanced preservation potential of glacial deposits, skewed

toward the glacial marine sediments, is well established for all the pre-Pleistocene intervals affected by extremely cold climate. Preservation potential of any succession of sediments depends on a delicate balance of long- and short-term components. The long-term component is largely controlled by the geotectonic setting (i.e., basin type) and the related tectonic activity of the basin after deposition (i.e., postdepositional erosion); the short-term component is connected to the erosion and/or reworking coeval with deposition ([Nystuen, 1985](#)).

For subaerial glacial environments the short-term component of the preservation potential is significantly low due to potential erosion and/or reworking by fluvial and aeolian processes. This low preservation may be augmented by the effects of glacio-isostatic rebound.

Ice advances tend to remove the record of the previous depositional phases ([Hambrey, 1994](#)). Consequently, the preservation potential of marine glacial retreat deposits is greater than for the glacial advance deposits, making the former most likely to be preserved in submarine proglacial environments where they are rapidly covered by subsequent transgressive deposits during rapid sealevel rise or retreat of the ice front out of the marine basin.

In this chapter we highlight recent advances in the understanding of facies models of glaciated basins on three specific areas: i) continental glacial facies ii) glacial marine and transitional facies associations, and iii) paleoenvironmental reconstructions based on surface-subsurface regional analysis.

3.1. Continental glacial facies

Glacial accumulations entirely deposited in continental settings are scarce in the basins along the southwestern margin of Gondwana, probably due to the low preservation potential resulting from the postglacial fluvial reworking of glacial sediments. In the Paganzo Basin, two examples of glacial continental deposits were described in the lower part of the Agua Colorada and Guandacol Formations.

In the Las Gredas locality ([Fig. 5](#)), [Limarino and Gutierrez \(1990\)](#) identified in the Agua Colorada Formation continental glacial deposits represented by two distinct types of diamictites. On the one hand, those interpreted as tillites, described as chaotic, massive, coarse-grained, matrix-supported conglomerates. The matrix/clast (M/C) ratio is uniformly higher than 2. Clasts, up to 250 cm in diameter, occur embedded in a muddy matrix; they show a characteristic pentagonal profile with frequent faceted and striated surfaces. Commonly the striae are aligned parallel to subparallel to the longer axis of the clasts, though more than one pattern of orientation was frequently noted. The second type of diamictite, interpreted as reworked deposits, corresponds to matrix-supported paraconglomerates in tabular beds. Beds show normal grading, crude parallel stratification, and rarely clast imbrication. In this kind of diamictite, the M/C ratio decreases sharply, and in some cases, the fabric tends to be clast-supported.

The diamictites of Las Gredas succession are part of a basal terrestrial glacial facies. It is associated with lacustrine and fluvio-glacial facies. The glacially-influenced, diamictite-dominated facies form a laterally irregular horizon at the base of the Carboniferous succession. It comprises thick (up to 2 m thick) beds of variable thickness of massive coarse-grained diamictites, massive pebbly mudstones, tabular beds of reworked diamictites and ribbon bodies of cross-bedded sandstone and fine-grained conglomerates. The sandstones, dominantly medium- and coarse-grained, form up to 40 cm thick lenticular beds with convolute lamination and occur intercalated within the massive diamictites. The massive diamictites are interpreted as tillites. The stratified diamictites likely resulted from a brief fluvial reworking of previously deposited glacial sediments, or discrete episodes of gravity flows favored by unstable slopes off moraines. The lenticular beds of cross-bedded sandstones encased into the massive diamictites most probably represent the fill of eskers believed to form when sediment carried by glacial meltwater was deposited in subglacial tunnels.

The lacustrine facies association consists of shales, mudstones, fine-

and medium-grained, horizontal-laminated and ripple cross-laminated sandstones with intercalations of thinly bedded diamictites and thin (up to 10 cm thick) tabular beds of pebbly mudstones and crudely stratified diamictites. A remarkable feature of the shales is the existence of dropstones up to 13 cm in diameter and rhythmites interpreted as resulting from the alternate pulsations of underflow and overflow currents into the lake. Mudstones and fine-grained sandstones from the lacustrine deposits yielded palynological assemblages composed of monosaccate pollen grains and spores of the Subzone A of the DM Biozone (Césari and Gutiérrez, 2001). Neither acritarchs nor scolecodonts, frequently found in open marine and fjord environments, were recovered in these lacustrine deposits.

The third facies association comprises cross-bedded coarse- and medium-grained sandstones and clast-supported conglomerates in lenticular beds. This sandstone-conglomerate assemblage, interpreted as glaciofluvial deposits, rest erosionally on the glacially-influenced and lacustrine deposits.

Another example of continental glacial deposits in the Paganzo basin is described by Valdez Buso et al. (this volume) in the Guandacol Formation at the Vichigasta locality (Fig. 5). As in various localities in the Paganzo basin, palynomorphs of Bashkirian age have been identified in the flooding event of the Vichigasta paleovalley (Stage 3, Valdez Buso et al., this volume).

3.2. Glacial marine and transitional facies associations

Two main facies associations have been described for marine glacial retreat deposits: a Valley-Glacier-Retreat Facies Association developed along basin margins under predominantly shallow subaqueous and intermittent subaerial conditions and a Submarine-Retreat Facies Association in open-marine areas.

We herein provide an updated view of (i) a specific, and widespread, Valley-Glacier-Retreat Facies Association (fjord association) characterized by confined, low initial subsidence along the paleovalley margins with irregular topography and (ii) Submarine-Retreat Facies Association (glaciomarine apron association) developed in areas with increased accommodation space. These two facies associations can be considered as the end members of a spectrum controlled by accommodation space (high in subsiding, open marine shelf areas or moderate to low along indented coastlines with valleys/fjords) and the spacing of points of sediment injection (linear or closely spaced source points or isolated source points confined to widely spaced valleys or fjords).

3.3. Valley-glacier-retreat facies association (fjord association)

Late Carboniferous (Pennsylvanian) deposits exposed in the western part of the Paganzo basin provide an excellent opportunity for the characterization of paleovalley (fjord) facies associations (Limarino et al., 2002, 2006) or valley-glacier-retreat facies association (López-Gamundí, 1997, 2010). Fjords are deep, glacially-carved valleys with steep sides that have been invaded by the sea; they can reach several kilometers in length and are relatively narrow in width (<10 km). Buatois and Mangano (1992) suggested on the basis of trace fossil studies that the sedimentation in the Paganzo basin could have occurred in a fjord like environment. Later field descriptions of a paleovalley at Talacasto (Fig. 5) and its sedimentary fill (Martínez and Astini, 1992; Martínez, 1998; Aquino et al., 2014) suggested the presence of a fjord-like facies assemblage associated with the Carboniferous strata in the western Paganzo basin. Martínez (1998) characterized deposition in this part of the basin during glacial and early postglacial phases as “a brackish-water environment in the paleovalley due to mixing with freshwater near the glacial ice-front”. Subsequent studies provide further details on this facies association and expanded its distribution further south and north along the western Paganzo basin (Dijkstra et al. 2006a, 2006b; Kneller et al., 2004; Limarino et al., 2006; Aquino et al., 2014; Alonso-Muruaga et al., 2018).

Five main localities are herein described as outstanding examples of such association; those outcrops are: 1) Agua Colorada, 2) Veladero, 3) Huaco area 4) Talacasto and 5) Quebrada Grande-Quebrada de Las Lajas (Fig. 5).

3.4. Agua Colorada

The Carboniferous sequence (lower section of the Paganzo Group) exposed in the central part of the Sierra de Narváz (Agua Colorada locality, Fig. 5) exhibits a complex assemblage of diamictites, conglomerates, sandstones, carbonaceous mudstones and thin coal beds (Limarino et al., 2010). The 140 m thick section overlies granites of the Ordovician Suri Formation. Diamictites can be differentiated into two main types: monomictic and polymictic diamictites. The former is confined to the basal interval of the section and is characterized by massive, matrix-supported diamictites with striated clasts, indicating direct glacial deposition (tillites), and only locally modified by gravity flow resedimentation and fluvial reworking. Stratigraphically on top of the monomictic diamictite association lies the second diamictite type and associated facies. This interval consists of massive, matrix-supported diamictites commonly interbedded with conglomerates, sandstones and scarce mudstones, interpreted as debris flow deposits. The monomictic diamictite association and overlying polymictic association present in Agua Colorada indicate glacially influenced sedimentation and subsequent glacier retreat respectively. The section continues with mudstones and sandstones arranged in coarsening and thickening upward cycles deposited during a postglacial marine transgression that flooded the fjord. The uppermost part the succession comprises sandstones interbedded with mudstones, and subordinate conglomerate deposited during postglacial conditions in the fjord system.

3.5. Veladero

In this locality, the lower part of the Carboniferous (Pennsylvanian) Quebrada Larga Formation can be differentiated into three main stratigraphic intervals, all of them confined to a paleovalley carved into the granitic and high-grade metamorphic basement (Limarino et al., 2014a). The lower interval has a thickness that varies from 1 to about 40 m depending on the strata's position relative to the paleovalley axis and is composed of different diamictite types, mainly including massive and stratified diamictite, interpreted to record reworking of previously deposited poorly sorted glacial sediments. The middle interval comprises shale with dropstones, locally interbedded with lenticular beds of diamictite, and large-scale cross-bedded sets of sandstone and conglomerate in Gilbert-type deltas, recording a relatively more advanced stage of deglaciation followed by a marine transgression and formation of the fjord setting, and a later progradation of deltas within the fjord. The upper interval, is characterized by cross-bedded sandstone and conglomerate deposited in braided fluvial plains.

3.6. Huaco area (Loma de los Piojos - Huaco anticline - Pozuelos)

Several outcrops in the Huaco area (Fig. 5) provide examples of paleovalleys with glacial infill. In Loma de Los Piojos (Fig. 6) glacial and postglacial deposits of the Guandacol Formation occur within a paleovalley with steep side walls and a relatively flat floor (Alonso-Muruaga et al., 2018). The paleovalley is carved on Lower Devonian sandstones (Talacasto Formation) and Mississippian mudstones and sandstones (Loma de los Piojos Formation). Stratigraphic relationships and palynological assemblages recovered in mudstones contained in the paleovalley fill suggest an early Pennsylvanian age for both the incision and the paleovalley fill. A bedrock striated pavement, originally outlined by Keidel (1921), was developed on the pre-Carboniferous basement (López-Gamundí and Martínez, 2000; Alonso-Muruaga et al., 2018). It shows fluted-like features reminiscent of “roche-moutonnée” forms with

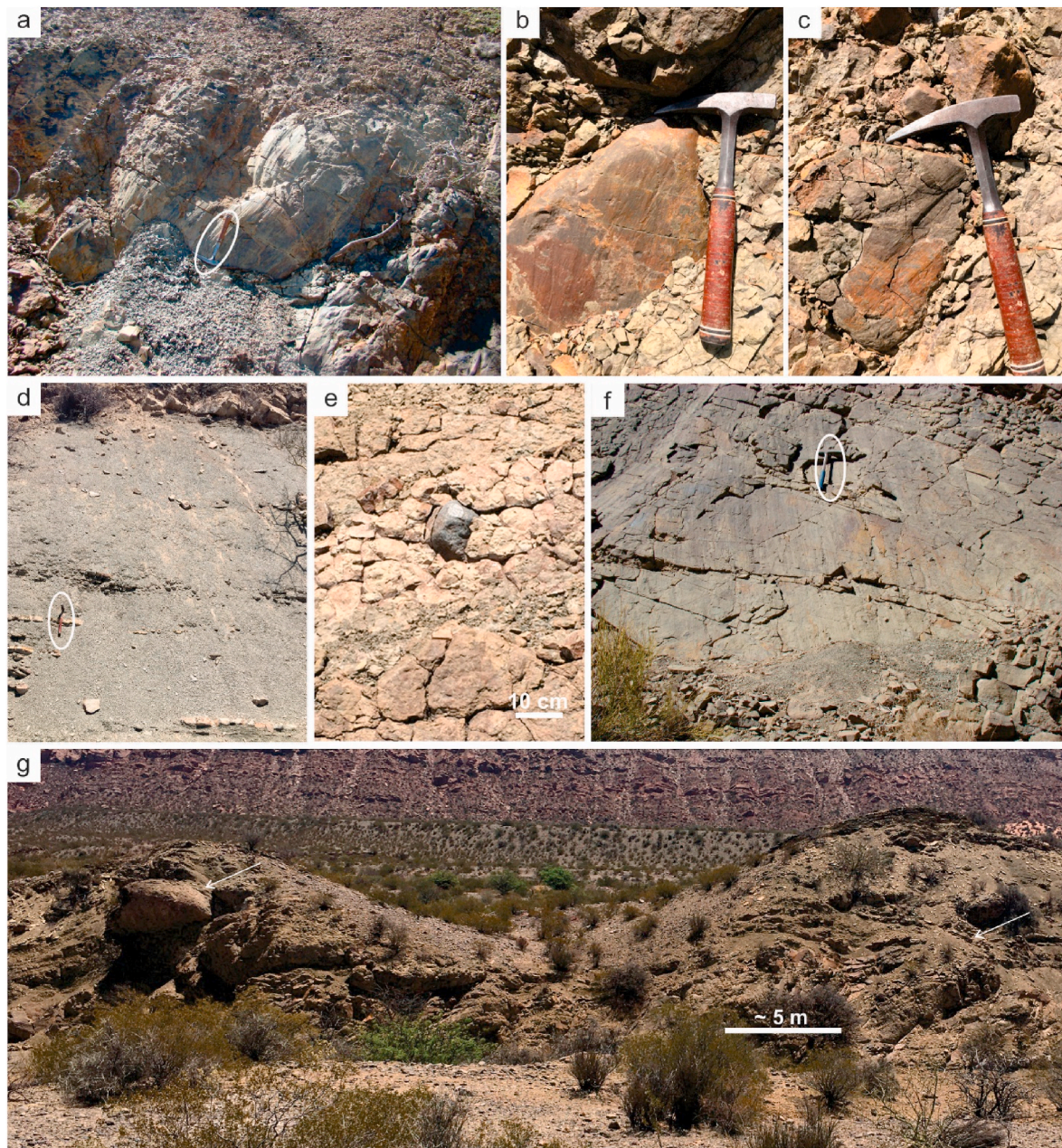


Fig. 6. Valley-Glacier-Retreat Facies Association (fjord association). Huaco area at the Loma de los Piojos locality. (a) Striated pavement located on the contact between Guandacol Formation and Talacasto Formation. (b) and (c) matrix-supported, massive diamictites and intraformational boulder pavement. Note striations on the polished surface of boulders of the pavement oriented along the long axis of the clast. (d) Laminated mudstones with dropstones, locally interbedded with sandstones. (e) Isolated “outsized” clast (dropstone). View from the top of the bed. (f) Intraformational striated surface on sand-rich, matrix-supported diamictites. (g) Slumped and deformed diamictites; arrow on the left points a ball and pillow structures, arrow on the right indicates top of a meter scale, recumbent fold.

well defined ridges, nailhead striations and crescentic marks and gouges (Martínez, 1998; López-Gamundí and Martínez, 2000). A thin (less than 2 m) lensoid or mound-shaped massive diamictite covers the pavement. Clasts have facets and striations and others are bullet-shaped indicative of subglacial abrasion and deposition by lodgment processes (López-Gamundí and Martínez, 2000). The basal diamictite is overlain by, and grades laterally into, laminated mudstones with outsize clasts and thin interbeds of medium- to fine-grained sandstones. Intraformational (soft bed) striated pavements associated with massive and stratified diamictites also occur (Alonso-Muruaga et al., 2011; Alonso-Muruaga et al., 2018). Stratified diamictites with high matrix/(larger)clast (M/C) ratio, interpreted as subaqueous debris flow

deposits, are interbedded with laminated mudstones with dropstones. They are characterized by meter-scale folding, possibly triggered by subaqueous slumping. Clast-supported conglomerates and sandstones, locally showing intense soft-sediment deformation, rest sharply on the stratified diamictite facies.

The glacial and glacially-influenced facies described above represent a wide variety of environments from subglacial setting to proglacial sedimentation in glaciallacustrine and glaciofluvial environments that were locally affected by mass transport processes, possibly ice push and/or overloading due to minor glacier readvance within the valley.

This glacial facies assemblage passes upward into at least two vertically stacked coarsening- and thickening-upward cycles consisting

of laminated mudstones and sandstones, without any glacial imprint, which are interpreted as prograding deltaic systems following cession of glacial activity.

Twenty-eight km to the northeast of the Loma de Los Piojos locality, a similar succession has been described at Huaco anticline (Fig. 7). Glacial deposits of the Guandacol Formation are well exposed on both flanks of the Huaco anticline (López-Gamundí and Martínez, 2000; Limarino et al., 2002; Pazos, 2002; Marenssi et al., 2005; Alonso-Muruaga, 2015). The exposed Guandacol interval, about 170 m thick, consists of diamictites, mudstones and sandstones. It rests on Ordovician limestones of the San Juan Formation. A pavement is carved on Ordovician limestones. Parallel to subparallel glacial striations were recognized on the pavement surface (López-Gamundí and Martínez, 2000). The basal part of the section at Agua Hedionda comprises at least three diamictite types: massive or structureless, crudely stratified and thinly-bedded. The massive diamictite facies occurs in lensoidal beds up to 1.5 m thick, commonly overlying the striated pavement carved on Ordovician basement, and includes abundant limestone clasts, and subordinate number of granite, quartz, and high-grade metamorphic clasts. Some clasts are striated, faceted, and bullet-shaped. Locally, the diamictites are associated with intraformational striated surfaces (Alonso-Muruaga et al., 2011). The massive diamictites are interpreted as lodgment tills. Cross-bedded, coarse-grained sandstones locally occur interbedded with the diamictites, or filling paleotopographic depressions, suggesting the presence of outwash processes. The stratified and thinly-bedded diamictites tend to overlie the massive diamictite beds. Both stratified and thinly-bedded types show high M/C ratio, form tabular beds. Individual bed thickness range from some centimeters for the thinly-bedded type to few decimeters for the stratified type. The thinly-bedded diamictites often present outsized clasts disrupting the stratification. Trace fossils occur at the bed tops or bottoms. These facies mainly record sedimentation from thin, subaqueous cohesive debris flows (cf. López-Gamundí, 1991) with ice-rafting and rain-out processes during glacial retreat.

The thinly-bedded stratified diamictites pass upward into mudstones with dropstones interbedded with millimetric to centimetric beds of very fine-grained sandstones with marine inarticulated brachiopods. These deposits represent suspension fall-out and ice-rafted rain-out sedimentation with episodic storm-generated currents (possibly underflows) during a transgressive event that flooded the fjord. This transgressive event peaked with the deposition of a black shale interval, which, in turn, is followed by deltaic coarsening- and thickening-upward mudstone-sandstone cycles.

The northernmost outcrops of glacial deposits in the Huaco area occur at the Los Pozuelos creek (Fig. 5), where a good example of the fjord sedimentation is exposed (Marenssi et al., 2005, Fig. 8 a). The basement, Ordovician limestones, shows polished (Fig. 8 b) and striated (Fig. 8 c) surfaces on which rest massive (Fig. 8 d) and crudely stratified diamictites. The basal diamictite-bearing interval reaches a thickness of up to 26 m and consists of stacked massive or crudely stratified beds (up to 2 m thick) of matrix-supported diamictites (Fig. 8 e). The glacial succession is capped by a transgressive interval comprised of dropstone-bearing shales (Fig. 8 f), wave-rippled sandstones (Fig. 8 g), and subordinate micro-hummocky cross-stratified sandstones.

3.7. Talacasto

This paleovalley (Figs. 5 and 9) was carved in Devonian metasediments (Punta Negra Formation). Injection structures such as conglomerate clastic dykes fill fractures in the basement (Martínez and Astini, 1992). Evidence of a glacial origin for the Talacasto paleovalley is provided by a U-shape cross section and a flat floor, with steep (>45°) to very steep locally overhanging valley walls (Martínez and Astini, 1992; Martínez, 1998; Aquino et al., 2014) and glacial striae on the valley walls (Aquino et al., 2014). Along the paleovalley walls the angular relationship between the paleovalley strata and basement varies from dominantly subvertical to vertical with onlaps on the basement margins at variable angles from 15° to 35° (Fig. 8 a and b; Aquino et al., 2014).

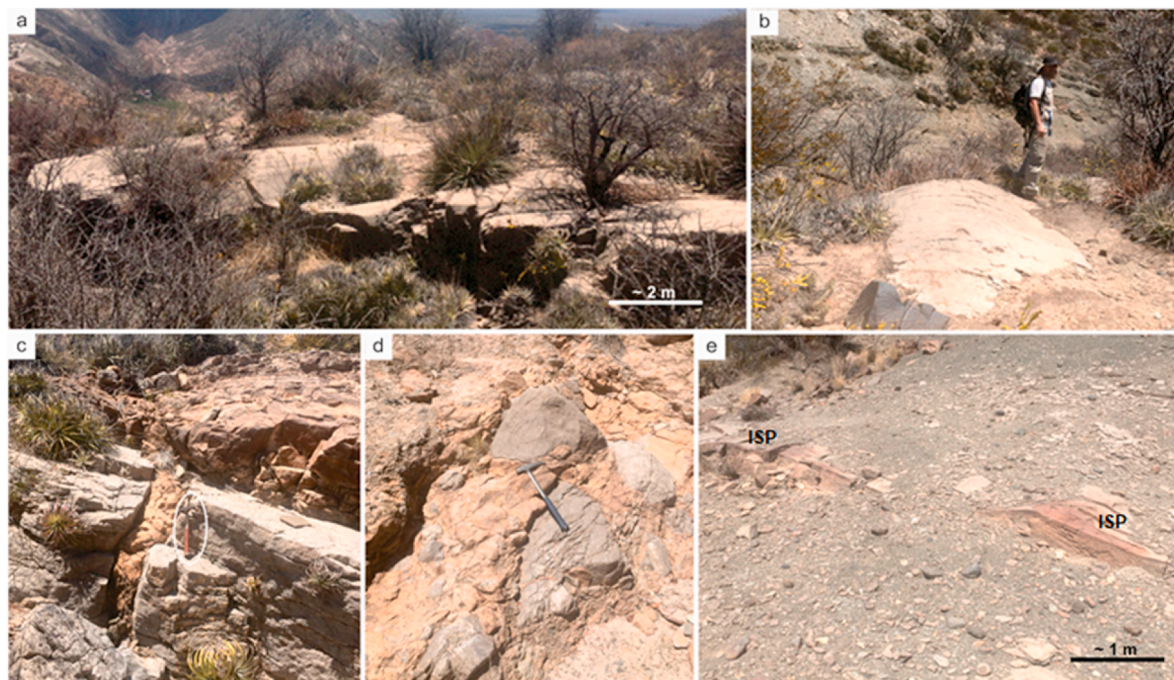


Fig. 7. Valley-Glacier-Retreat Facies Association (fjord association). Huaco area at Huaco anticline. a) General view of polished and striated pavement carved in Ordovician limestones (San Juan Formation) b) Detail of the striated pavement in (a), showing a “whale back” - like relief. c) Injected diamictite into the limestone basement. d) Diamictites with limestone clasts. e) intraformational striated surface/pavement (ISP) developed on light red sandstone bed top; the striated surface shows gentle undulations of meter scale wavelength. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

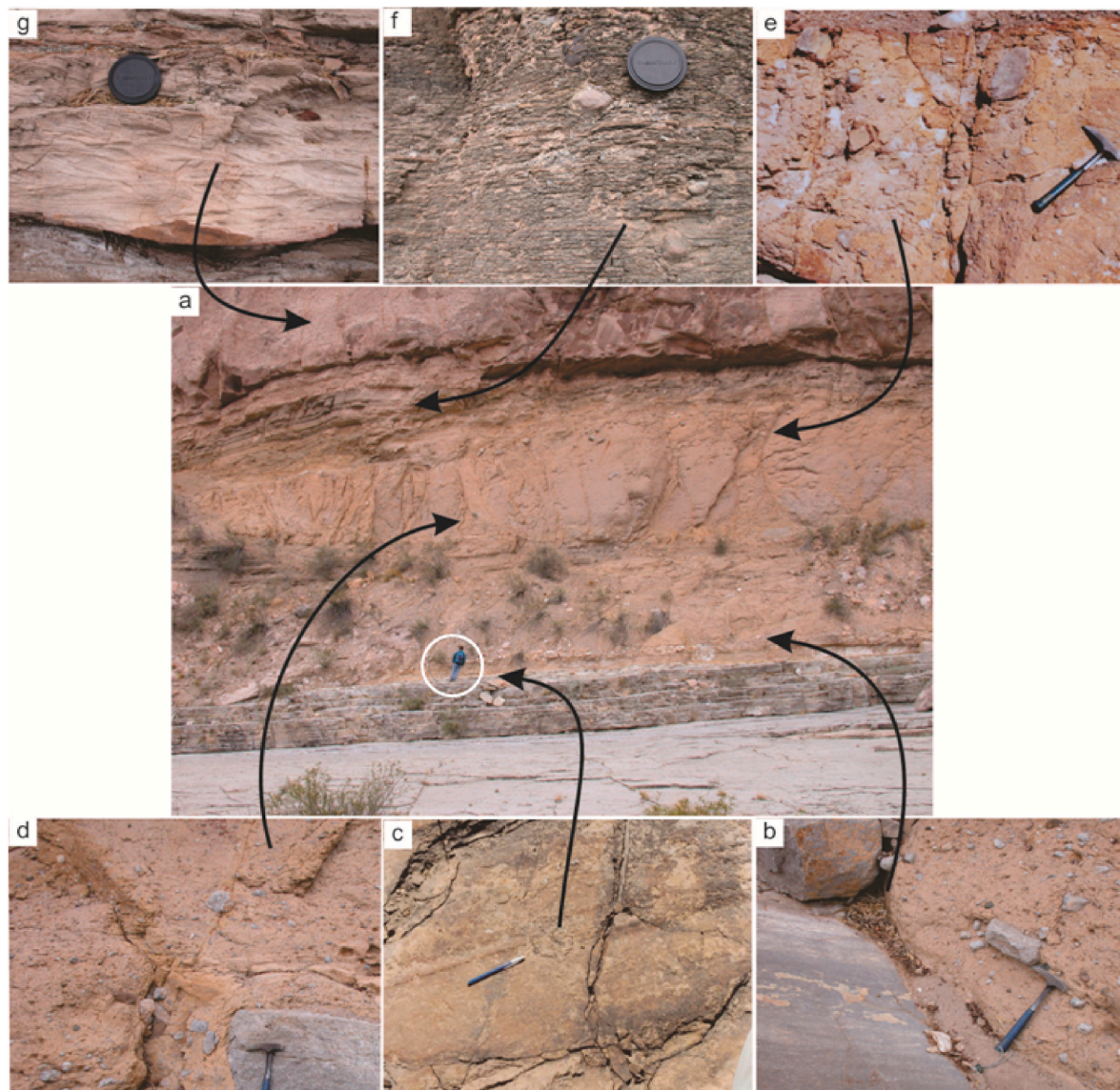


Fig. 8. Valley-Glacier-Retreat Facies Association (fjord association) in the Huaco area at the Pozuelos creek locality. a: General view of the Ordovician limestone basement and the overlying glacial deposits, b: contact between the Ordovician limestones and the massive diamictites interpreted as tillites, note the polished surface of limestone with small striations, c: glacial striations carved into the Ordovician limestones, d: detail of the massive diamictites found in the middle section of the outcrops, note dominant presence of limestones in the clast population and the wide range of clast sizes, e: poorly-stratified breccias found near the top of the glacial diamictites, f: shales with dropstones overlying the diamictite interval; these fine-grained rocks mark the flooding of the fjord system during glacial retreat, g: Medium- and fine-grained sandstones with wave-ripple cross-lamination; note the sharp contact with the underlying interval of shales with dropstones.

The exposed succession, approximately 170 m thick, exhibits a general fining and thinning upward trend with channelized conglomerates and sandstones at the base of the paleovalley. The basal coarse (maximum clast size (MCS): 55 cm) conglomerates are massive, inversely graded or with inverse to normal grading with few striated clasts and are interpreted as non-cohesive debris-flow deposits. They are associated with crudely stratified conglomerates (MCS: 18 cm) and pebbly sandstones, often with trough or planar cross bedding (Martínez and Astini, 1992). The conglomerate-dominated basal interval is interpreted as a glacially-influenced subaqueous fan delta. It grades up to a sand-prone interval (uppermost 60 m) dominated by coarse-grained sandstones with dispersed pebbles and cobbles and medium- to fine-grained sandstones with flute clasts, normal grading and climbing ripple lamination on the top of each bed, interpreted as deposited by low density turbidity currents. Lonestones up to 12 cm in diameter that pierce and deform underlying strata also occur. The upper third of the Talacasto succession includes sandstone-mudstone heterolithic strata interpreted as low

density Tbc and Te Bouma turbidites capped by micritic mudstones (Martínez and Astini, 1992), which illustrates the marine flooding following the early glacially-influenced sedimentation dominant during the early stage of the paleovalley fill. Paleocurrents from cross stratification and ripples shows a clear southward component along the paleovalley axis (Aquino et al., 2014).

The same paleovalley fill has been subdivided in seven units in a detailed study by Aquino et al. (2014). The basal 75 m - thick Unit 1 is equivalent to the conglomerate-dominated basal interval of Martínez and Astini (1992). Clinoforms in fine-grained conglomerates and coarse-grained sandstones of Unit 1 indicate flow toward the south. The basal part of Unit 1 (packages 1 and 2) is interpreted by Aquino et al. (2014) as an ice-contact delta (sensu Lønne, 1995), equivalent to a Gilbert delta in front of a glacial margin.

Units 1 through 5 show a single cycle of deglaciation, and most probably represent the succession logged by Martínez and Astini (1992). Outsized clasts (dropstones) have been observed at the top of the

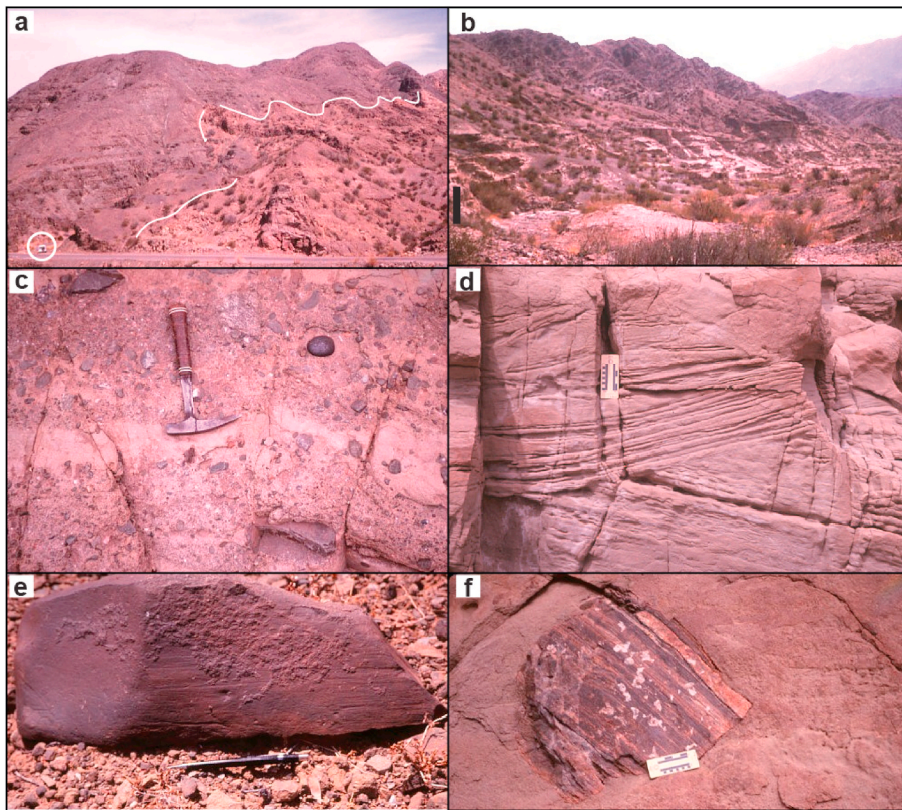


Fig. 9. Valley-Glacier-Retreat Facies Association (fjord association). Talacasto locality, a: steep contact between Devonian sedimentary basement (background) and paleovalley fill, car as scale on lower left corner, b: onlap of paleovalley strata (light colored) onto Devonian metasediments (dark colored, background), c: pebbly sandstone and pebble to cobble conglomerate, d: medium-to coarse-grained sandstones with unidirectional planar to trough cross bedding, e: faceted and striated boulder, f: outsized boulder in coarse-grained crudely stratified sandstone.

fining-upwards cycles in Unit 1. Unit 6 (a chaotic, massive and normally graded, moderately sorted conglomerates), interpreted as moraine complex systems, marks the beginning of a new cycle of glacier advance (Aquino et al., 2014). This conglomerate interval, fed laterally from the valley wall, is followed by deformed heterolithic facies of diamictites and clast-supported conglomerates (Unit 7) assigned to a mass transport complex. All units show evidence of glacial influence such as dropstones and striated clasts, which indicates that the glaciers were always present in the valley or in adjacent areas during sedimentation.

3.8. Quebrada Grande - Quebrada de Las Lajas

The southernmost succession that records deglaciation in a paleofjord in western Paganzo basin has been described in the Quebrada Grande (Fig. 5) by Bercowski et al. (1991) and Kneller et al. (2004). The geometry of the trunk valley and its tributaries is controlled by the underlying basement lithologies: a deep valley, with steep sides (with gradients in excess of 40°) exists where the bedrock is massive limestones, and a broader, shallower valley, exists where the bedrock is generally composed of fine-grained siliciclastic deposits (Kneller et al., 2004). The valleys are locally floored with diamictites, including both *in situ* tillites and remobilized diamictites. The major part of the valley fill consists of a ≤150-m-thick dropstone-bearing, mudstone-dominated succession. Dropstones decrease in abundance down the fjord. The entire succession is overlain by > 300 m of sandy turbidites, of which the upper part includes large mass-transport complexes. Debris flows, locally with meter-scale blocks of Ordovician limestone from the valley side, were triggered by rockfalls from the steeper valley sides.

Five stages of the paleovalley infill have been recognized by Kneller et al. (2004):

Stage I consists of massive diamictites with large boulders up to 3 m in diameter, interpreted as lodgment tills, and crudely stratified matrix-supported conglomerates interbedded with the resedimented diamictites, melt-out tills and/or cohesive debris flows, and current-worked

fine-grained sandstones. These deposits were probably laid down subglacially and indicate a glacially occupied valley during this stage.

Stage II is characterized by shallow marine proglacial, axial fan delta clinoforms up to 10 m high with topsets of very coarse-grained sandstones. The clinoforms grade laterally to siltstones rhythmities with dropstones and thinly laminated and cross-laminated sands interpreted as deposits of suspension underflows. The Stage II deposits overlie basement or stage I diamictites.

Stage III is characterized as proglacial, gravelly subaqueous outwash fans with matrix-supported conglomerates sufficiently widespread to be used as stratigraphic markers interpreted as debris flows. They are associated with winnowed layers and turbidites with dropstones and meltwater plume deposits. The sea level rise was sufficiently rapid during stage III to overwhelm the sediment supply.

Stage IV is an approximately 100-m-thick package of ice-distal, high-density and low-density sandy turbidites free of dropstones. The tabular nature of individual beds suggests deposition on a subhorizontal fjord floor.

Stage V comprises a thick (at least 300 m) interval of mass flow complexes made up of i) folded and disrupted turbidites, (ii) bouldery or pebbly sandy mudstones and muddy sandstones with blocks of limestone several meters across, and (iii) clast- to matrix-supported conglomerates with abundant limestone boulders, ranging from chaotic and massive to moderately sorted. This stage represents a phase of catastrophic sedimentation due to destabilization of slopes.

Twenty km north of the Quebrada Grande paleofjord at Quebrada de Las Lajas (Fig. 5) another well preserved glacial paleovalley with slump/debris flow deposits, locally containing meter-scale blocks of Ordovician limestone from the valley side (Fig. 10 a, b) is exposed. The Quebrada de Las Lajas preserves up to a 350-m-thick valley-fill succession in a highly confined fjord setting (Dykstra et al., 2006a, 2006b). The cross section of the paleofjord is U-shaped with steeply dipping walls (50–70°) (Dykstra, 2005). The paleofjord was approximately 200 m wide at its head, and 1 km wide near its mouth (Dykstra et al., 2006b). The fjord facies

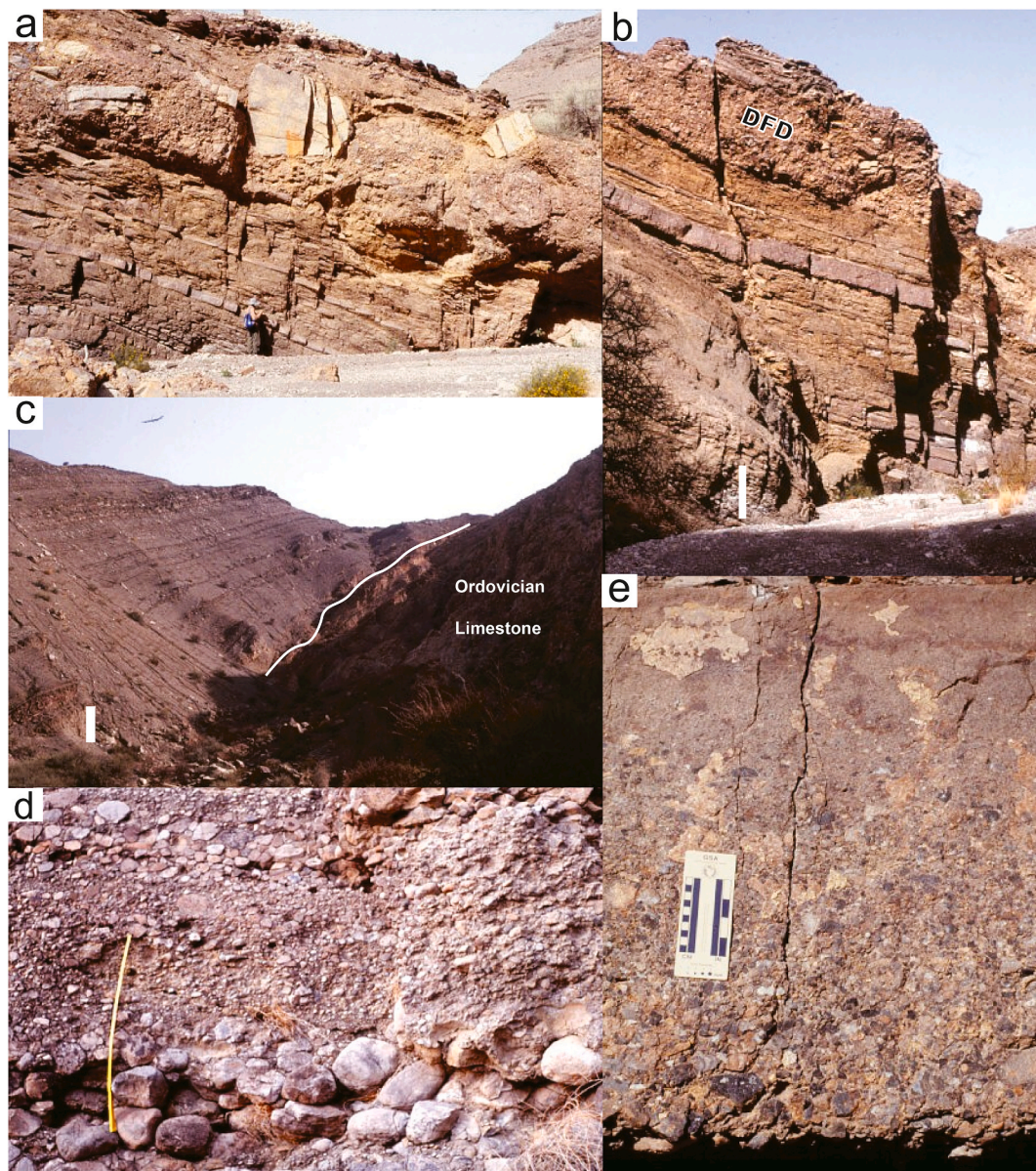


Fig. 10. Valley-Glacier-Retreat Facies Association (fjord association). Las Lajas locality. (a): debris flow (breccia) deposits with large limestone boulders, (b): debris flow deposit bed (DFD) capping a section of sandstones and mudstones, white bar scale: 2 m, (c): basal interval of siltstones and sandstones with dropstones with subordinate conglomerate beds resting on Ordovician limestone basement present along the paleovalley axis, white bar scale: 2 m, (d): normally graded conglomerate with coarse (boulder-sized) basal interval, yellow bar scale: 1 m, (e): normally graded pebbly to coarse-grained sandstone. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

association has been subdivided in four units that reflect four stages of evolution of the valley fill.

Stage 1 is characterized by ice-contact Gilbert-type deltas and proglacial (lacustrine) fans; it is divided into two facies associations: (i) a subglacial tillite and proglacial delta assemblage of massive or stratified conglomerates interbedded with subordinate sandstones and siltstones, and, (ii) proglacial lacustrine assemblage of thin-to medium-bedded, well-laminated siltstones and sandstones with dropstones with rare conglomerate beds, dominant down the axis of and away from the walls of the paleovalley (Fig. 10c). A study of pollen and spores indicates a nearshore marine environment for stages 2 and 3 (Césari and Bercowski, 1997).

The onset of *stage 2* is interpreted as a postglacial marine incursion with an increase in water depth punctuated by turbidity currents. Massive or laminated shales deposited throughout the paleofjord prevail in the basal interval of this stage. Conditions of relatively low coarse-

grained sediment input persisted during deposition of 135 m of section before a transition to stage 3.

Stage 3 represents the establishment of a more consistent sand supply with thick, sheetlike turbidite sandstones and coarse-grained fluvial sediment up dip in the delta. Several large-scale mass-transport events occurred during stages 2 and 3 (Fig. 10d and e).

Stage 4 records progradation of a coarsening- and shoaling-upward, coarse-grained fan delta fed by a fluvial system from the head of the paleovalley into the marine waters in the fjord.

Although mass-transport deposits are common throughout the paleofjord fill, they are most abundant in Stages 1 and 4. They include a wide range of facies including deformed beds, slumps and slides of various scales.

3.9. Facies model and evolution of fjord infill

Fjords are deep glacial valleys characterized by steep sides that have been invaded by the sea; their drainage patterns are relatively straight. The infill evolution of the Carboniferous paleofjords of the western Paganzo basin has been interpreted as the stratigraphic expression of deglacial sea-level rise and post-glacial rebound (Dykstra et al., 2006a, 2006b). Glacioeustatic sealevel rise, meltwater discharge, and subsequent postglacial rebound were the key external controls on the depositional systems. The abundance of deposits derived from mass wasting suggests that the Paganzo paleofjords were associated with tidewater glaciers and therefore can be considered temperate fjords (Powell, 2005; Howe et al., 2010). The evolution of these narrow, overdeepened valleys exhibits four distinct stages.

- (i) Incision of paleovalley and deposition of subglacial diamictites and an ice-contact delta: characterized by the glacial incision carved in the basement, formation of striated bedrock (hard bed) pavements on the valley floor by glacial abrasion and subglacial deposition. Paleovalley floors tend to be flat along the axis (Alonso-Muruaga et al., 2018) or with steps (Kneller et al., 2004). U-shaped cross sections with steep sides are common (Kneller et al., 2004; Dykstra et al., 2006a; Aquino et al., 2014). The infill consists of lodgment till deposits, ice-contact deltas and subaqueous outwash fans.
- (ii) Early Transgressive stage: this stage records a progressive decrease of sediment delivery due to the retreat of the ice mass. It is characterized by resedimentation of subglacial material by subaqueous sediment gravity flows and slumps in proglacial settings mainly represented by stratified diamictites and ice rafted deposition (dropstones and iceberg dump diamictites). Large-scale mass-transport deposits occur coincident with this glacioeustatic flooding, which may indicate that there may be a strong correlation between mass transport and increased relative sealevel. Poorly consolidated glacially deposited sediments are prone to instability and subsequent mass transport and resedimentation due to high sedimentation rates (Elverhøi et al., 2002; Benn and Evans, 2010). Finer-grained deposits may host ichnofossils and body fossils. Normal marine or brackish conditions may dominate this stage.
- (iii) Maximum flooding (late transgressive stage): characterized by almost complete shut-off of sediment delivery. It consists of black shales or laminated mudstones related to a marine incursion that flooded the valleys; normal marine or brackish conditions may dominate this stage. Minor sediment input in the parts of the depositional system distal to the sediment source where marl and organic rich horizons, indicative of periods of clastic starvation, may be present. Organic rich layers indicate a significant imbalance between accommodation space and sediment supply. In modern fjords organic carbon content is found inversely proportional to the suspended load delivered to these fjords (Syvitski et al., 1990). Slumps and debris flow deposits with sources off the paleofjord steep walls may be present. Dropstones may be present in the trunk valley but not in tributaries (Kneller et al., 2004). The scarcity to absence of dropstones suggests a major contraction of the ice with ice retreat away from a shoreline.
- (iv) Highstand: progradation of a fluvial-deltaic system from the head of the paleovalley into the fjord. Common presence of Gilbert-type, deltaic clinoforms, formed by progradationally stacked coarsening and thickening cycles of mudstone-sandstone and exceptionally conglomerates (Martínez and Astini, 1992; Dykstra et al., 2006a; Limarino et al., 2014a; Aquino et al., 2014). This reinvigoration of the sediment supply may have been associated with postglacial-rebound-generated uplift, which is known from the Paganzo Basin (Bercowski and Milana, 1990; Kneller et al., 2004; Limarino et al., 2002). A late highstand phase is

characterized by fluvial sedimentation in proximal positions and transitional deltaic facies basinward.

The dropstone-bearing section assigned to the transgressive interval (early transgressive stage) in western Paganzo contains the low diversity *Didymaulichnus* trace-fossil assemblage, interpreted to represent brackish-water conditions in a stressful environment (Buatois et al., 2006, 2010). This interval is overlain by non-bioturbated black shales (late transgressive stage) succeeded in turn by siltstones and current-rippled fine-grained sandstones with a *Mermia* trace-fossil assemblage (Buatois et al., 2010), interpreted as the product of delta-fed underflow currents and suspension fallout. The presence of non-bioturbated beds with acritarchs in this *Mermia*-bearing interval indicates the existence of intermittent periods of brackish-water conditions (Buatois et al., 2010).

3.10. Submarine-retreat facies association (glaciomarine apron association)

The basal fill of these units is characterized by massive, mostly featureless clast-supported conglomerates with few striated and polished clasts indicative of glacial material (El Paso and Majaditas Formations) followed invariably by a fining-upward succession of proximal glaciomarine massive to stratified clast poor diamictites, thinly bedded diamictites with ice-rafted debris (IRD), mudstones with isolated dropstones, interpreted as the combined product of rain out from icebergs and melt water plumes, gravity flows (mostly cohesive debris flows) and ice rafting (López-Gamundí, 1989, 1991), capped by postglacial IRD-free mudstones. Locally contorted sand masses in diamictite beds, associated with low- and high-density turbidites, suggest a mass-emplacment mechanism (e.g., Ansilla Formation, López-Gamundí, 1986b). The massive to poorly bedded diamictite, locally clast supported and occasionally with a-axis clast imbrication, below the intertill boulder pavement (one type of diamictite underneath and a different one above, Figs. 11a and b) in the upper half of the Hoyada Verde glacial succession is at least partially interpreted as subglacial deposits. The esker-like sandstone channels encased in the structure-less diamictites have been interpreted as deposited in subglacial tunnels (González, 1981). Further evidence of subglacial abrasion is provided by an intratill boulder pavement (encased by the same diamictite type below and above the pavement, Figs. 11c and d) present in the lower half of the Hoyada Verde section (López-Gamundí et al., 2016). Both inter- and intratill boulder pavements were probably originally laid down as residual gravel lag surfaces, formed by the erosion of underlying diamictite when wave and/or tidal reworking of fines was effective and subsequently abraded by advancing glaciers (López-Gamundí et al., 2016). The upper interval in the Hoyada Verde is characterized by well-bedded diamictites with scarce sandstone lenses. Two diamictite subtypes (thick-bedded diamictites with an average thickness of 50 cm and common soft-sediment deformation and thin-bedded variety with thicknesses exceptionally up to 20 cm) have been interpreted as subaqueous gravity flow deposits. This diamictite section grades up to pebbly shales facies. Dropstones present in the thin bedded diamictites and pebbly shales indicate continuous ice rafting.

The Hoyada Verde inter-till boulder pavement is overlain by a thin (up to 30 cm), discontinuous, patchy veneer of massive diamictite followed by a thin (less than 30 m) interval of thin-bedded diamictites with IRD, interpreted as the combined product of subaqueous, cohesive, debris flows and ice rafting (López-Gamundí, 1991). The thin-bedded diamictites grade upward into postglacial IRD-free shales. The thin patchy massive diamictite potentially represents the only record of subglacial sedimentation immediately after glacial abrasion that produced the intertill boulder pavement.

Lateral variations along the margins of the Calingasta-Uspallata Basin show a range of facies associations in open to moderately topographically-confined conditions on one side to fjord-like facies

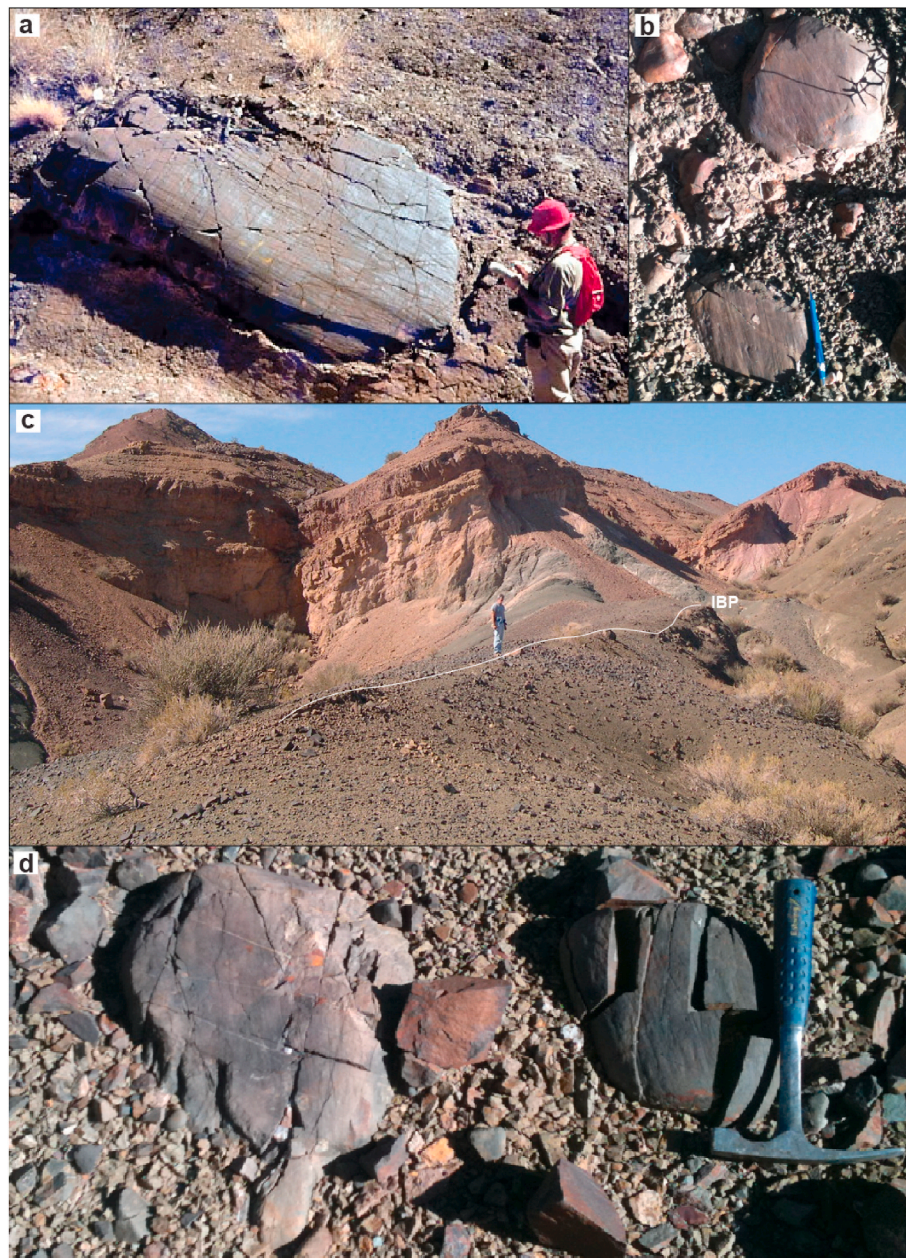


Fig. 11. Submarine-Retreat Facies Association (glaciomarine apron association) Hoyada Verde locality. a: intertill boulder pavement, faceted boulder with striations along a long axis, b: detail of intertill boulder pavement with faceted with striated surface, c: intratill boulder pavement (IBP) associated with massive diamictites below and above the pavement, d: detail of IBP with faceted clasts with striated surface.

associations indicative of high topographic confinement. The former association is well represented in the northern part of the basin by the Hoyada Verde (Fig. 11), El Paso (Fig. 12a and b) and Majaditas (Fig. 12c and d) formations whereas the Agua de Jagüel Formation (López-Gamundí, 1984) in the southern part of the basin constitutes the fjord-like or paleovalley end member (Ciccioli et al., 2008; Henry et al., 2010). The postglacial interval, dominated by fine-grained facies and uniformly distributed across the basin without major lithological changes, contrasts with the variability observed in short distances between the two end member associations of the underlying glacial sediments. The same IRD-free mudstone-rich postglacial section can be followed along an almost continuous belt of exposures for kilometers from the Hoyada Verde/El Paso outcrops to the north to the Majaditas section to the south with no major facies changes and common faunal and microfloral contents (Vergel et al., 2016). Further, facies-based and faunal correlations allowed extending this postglacial interval further to

the north to the Casa de Piedra Formation (Banchig et al., 1997; Lech et al., 1998) and to the south to the Agua de Jagüel Formation (Sterren et al., 2020) for a total distance of approximately 100 km (López-Gamundí et al., 2018).

Indirect evidence of such variability is provided by the analysis of the associated faunas. The postglacial mudstones with marine invertebrate of the AT/RM fauna of late Serpukhovian-Bashkirian age present in the Agua de Jagüel Formation were deposited in relatively restricted (paleofjord) parts of the basin. This fauna is characteristic of restricted environments with relatively low concentrations of oxygen and nutrients on the seafloor (Sterren et al., 2020). Geochemistry of strata in the Agua de Jagüel Formation also suggests the occurrence of anoxic bottom waters (Henry et al., 2010). These conditions are consistent with topographically-confined, fjord-type, indented coastlines as proposed by Henry et al. (2010) for the Agua de Jagüel section. The presence of a very impoverished record of the AT/RM Fauna in the Agua de Jagüel

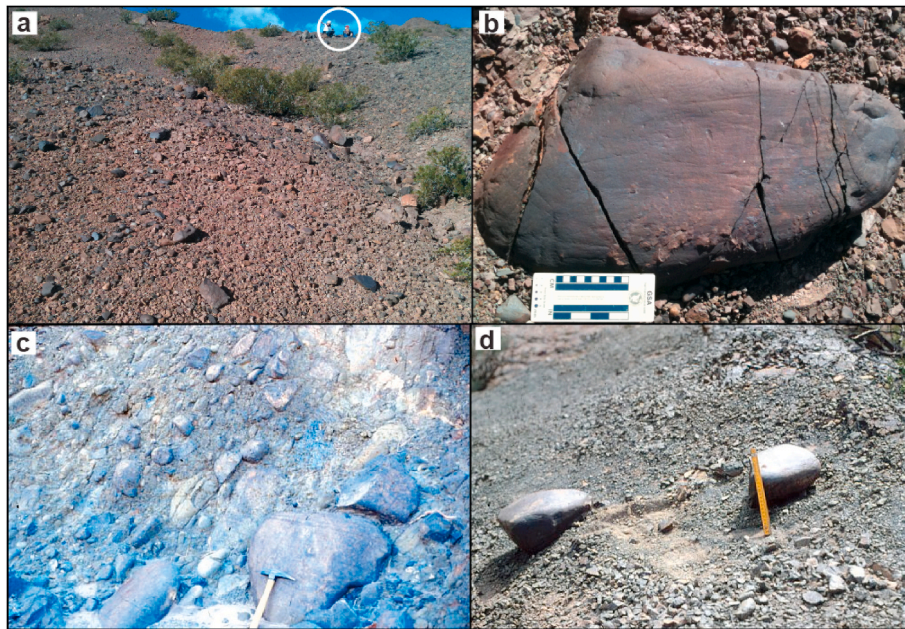


Fig. 12. Submarine-Retreat Facies Association (glaciomarine apron association) a: basal fill of El Paso Formation, massive diamictites, b: bullet-shaped clast with striations in basal diamictites, El Paso Formation, c: basal fill of Majaditas Formation, bouldery conglomerates with sand-rich matrix, d: faceted boulders in Diamictite Member, Majaditas Formation, scale: 25 cm.

Formation deposits in comparison with that of the El Paso Formation, could indicate a more stressful environment for benthic colonization, probably related to restricted oceanic water circulation (Sterren et al., 2020). The marine flooding over drastically different coast configurations and the availability of nutrient and oxygen in the water column would have propitiated the development of faunas with important differences in the taxonomic composition and the paleoecological structure (AT/RM and *Levipustula* Faunas) occurring at the same time interval.

3.11. Tarija Basin: paleovalleys and paleofjords in a glaciated basin

Glacial valleys and associated paleofjord incisions constitute the main evidence of direct ice activity in the Tarija Basin. Due to their dimensions subsurface seismic studies are necessary to fully describe their nature and extent. Multiple glacial cycles bound by glacial erosion surfaces (GES) each associated with their own internal complex geometries and facies partitioning have been described in the Tarija Basin. A glaciated basin margin with four distinct domains have been mapped by Bache et al. (2012) and Rousse et al. (2018) through the combined analysis of outcrops along the western foldbelt and subsurface (2D seismic and few wells and cores tied in the northern part of the grid) in the foreland (Chaco plains). These four distinct domains are: i) subglacial domain (ice-streams and tunnel valleys), ii) proglacial domain (subglacial channel/valley mouth and ice contact fans), iii) diamictite-prone, proglacial domain (with ephemeral meltwater discharge) and, iv) periglacial or offshore domain (cf. Dowdeswell et al., 2015). Deeply incised sand-filled paleovalleys, typically 500 m deep and several kilometers wide, have been described in the Machareti Group (Tupambi and Tarija Formations) and Mandiyutí Group (Escarpment Formation) (Helwig 1972; Salinas et al., 1978; Starck et al., 1993a, b; Eyles et al., 1995; Cardona et al., 2000; Girolodi and Algeria 2005; Bache et al., 2012; Rousse et al., 2018). A wide range of facies from subglacial, proglacial, periglacial to ice distal shelf settings is interpreted.

The palaeovalley infills have been classically interpreted as continental tillites and subaerial meltwater channels deposited by ice margins advancing and retreating across the Chaco Basin from ice centers to the south and east (Helwig 1972; Salinas et al., 1978; Caputo and Crowell 1985).

Four glacial cycles have been recently proposed for the Carboniferous section of the Tarija Basin (Rousse et al., 2018): Cycle 1 comprises the Itacua Formation, Cycle 2 comprises the glaciogenic deposits of the Tupambi and Tarija Formations, Cycle 3 includes the Chorro and Taiguati Formations, and Cycle 4 is represented by the Escarpment Formation. Each cycle boundary is marked by an abrupt change in facies and high relief erosion surface (GES). Recent seismic analysis allowed refinement of the nature and dimensions of these ice-carved paleotopographies. Two scales of glacially eroded troughs have been identified in seismic: a) a network of narrow and locally branching tunnel valleys, and b) mega-valleys (ice-stream, >10 km) separated by large interfluvies where tunnel valleys may occur. Paleovalleys, c. 500 m deep and c. 7 km wide, assigned to the base of Cycle 4 (Escarpment Formation) have been seismically mapped by Bache et al. (2012). The palaeovalley basal infill is imaged as a chaotic seismic facies overlain by an aggrading-prograding prism. The chaotic seismic facies and the uniform gamma-ray (GR) electrofacies (facies 1 of Bache et al., 2012) are interpreted as glacial deposits of various origins (rainout diamictites, lodgment tills and glaciofluvial sediments). This seismic facies is covered by prograding clinofolds (seismic facies 2), which commonly extends out to the interfluvies. Seismic facies 2 and GR electrofacies 2 (interpreted as cleaning/coarsening-upward siltstone to coarse-grained sandstone units) have been likened to deltaic progradations (Bache et al., 2012). Seismic facies 3 (with a channelized to slightly chaotic seismic facies) and GR facies 3 (dirtying/fining-upward cycles) that typically extend beyond the paleovalley incisions; this facies has been interpreted as highly aggrading fluvial deposits.

Four types of glacial or paleovalleys have been identified): (i) tunnel valleys, (ii) ice-stream cross shelf troughs (lowland), (iii) valleys carved by mountain glaciers (highland) and (iv) fjords (Bache et al., 2012). The widths of the Machareti paleovalleys range between 3 and 21 km, with a mean width of 7 km. The incisions are deep, with 500–700 m between the shoulders and the bottoms of the valley thalwegs. The palaeovalley drainage patterns are straight to slightly sinuous; their sections are both U- and V-shaped which indicate coexistence of abrasion and hydraulic processes of erosion. The drainage patterns are straight to slightly sinuous. Based on the width and length parameters, Bache et al. (2012) suggested that paleovalleys observed in the Tarija Basin correspond to

ice-streams that cut cross-shelf troughs and/or fjords.

The thin (approximately 60 m) Cycle 1 (Itacua Formation) is made up of proglacial diamictite and sand-prone gravity flow deposits often contain slumps and shales with dropstones (Souza Cruz and Rocha, 2008; Rousse et al., 2018).

The Intra-Machareti, Tupambi (Cycle 2) mostly consists of subglacial sandstone channel and subordinate diamictites; it represents an early stage of heterolithic infill during an ice maximum phase. The cycle is characterized by subglacial and glaciofluvial facies (cross-bedded sandstones and conglomerates, subordinate debrites (diamictites)) deposited beneath ice-sheets due to meltwater circulation. Cross bedding and climbing ripples in sandstones indicate paleocurrents towards the W-NW (Souza Cruz and Rocha, 2008). The basal part of the cycle is erosive and formed by conglomerates with clasts up to boulder size (Souza Cruz and Rocha, 2008). The Tupambi sediments fill the basal portion of the paleovalleys onlapping the irregular paleorelief (Starck et al., 1993a; Starck et al., 2002); upsection they cover the interfluvial developing significant lateral continuity (Starck and del Papa, 2006). Dimensions of these sand-prone and diamictite-prone infills of glacial paleovalleys range between 1 and 5 km in width and around 100 m in thickness, and have been mapped in the subsurface for tens of kilometers. These Tupambi paleovalleys show a NW/SE orientation with perpendicular, smaller tributary valleys (tunnels) (Rousse et al., 2018). Exceptionally paleovalleys form elongated depressions at least 100 km long, limited by the 2D data set extent and by erosion of the glacial succession off the Izozog High (Bache et al., 2012).

The Tarija Formation (upper part of Cycle 2) is characterized by proglacial to ice-marginal gravity flows (mostly debris flows) deposits and channelized sandstones (diamictite-prone facies assemblage) in its lower half (Fig. 13 a and b) and an upper half made up of turbidites and stratified diamictites with scattered faceted and striated (Fig. 13 c), outsized clasts in a ice distal setting (Frakes and Crowell, 1969; Frakes et al., 1969; López-Gamundí, 1986a; Starck et al., 1993a; Rousse et al., 2018). Glacial abrasion in the Tarija Formation is evidenced by the presence of striated surfaces on sandstone bed tops in several levels in the Tarija Formation (Starck et al., 1993b). Along the outcrop belt López-Gamundí (1986a) subdivided the Tarija Formation succession into two major facies (1 and 2) and a subordinate facies (3):

- 1) Diamictites (pebbly mudstones) with low M/C ratios. Subfacies (1a) massive to poorly stratified diamictites (MCS = 50 cm) (Fig. 13 a); and (2b) thinly-bedded (15 cm), normally graded pebbly mudstones (MCS = 10 cm).
- 2) Sandstones in lenticular bodies of variable dimensions (3–20 m wide, 1–5 m thick) with ripples, small-scale cross bedding and upper regime planar lamination. Soft-sediment deformation (due to slumping and differential loading) is common in the sandstone bodies (Fig. 13 b).
- 3) Thinly bedded fine-grained sandstones, occasionally with ripple-cross lamination, and laminated mudstones.

Thin layers of fine-grained sandstones and mudstones in the lower part of the Tarija Formation contain grazing trail trace fossils of the *Mermia* assemblage (Buatois and del Papa, 2003). Taxonomic composition, nature of preservation, and absence of typical ichnotaxa of marine environments suggest freshwater conditions in fjords affected by a strong discharge of fresh water due to melting of the ice masses (Buatois et al., 2010)

Marine invertebrate fossils have been documented from the shales of the Taiguati Formation (uppermost part of Cycle 3). The association is characterized by bivalves, gastropods and brachiopods of the *Levisputula* Biozone (Rocha-Campos et al., 1977; Trujillo Ikeda, 1989).

The Escarpment Formation (Cycle 4) is characterized by proglacial outwash fan/ice-contact fan sand prone facies made of high-energy, scoured-based, coarse-grained sandstones and conglomerates with tabular and trough cross-bedding and finer-grained sandstones with climbing ripples (Starck and del Papa, 2006; Souza Cruz and Rocha, 2008; Rousse et al., 2018) and subordinate diamictites in the lower third. Paleovalleys with a predominant SE–NW direction have been previously described in the Escarpment Formation (Helwig, 1972; Salinas et al., 1978; Eyles et al., 1995). This paleotopography led to the deposition of more than 500 m of the Escarpment Formation in the valley axes and it can be absent in the paleohighs or interfluvial (Starck and del Papa, 2006).

Apart from a local 150 m deep incision at the top of the Escarpment Formation in the southernmost part of the basin, the overlying San Telmo Formation shows a constant thickness, with stratigraphic markers traceable for several tens of kilometers (Starck and del Papa, 2006). No major evidence of paleovalleys due to glacial abrasion has been found.



Fig. 13. Tarija Formation. a: massive diamictites with deformed sandstones, b: sandstone bed with soft sediment deformation encased in massive diamictites; note partial disaggregation of sandstone bed, c: granite clast with striations, scale length: 15 cm.

Souza Cruz and Rocha (2008) have described three facies associations in the San Telmo Formation through core data: (a) fine-to coarse-grained, well sorted sandstones with cross-bedding, interbedded with subordinate mudstones, interpreted as fluvial braided deposits or as ephemeral channels deposits formed in arid climate, (b) fine-to medium-grained, locally deformed, normally graded sandstone interpreted as low density Ta-e Bouma turbidite beds interbedded with lacustrine delta-front shale, and, (c) fine-to coarse-grained massive, mud-rich sandstones with scattered granules, interbedded with fine-to very fine-grained, sandstones with climbing ripples, interpreted as hyperconcentrated gravitational flows. The basal part of San Telmo Formation (Yaguacua Member) comprises shales and diamictites with polished, faceted, and striated clasts up to 5 cm in diameter (Starck and del Papa, 2006). In subsurface marine gastropods and brachiopods have been described in the Yaguacua Member (Rocha Campos et al., 1977). The Yaguacua Member passes upsection to fluvial sandstones interbedded with large-scale, cross-bedded eolian sandstones (Chimeo Member).

4. Deglaciation sequences an operational tool of analysis

Deglaciation is commonly inferred from sedimentary successions that include ice-contact deposits overlain by more ice-distal glaciomarine depositional facies (Boulton, 1990; Lønne et al., 2001; Jakobsson et al., 2011; Dowdeswell et al., 2015).

Deglaciation sequences or successions constitute the main record of pre-Pleistocene glacial activity in sedimentary basins reflecting deposition mainly during the retreat of ice sheets, ice caps and alpine glaciers and successive deglaciation. The nature and mode of deposition of deglaciation sequences are mainly controlled by complex interactions between the rate of ice margin retreat (slow, punctuated, rapid), changing morphology and cross-sectional area of glacial valleys/fjords (pinning points), coupling of ice and sediment dynamics, glacial dynamics, and evolution of the sediment entry points (Brookfield and Martini, 1999; Bennett and Glasser, 2009; Cowan et al., 2010; Brinkerhoff et al., 2017).

Visser (1997) introduced the concept of deglaciation sequences for the well established glaciation in the Karoo and Kalahari basins of Southern Africa. He introduces the term deglaciation sequence to designate “an upward-thinning sediment body (or package) deposited seaward of the ice-grounding line during a major recessional phase of a marine ice margin”. He defined four deglaciation sequences in the Karoo Basin which have also been identified in the Kalahari Basin (Bangert et al., 1999; Stollhofen et al., 2000, 2008).

Although deglaciation sequences show high facies variability controlled by local tectonic setting and ice sheet retreat patterns they all have a common vertical stacking pattern. A deglaciation sequence is defined as an upward fining succession of (i) basal massive diamictites, (ii) stratified diamictites, (iii) sandstones deposited mostly by gravity flows in a variety of subenvironments (outwash, grounding line fan, and delta front), (iv) mudstones and rhythmites derived from turbid melt-water plumes with IRD.

Since its introduction this concept has been applied in other basins where the late Paleozoic glacial record of southwestern Gondwana has been studied e.g., the Paraná basin in Brazil (Vesely and Assine, 2006) and Calingasta-Uspallata and Paganzo basins in west central Argentina (Limarino et al., 2002). A similar approach has been used by Rouse et al. (2018) to study the stratigraphically complex multiple stacked or nested glacial sequences of the Carboniferous of the Tarija Basin. Each sequence is defined as “a major glacial advance/retreat cycle” and hence can be likened to the deglaciation sequence. The concept has been extended to other pre-Pleistocene glaciation such as the late Ordovician one mainly recorded in North Africa (Ghiene, 2003; Le Heron et al., 2010; Moreau, 2011).

The passage from glacially influenced deposits to postglacial sediments (herein labeled as the glacial - postglacial transition) is defined as the transition from two key lithologies: glacial diamictites and

postglacial mudstones (Visser, 1996). Three types of transitions were defined:

The *type 1* transition sequence consists of clast-poor diamictite deposited by rain-out and sediment gravity flow processes, with an abrupt contact with the overlying mudrocks. This transition is interpreted as the result of collapse of the ice shelf caused by rapid ice margin retreat. This glacial-postglacial transition type is similar to the base-cut TST (transgressive systems tract) for a period of rapid ice melting characterized by low sediment supply rate with respect to sea-level rise rate (López-Gamundí, 2010). This transition type could also be produced by ice retreating up a fjord with the ice retreating across subbasins. In this scenario debris are being trapped within the subbasin closest to the ice and icebergs being trapped by the sill at the end of the nearest subbasin.

The *type 2* transition sequence consists of clast-poor diamictite with sandstone bodies overlain by mudrock with dropstones of ice-rafted debris (IRD) or thin beds of diamictite interpreted as discrete debris flows, often with IRD. Deposition of the diamictite facies occurred during oscillations of an ice front with “short stable periods of surface melting of the ice resulting in the formation of small subaqueous outwash fans” (Visser, 1996).

The *type 3* transition sequence consists of boulder-rich, locally clast-supported diamictite interbedded with massive or sandstone-bearing diamictite overlain by mudrock with IRD or thin interbedded debris-flow diamictites. Visser (1996) assigned this transition type to “highly unstable ice front conditions and fluctuating sea-levels with periods of rapid deposition probably by collapsing ice streams.”

The deglaciation sequence framework proposed by Visser (1997) can be used as a reference for the adjacent Kalahari Basin and further for the Paraná Basin, Falkland/Malvinas and tentatively for the VFB/Claromecó glacial sections. Glaciation in the Karoo Basin, where four deglaciation sequences have been recognized, is considered as the product of deposition by an early, stable, slowly retreating, predominantly subpolar marine ice sheet becoming a late stage, more temperate, unstable, more rapidly retreating ice margin (Visser, 1989, 1991, 1997). The uppermost three cycles (DS II, III, IV) range up to approximately 350 m in thickness.

In the Paraná Basin, mostly located in Brazil with a southernmost extension into Uruguay (Cuenca Norte, De Santa Ana et al., 2006), a well-based stratigraphic framework proposed by França and Potter (1991) subdivided the Itararé Group into three stratigraphic units: Lagoa Azul (present primarily in the subsurface with a maximum thickness of over 500 m), Campo Mourão (with a maximum thickness of 900 m), and Taciba (up to 420 m of thickness) formations. Each unit is made up of a fining-upward facies succession composed of a thick basal sand-rich interval followed by a fine-grained section with diamictites, sandstones, rhythmites, and shales (dos Santos et al., 1996; França and Potter, 1991; Rocha-Campos et al., 2008). Each of these successions is capped by marine shales, from base up, namely, Folhelho Roncador, Lontras Shale and Passinho Shale/Rio do Sul. This tripartite subdivision of the Itararé Group already reflects the characteristic retrogradational stacking pattern of the deglaciation sequences. Subglacial tillites resting on striated and/or glaciotectionized basement (Bigarella et al., 1967; Rocha-Campos et al., 1988; Pérez-Aguilar et al., 2009; Rosa et al. 2016; Assine et al., 2018) and deformed glacial sediments identified at different stratigraphic levels in the Itararé Group also were used as indicative of multiple ice advances (dos Santos et al., 1996; Fedorchuk et al., 2019; Rosa et al., 2019). Further, numerous glacially striated, soft sediment surfaces suggest a complex history of higher frequency, probably local, glacier advances and retreats along the basin margins (Vesely and Assine, 2002; Trosdorf et al., 2001, 2005). Many of these surfaces have been reinterpreted as scour by iceberg keels (Rocha-Campos et al., 1994; Vesely and Assine, 2014).

A later refinement (Vesely and Assine, 2006) identified five unconformity bounded sequences with fining-upward vertical facies in the Itararé Group: one in the basal Lagoa Azul, three in the Campo Mourão Formation and one in the uppermost Taciba Formation. Laterally

discontinuous tillites associated with subglacial features indicative of glacial abrasion such as striated pavements (Bigarella et al., 1967; Rocha-Campos et al., 2008; Pérez-Aguilar et al., 2009) and boulder beds (Rocha-Campos et al., 1968, 1969, 1976; Lucio and Rocha-Campos, 2000) occur at the base of the sequences. The subglacial component made up of thin lodgment till layers is commonly absent at the base of these sequences. The lowermost portions of the sequences are dominated by polymictic conglomerates and cross-bedded sandstones deposited by meltwater-derived, highly concentrated hyperpycnal flows in subaqueous proximal proglacial outwash fans (Vesely and Assine, 2006; Vesely, 2007). This sandstone-conglomerate assemblage is usually overlain by diamictites, parallel-bedded and rippled sandstones, and Bouma-like turbidite facies sequences interpreted as deposits of distal outwash fan lobes (Vesely and Assine, 2006). The tops of the deglaciation sequences are positioned in mud-rich marine sediments with glacial activity reduced to IRD material (dropstones) or no evidence of glacial influence. Variations of transition types (1, 2 and 3, Visser, 1996) are controlled by the interplay between sediment supply rate and relative sealevel rise rate during glacier retreat.

Vesely et al. (2018) identified an additional type of glacial advance and retreat succession for cyclic diamictite-mudstone succession in the Bashkirian-Moscovian Campo do Tenente Formation in the Paraná Basin. The succession is composed of two facies, F1 and F2. F1 consists of sheet-like, 8-m-thick, clast-poor diamictites lacking striated, polished, or faceted clast and contains plastically deformed m-thick blocks of fine to gravelly sandstones. Facies F2 is up to 6 m thick and consists of a 20-cm-thick basal zone of IRD-free rhythmites that grade upward into rhythmites containing multiple cycles of increasing-decreasing IRD. Lonestones include abundant sediment pellets up to 5 cm in diameter consisting of mud, sand, and gravel and striated extrabasinal clasts up to 40 cm in diameter. The F2 facies is punctuated by crudely stratified, cm-scale clast-poor diamictite. The uppermost part of the F2 facies consists of up to 2 m of IRD-free shale which is in sharp contact with Facies 1 diamictite in the overlying cycle. Vesely et al. (2018) attributed these facies to glacial advance and retreat cycles. Facies F1 is interpreted as non-glacial mass-transport deposits derived from delta collapse during intervals of maximum ice retreat onto the land surface, while Facies F2 represents advance of the ice front into the marine basin and the delivery of IRD to ice-distal sites during maximum ice advance.

Dip-oriented cross sections with surface and subsurface data show a range of facies associations in valleys filled by relatively thin successions comparable with glacially carved shallow troughs along the basin margin to thicker deglaciation sequences basinward. This fjord facies association is well preserved particularly along the southeast margin of the basin. Although, not all of the valleys are the result of glacial activity, especially those in the Morro Do Papaléo area of eastern Rio Grande do Sul state in the southeastern margin of the basin (Fedorchuk et al., 2019).

The Permo-Carboniferous succession of the Falkland/Malvinas Islands (Fig. 1) is lithostratigraphically known as the Lafonia Group. The basal glacial section is known under different names, Lafonian Diamictite (Frakes and Crowell, 1967) and Fitzroy Tillite Formation (Aldiss and Edwards, 1999). The Fitzroy Tillite Formation overlies the Bluff Cove Formation probably either unconformably or disconformably, while in the west it overlies the Devonian Port Stanley Formation unconformably. It is composed of massive to very weakly stratified thick-bedded sandy diamictites, with minor intercalations of laminated mudstones and small channelized sandstone beds. Two main facies associations, interpreted as grounded ice-sheet conditions which grades basinward into marine conditions with floating ice have been identified; the transition between both associations have been described as lateral proximal-distal changes and also vertically with massive diamictites with evidence of subglacial deposition at the base grading up to dominantly thick-bedded diamictites and IRD-bearing shales (Frakes and Crowell, 1967). In the west (West Falkland Island), they proposed that a grounded ice sheet led to the formation of subglacial deposits that

incorporated linear sand bodies originating as eskers. Linear sandstone bodies, some are than 1 km long, up to 10 m wide and 5 m thick (Frakes and Crowell, 1967), are enclosed within the diamictite. These bodies consist mainly of well-sorted fine- and very fine-grained sandstone, with some siltstone, medium- or coarse-grained sandstone and clast-supported conglomerate. These linear sand bodies, interpreted as subglacial or englacial eskers, and sand sheets, interpreted as subglacial deltas formed at an ice-sheet grounding line. Farther east, the diamictite facies with soft sediment deformation seems to be related to the slumping and remobilization of till deposits on a submarine slope. In East Falkland Island a massive diamictite was considered to be a glaciomarine rainout (?) deposit in an open shelf (Frakes and Crowell, 1967).

In the 850-m-thick diamictite-dominated Fitzroy Tillite Formation two laminated intervals stand out. One interval, the top of which is approximately 120 m below the top of the Fitzroy Tillite, exposed in East Falkland has a thickness of 20 m and consists of alternating diamictite and laminated mudstone beds with sporadic dropstones (Bellosi and Jalfin, 1984; 1987). In West Falkland a laminated 5–6 m thick unit of sandstones and siltstones with sporadic dropstones, placed approximately 240 m below the top of the Fitzroy Tillite has been described as a distinct Quark Pond Member by Aldiss and Edwards (1999) and Meadows (1999). Although the exact stratigraphic levels of these two intervals within the Fitzroy Tillite Formation is still a matter of debate Horan et al. (2018) speculated that these intervals reflect the vertical 'stacking' described by Visser (1997a) from the Dwyka deglaciation sequences.

The glacial interval (Fitzroy Tillite Formation) grades up to the postglacial Port Sussex Formation (Bellosi and Jalfin, 1984; López-Gamundí, 1997; Horan et al., 2018). The Fitzroy Tillite interval is succeeded abruptly by the thinly interbedded diamictites, turbidite sandstones and laminated mudstones of the Hells Kitchen Member (Port Sussex Formation). This transition between the glacial Fitzroy Tillite Formation and the postglacial Hells Kitchen Member and Black Rock Member of the early Permian Port Sussex Formation has been thoroughly described in outcrops (Frakes and Crowell, 1967; Bellosi and Jalfin, 1984) and borehole cores (Stone, 2011; Horan, 2015; Horan et al., 2018).

The upward transition from the Hells Kitchen Member into the entirely postglacial mudstones of the Black Rock Member occurs for about 2 m (Horan et al., 2018). In cores the Fitzroy uppermost diamictites are succeeded by an approximately 8 m - thick succession of alternating intervals dominated by mudstones with IRD (Hells Kitchen Member Zones 1 and 3) or granular and pebbly stratified diamictites (Hells Kitchen Member Zones 2 and 4). This succession is overlain by the postglacial interval of IRD-free mudstones with parallel and wavy lamination and intraclasts (Black Rock Member Lower Section) followed by (Black Rock Member Upper Section) organic rich mudstones. The total organic carbon content obtained from cores does not exceed 3% (Horan, 2015), although it can reach 40% locally (Marshall 1994a; Trewin et al., 2002; Stone and Horan, 2016).

A close relationship between the late Paleozoic successions of the Karoo basin and the Falkland/Malvinas Islands has been highlighted on paleomagnetic, paleogeographic and stratigraphic grounds (Visser, 1987; Marshall, 1994b; Aldiss and Edwards, 1999; Trewin et al., 2002). A stratigraphically close correlation of the Karoo Supergroup from the Karoo Basin with the Lafonia Group of the Falkland Islands has been proposed by Aldiss and Edwards (1999) and later refined by Trewin et al. (2002). In addition to the correlation of the Fitzroy Tillite Formation with the tillites of the Dwyka Group in South Africa, the Port Sussex Formation was thought to correlate with the lower part of the Ecca Group; specifically, the Hells Kitchen Member with the lower part of the Prince Albert Formation. Two cycles in the Falkland/Malvinas (massive diamictite overlain by the Quark Pond Member and diamictites overlain by the Hells Kitchen Member) defined by Aldiss and Edwards (1999) have been tentatively equated with the uppermost Cycles 3 and 4

defined in South Africa (Visser 1997). The correlation can be further supported by the lithological similarities between the Hells Kitchen Member and the Prince Albert Formation at the top of Cycle 4 in the Karroo basin. Recent CA-TIMS dates for the basal Prince Albert provided a $282.17 \pm 0.32/-0.44$ Ma age (Griffis et al., 2019).

The Sauce Grande Formation exposed in the Ventana foldbelt (Fig. 2) consists mostly of glaciomarine diamictites with subordinate sandstones and shales (Coates, 1969; Amos and López-Gamundí, 1981a, b). The exposed strata range in thickness between 500 and 1100 m (Andreis et al., 1987). It has been subdivided into three cycles (megaciclos, Andreis and Torres Ribeiro, 2003) or members (Lesta and Sylwan, 2005). The lower half to two thirds of the Sauce Grande Formation comprises a fining upward succession (megaciclo medio of Andreis and Torres Ribeiro, 2003) of mostly massive diamictites with scarce lenses of sandstone near the base and top (Coates, 1969; Andreis, 1984) capped by thin (>50 m) intervals of shale and sandstone, with horizontal lamination or ripples (megaciclo medio of Andreis and Torres Ribeiro, 2003). Palynological content found in the middle interval is accompanied by a solitary acritarch (di Pasquo et al., 2008) which, in addition to the presence of a solitary marine bivalve fossil (*Astartiella? pusilla*) found in the diamictites (Harrington, 1955), confirms the marine origin for the Sauce Grande Formation. The two lower megacycles of Andreis and Torres Ribeiro (2003) are interpreted here as deglaciation sequences in the Sauce Grande Formation.

The upper megacycle (megaciclo superior of Andreis and Torres Ribeiro, 2003) is characterized by diamictites, confined to the lower part, and sandstones and mudstone-dominated facies in the upper interval (Andreis and Torres Ribeiro, 2003). Diamictites, massive or thick-bedded with contorted sandstone beds, are associated with subordinate interbeds of rippled sandstones with evidence of wave and current action, conglomerates, rhythmite couplets of mudstone and fine-grained sandstone and shales with common dropstones (Amos and Lopez-Gamundi, 1981a,b; Andreis, 1984; Andreis and Torres Ribeiro, 2003). The glaciomarine diamictites have been interpreted as probably rain-out tills with variable participation of sediment gravity flows (Coates, 1969; Harrington, 1980; Andreis, 1984; Andreis and Torres Ribeiro, 2003). The uppermost Sauce Grande Formation grades up to the Piedra Azul Formation. This unit, which reaches 300 m in thickness, has been subdivided informally into a lower mudstone-rich interval and an upper sandstone-dominated section. The contact with the underlying Sauce Grande diamictites is gradational; thin-bedded (>30 cm) diamictites tend to disappear up section in detriment to mudstones (Andreis and Japas, 1996). Fine- and medium-grained sandstones with horizontal lamination and current and oscillatory ripples are dominant in the upper section of the Piedra Azul Formation (Andreis and Japas, 1996). An upper deglaciation sequence composed of the upper megacycle of the Sauce Grande Formation and the mudstone-dominated lower section of the Piedra Azul Formation is herein proposed. The presence of the *Eurydesma* fauna in the Bonete Formation suggests that the Sauce Grande-Piedra Azul glacial-postglacial transition is pre-late Asselian.

Based on the seismic facies and GR electrofacies coupled with facies analysis along the outcrop belt, Bache et al. (2012) defined in the Tarija basin deglacial sequences. Each sequence consists of a basal facies 1 dominated by glacial sedimentation followed by facies 2, predominantly deltaic, that in turn grades upsection to fluvial deposits of facies 3. A similar stacking pattern has been described for the Upper Ordovician paleovalleys of northern Africa filled by glacial deposits usually with coarse-grained deposits (conglomerates, diamictites) at the base, passing upsection into prograding deltas and further into aggrading fluvial deposits (Ghienne et al., 2003; Le Heron et al., 2004). Deposition during deglaciation or marine transgression has been suggested for this Upper Ordovician succession (Powell et al., 1994; Ghienne and Deynoux 1998; Le Heron et al., 2004; Ghienne et al., 2007).

5. Climate change drivers in basins of west central Argentina

During glaciation, a series of uplifts characterized the western margin of Gondwana adjacent to the Calingasta-Uspallata, Río Blanco, and Paganzo Basins. Major topographic elements included the Protoprecordillera, Pie de Palo Arch, Famatina Arch, Sierra de Chepes, and the Pampean Arch (Figs. 2 and 5). Controversy surrounds the elevation of the Protoprecordillera and some reports suggest that it was a relatively minor feature, and is instead referred to as the Tontal Arch (cf. Aquino et al., 2014; Valdez Buso et al., 2017; 2020; Milana and di Pasquo, 2019). However, radial glacial flow directions away from this feature and the occurrence of deep paleovalleys (fjords) at least 2.5 km deep (Agua de Jagüel paleovalley; López Gamundi and Martínez, 2000; Henry et al., 2010) indicate that the Protoprecordillera was a substantial topographic feature along the western margin of Gondwana, perhaps not unlike the modern fjords and uplands of southern Chile. This highland formed as an obducted, accretionary-prism/fold-and-thrust belt during the Chañic Orogeny (Ramos et al., 1984, 1986, 1988; López-Gamundí and Rossello, 1993; López-Gamundí et al., 1994; Limarino et al., 2002, 2006, 2014). The Pie de Palo Arch and possibly the Famatina Arch also served as glacial centers as suggested by ice flow directions, clast compositions, and detrital zircon ages (cf. López Gamundi and Martínez, 2000; Henry et al., 2010; Craddock et al., 2019; Pauls, 2020). The Sierra de Chepes (Fig. 5) in the eastern Paganzo Basin is cut by the narrow Olta-Malanján paleovalley. Its fill consists of alluvial fan, lacustrine and deltaic facies and it does not appear to have been glaciated (Moxness et al., 2018; Pauls et al., 2019, this volume; Pauls, 2020).

Collapse and breaching of the Protoprecordillera as a topographic barrier began in the Bashkirian and continuing into the Permian (Ramos, 1988; López-Gamundí et al., 1994; Limarino et al., 2002, 2006, 2014; Net and Limarino, 2006; Isbell et al., 2012). Evidence for this collapse includes: 1) disappearance of glacial deposits in the western basins and paleovalleys, 2) transgressive strata in basins on both sides and within paleovalleys of the Protoprecordillera, 3) occurrence of marine faunas in paleovalleys within the Protoprecordillera, 4) detrital sediment in the Andean Cordillera region sourced from the Pampean ranges, 5) subsidence within the paleovalleys (e.g., Agua de Jagüel paleovalley) that allowed for Pennsylvanian and Permian shoreface deposits to lap onto and overtop paleovalley walls (cf. López-Gamundí et al., 1994; Taboada, 1997; Net and Limarino, 2006; Limarino et al., 2006, 2014a, b; Henry et al., 2008, 2010; Isbell et al., 2012; Milana and di Pasquo, 2019; Pauls, 2020; Pauls et al., this volume). Concomitant with the collapsing Protoprecordillera was an outboard shift in the position of the active eastward-dipping subduction zone, which resulted in development of a wide (>200 km) Pennsylvanian-Permian volcanic arc along the present Argentine-Chilean border (Ramos et al., 1984, 1986; López-Gamundí and Breitzkreuz, 1997; Mpodozis and Ramos, 1989; Azcuy et al., 1999; Willner et al., 2005, 2008; Limarino et al., 2006, 2014; Limarino and Spalletti, 2006; Ramos and Folguera, 2009; Spalletti et al., 2012; Einhorn et al., 2015).

The formation of glaciers is dependent on insolation, sources of moisture, and the persistence of snow from year to year. Glacial formation depends on the relationship between global and regional driving factors (Fielding et al., 2008; Isbell et al., 2012; Montañez and Poulsen, 2013; Limarino et al., 2014b; Moxness et al., 2018; Pauls, 2020). A number of factors control insolation. These include: solar luminosity, orbital parameters (eccentricity, obliquity, and precession), abundance of greenhouse gases (primarily CO₂), latitude, and elevation (cf. Broecker and Denton, 1990; Benn and Evans, 2010; Isbell et al., 2012; Limarino et al., 2014b; Montañez and Poulsen, 2013). These factors combine to define an elevation for a given latitude above which local annual insolation is low enough to allow for the net accumulation of ice and the formation of glaciers. This elevation, known as the equilibrium line altitude (ELA), separates areas of net accumulation from areas of net ablation and hence the elevation at which glaciers form varies with latitude across the surface of Earth (Broecker and Denton, 1990; Benn

and Evans, 2010). The ELA is at or near sea level in the polar regions and rises to nearly 6,000 m in the tropics. The ELA increases in elevation during interglacial intervals and decreases during glacial epochs. Moisture sources also influence glacial formation by raising and lowering the ELA relative to the local elevation of the land surface and due to topographic orographic barriers. Moisture is a major limiting factor and can influence glacial formation and local climate due to an orographic effect as air rises over mountain belts (e.g., Protoprecordillera), moisture shadows (eastern Paganzo Basin), or due to proximity to a seaway (e.g., Chaco-Paraná and Paraná basins).

Evidence for the collapse of the Protoprecordillera, development of a western volcanic arc, and changing climatic conditions in the western basins appears to be concomitant. Initial elevation of the Protoprecordillera would likely have raised the range above the ELA, allowing for glacial nucleation. This range would have also resulted in an orographic effect for moisture laden air coming from the Panthalassan Ocean (cf. Broccoli and Manabe, 1997; Partridge, 1997; Ruddiman and Prell, 1997). However, moisture would have been limited to the Protoprecordillera and nearby ranges as a precipitation shadow would have prevented enough moisture from reaching the eastern Paganzo Basin to grow glaciers in the Sierra de Chepes (Moxness et al., 2018; Pauls et al., 2019; Pauls 2020). Collapse of the Protoprecordillera below the ELA in the late Bashkirian and the ensuing transgressing seas breaching the range likely delivered postglacial moisture throughout the Paganzo Basin, which allowed for peat accumulation in the basin. However, continued collapse of the range, withdrawal of the sea, and an elevating volcanic arc farther to the west would have inhibited glaciation in western Argentina due to subsidence of the range below the ELA, and the development of a new orographic barrier (the volcanic arc). These factors would have increased aridity across much of western Argentina during the middle-late Pennsylvanian and into the early Permian. This is in stark contrast to continued Pennsylvanian glaciation and the development of the ensuing Early Permian, coastal-plain, coal measures in the Paraná Basin. This west to east paleogeographic disparity across western Gondwana likely resulted from variations in moisture regimes as a widespread seaway in the eastern basins (e.g., Paraná Basin) allowed glaciers to form in the uplands of western Africa later in the Pennsylvanian, which delivered ice to eastern South America (cf. dos Santos et al. 1996; Visser, 1987, 1997; Vesely and Assine, 2006; 2011; Holz et al., 2008; Rocha-Campos et al., 2008; Fallgatter and Paim, 2019; Rosa et al. 2016; Fedorchuk et al., this volume).

6. Sequence stratigraphic framework of late Paleozoic successions

6.1. Applying sequence stratigraphy to glacial successions: challenges and limitations

One of the major challenges when a sequence stratigraphy analysis is applied to any sedimentary succession is the identification of a hierarchy of sequences. Hierarchical classifications are used in sequence stratigraphy to assign spatial and temporal order to the sedimentary successions. There is, unfortunately, a tendency of mislabeling and even sometimes transposing the hierarchical order of sequences mainly due to misinterpretation of the magnitude in space and time of their limiting boundaries. Another tendency that often leads to confusion is to infer a sequence stratigraphic framework for a depositional succession based on the study of an areally restricted outcrop or based on one or a few closely-spaced wells and outcrops. Without lateral continuity via well spaced outcrops or wells the discrimination of discontinuities of autogenic (caused by factors generated within the depositional system itself) and allogenic (induced in the system by external forces) origin gets blurred and is almost impossible to discern.

Noted, a few specific differences between glacial and non glacial successions are critical when sequence stratigraphic analysis is intended to be applied to glacial successions. These are, namely:

i) Compared with non-glacial successions, sea level on glaciated continental shelves is not necessarily the only control on baselevel; the sole of the glacier is as well. The glacier has the ability to erode major regional unconformities by glacial abrasion (glacial erosion surfaces, GES) well below sea level (Powell, 2015). The glacier bed is also a primary source of large volumes of sediment introduced below sea level. The glacier margin also marks the 'shoreline' position and provides additional control on accommodation (Zecchin et al., 2015).

ii) Sediment supply, although included as an important variable (Posamentier and Vail, 1988), is in practice commonly simplified for sequence stratigraphic models and interpretations. This approach includes only those changes controlled by base-level oscillations, and related fluctuations of accommodation space (the algebraic summation of eustatic sealevel changes and tectonically-induced subsidence/uplift, Weimer, 1992). In non glacial successions, therefore, the primary driving forces of stratigraphic variability are nickpoint migration, shoreline erosion and shelfal incision and the change of accommodation space. Due to very high sediment yield particularly during ice retreat (Boulton, 1990; Cai et al., 1997) glaciated basins can accumulate up to hundreds of meters in thickness over very short time periods of tens (locally at the ice front) to hundreds-to-thousands of years. This catastrophic pouring of large amounts of sediments is common during deglaciation in postglacial and, to a lesser degree, interglacial times.

iii) Sedimentation rates reach a maximum during sealevel rise (and minimum during the sealevel fall) in glaciated basins. This is therefore a major diversion from classic sequence stratigraphic models. The transgressive systems tract (TST), usually the thinnest systems tract in non-glacial sequences, tends to be quite thick. It is considerably thicker than the underlying, when present, LST in glacial successions. Deglaciation sequences, like the ones described herein for the Paraná and Karoo basins and proposed for the VFB-Claromecó basin, are pertinent examples of depositional successions dominated by high sediment supply from retreating ice in highly subsiding basins.

Attempts to utilize sequence stratigraphy for the analysis of Pleistocene glacial successions have been numerous (Boulton, 1990; Eyles et al., 1991; Brookfield and Martini, 1999; Powell and Cooper, 2002; Hansen, 2004; Cummings et al., 2011; Pedersen, 2012; Powell, 2015). Applications to pre-Pleistocene cases (Deynoux and Proust, 1991; McDougall and Martin, 2000; Vesely and Assine, 2004; El-ghali, 2005; Eschard et al., 2005; Lang et al., 2012; Melvin, 2019) in particular to postglacial flooding assigned to a transgressive systems tract (TST) within a sequence (Ghiene, 2003; Loi et al., 2010; López-Gamundí, 2010) suggest eustatic mechanisms followed by isostatic adjustments along the formerly glaciated basin margins (Haldorsen et al., 2001; Mottin et al., 2018; Dietrich et al., 2018).

6.2. 2nd order and 3rd order sequences in glaciated basins

The late Paleozoic successions of southwestern Gondwana exhibit a common tripartite motif with i) a lower part characterized by glacial and glacially-influenced sediments, ii) a predominantly marine middle interval initiated with the postglacial transgression and in some cases punctuated by successive shorter-lived ingressions, and iii) an upper interval characterized by progradation followed by continentalization accompanied in many sectors by increasing aridization. These tripartite successions are equivalent to 2nd order sequences, with estimated durations of 10–80 Myr (Vail et al., 1977) or 10–100 Myr (Miall, 1995). Examples of such evolution can be identified in the thick late Paleozoic successions of the Paraná and Karoo basins and in the VFB. In the Paraná Basin only the glacial basal interval reaches around 1,500 m of thickness in basin center positions (França and Potter, 1991). The basal unconformity of these sequences is a significant discontinuity between a basement of sedimentary or metasedimentary Lower Paleozoic or granitic rocks and glacial deposits (latest Devonian/Mississippian unconformity, Milani et al., 2007; Eohercynian unconformity, López-Gamundí and Rossello, 1993). Basement rocks are commonly

affected by glacial abrasion as documented by striated pavements and subglacial erosional landforms exposed mostly along the eastern margin of the Paraná basin (Bigarella et al., 1967; Rocha-Campos et al., 2008; Rosa et al. 2016) and along its southern extension in Uruguay (Assine et al., 2018; Fedorchuk et al., this volume). The Permo-Carboniferous megasequence of the Paraná Basin, composed of the Itararé (base), Guatá, and Passa Dois (top) groups, has been defined as an unconformity bound 2nd order sequence or, similarly, as a transgressive–regressive second-order cycle (in the sense of Vail et al., 1977), with a maximum flooding surface within the shales of the Palermo Formation (upper unit of the Guatá Group) or Irati Formation. It has been known under different names (Seqüência Carbonífera-Eotriássica; Milani et al., 1994; Superseqüência Gondwana I, Milani et al., 2007).

The Fitzroy Tillite Formation rests on the Devonian Port Stanley Formation (West Falkland Group) disconformably or unconformably. An angular discordance of less than 20° was observed at the slightly irregular basal contact between the underlying Port Stanley Formation and the Fitzroy Tillite deposits. The glacial deposits grade upward into the mudstone-dominated Port Sussex Formation; this unit contains thin metabentonite layers indicative of contemporaneous volcanism (Aldiss and Edwards 1999; Trewin et al., 2002). Similarly, the late Paleozoic tripartite succession exposed along the VFB is assigned to a second order sequence with the postglacial Piedra Azul and *Eurydesma* fauna - bearing Bonete shales (part of the transgressive systems tract, TST) as the turning point from the underlying glacial Sauce Grande interval to the prodeltaic to fluvial complex of the Bonete and Tunas deposits, interpreted here as the highstand systems tract of the second order sequence.

Sequences of the same order, although thinner, can be identified in the Paganzo and San Rafael basins. In both basins the middle interval is also punctuated by short-lived marine incursions (Limarino et al., 2006; Gulbranson et al., 2010; Pérez Loinaze et al., 2014). In western Paganzo, in addition to the postglacial transgression in the Guandacol Formation dated at 318.79 ± 0.1 Ma, two more marine incursions, dated from tuffaceous intervals at 315.46 ± 0.007 Ma and 312.82 ± 0.11 Ma occur (Gulbranson et al., 2010). Valdes Buso et al. (2020) identified in the overlying Tupe Formation a transgressive interval dated as 315.46 ± 0.007 Ma (assigned to Pennsylvanian transgression 1, Limarino et al., 2006), which consists of a heterolithic succession of fine-grained sandstones with current and wave ripple-cross lamination and carbonaceous mudstones deposited in shallow marine and probable estuarine settings (Facies Association 3, Pérez Loinaze et al., 2014). The interval, dated as 312.82 ± 0.11 Ma (Pennsylvanian transgression 2, Limarino et al., 2006), is made up of shallow marine interlaminated successions of very fine- and medium-grained sandstones with horizontal lamination, and ripple-cross laminations, mudstones with invertebrates of the TS biozone (Cisterna et al., 2002, 2011), and marls (Facies Association 5, Pérez Loinaze et al., 2014).

In tectonically active basins along the proto-Pacific margin, i.e., Calingasta-Uspallata (Fig. 1), the Permo-Carboniferous sequence shows a major erosional discontinuity that subdivides the succession in a lower sequence that comprises the glacial section and postglacial deposits and an upper sequence with the shallow marine to continental deposits (López-Gamundí and Martínez, 2003). This discontinuity loses its erosional nature basinward in the northwestern part of the basin where it is only detected by facies changes across the surface (Alonso-Muruaga et al., 2018).

The deglaciation sequences described in this contribution, particularly those thoroughly studied in the Paraná and Karoo basins and exposed along the VFB, are proposed to be assigned to third order sequences. They consist, when present, of a thin lowstand system tract (LST) of subglacial deposits with evidence of glacial abrasion at the base. The LST is followed by thick glaciomarine and glacially influenced sediments discerned in two distinct facies associations (paleofjord and glaciomarine apron). These facies are part of a thick TST that culminates with marine shales that reflect interglacial or postglacial conditions during ice retreat. Thus, the deglaciation sequences are proposed to be

incomplete third order sequences made up of LST-TST or exclusively TST. Based on zircon radiometric dates, the duration of each major deglaciation sequence in the Karoo Basin was calculated by Isbell et al. (2008) to be between 3.6 and 8.2 Myr in duration. This falls within the 1–10 Myr range assigned to third order sequences as suggested by (Vail et al., 1977) and (Miall, 1995). It is telling that Visser (1997) described the SD III and SD IV as ‘composite cycles consisting of a stacking of smaller units’. These smaller units of Visser (1997) could be assigned to smaller scale, fourth-order sequences.

6.3. Glacial erosion surfaces as sequence boundaries

The multiple ice advances and retreats of different magnitude commonly associated with subglacial abrasion and marine flooding respectively, complicates the sequence stratigraphic analysis of glacial successions, in particular, the hierarchical order of sequences. Since glacial advances and retreats can be of different scales, the basal surface of a deglaciation sequence, usually a glacial erosion surface (GES), can cover a wide spectrum from regionally significant to local unconformities. Thus, glacial erosional surfaces of different hierarchies, commonly sequence boundaries in glacial successions, can also lead to misinterpretation. A hierarchical subdivision of GSEs, similar to that used for bounding surfaces and associated architectural elements in fluvial (Allen, 1983; Miall, 1985; Miall, 1989), eolian (Brookfield, 1977; Kocurek, 1988; Havholm and Kocurek, 1994) and deep-water successions (Pickering et al., 1995; Cullis et al., 2018), emerges as a necessity. Abrasion surfaces developed on bedrock (glacially scoured and polished exhumed landforms), boulder pavements, and striated soft sediment surfaces can be tentative categories for bounding-surface hierarchies and their associated sequences of different orders.

6.4. Eustatically and isostatically induced sealevel changes

The complex relationship between water depth during glaciation, position of the glacial front, sedimentation rates, and eustasy and isostatic adjustments has a significant influence on the sequence stratigraphy of glacial successions. Deglaciation produces large amounts of debris that may be deposited in subaerial or subaqueous proglacial settings. With the associated ice front retreat, isostatic rebound can locally outpace deglacial eustatic rise causing a rapid relative sealevel fall (Boulton, 1990; Powell and Cooper, 2002) as inferred from several study cases (e.g., Nemec et al., 1999; Massari et al., 1999; El-ghali, 2005; Nutz et al., 2015; Dietrich et al., 2017).

The balance between eustasy and isostasy is tilted toward the former in distal areas of the basin. These eustatically dominated zones (Boulton, 1990; Dietrich et al., 2018) prevail in the open glaciomarine depositional parts of the basin whereas isostatically dominated zones (Boulton, 1990; Dietrich et al., 2018), dominated by ice marginal deposition, are common close the ice front along the basin margins (Fig. 14). Haldorsen et al. (2001) envisioned Dwyka sedimentation in the northeast Karoo Basin as partitioned into two regions: the proximal coastal region characterized by ice marginal deposition and a distal deep region with open glaciomarine sedimentation. The proximal coastal region facies association (Valley Facies Association, Isbell et al., 2008) consists of a basal interval of up to 100 m of thickness of fairly massive, heterogeneous diamictites with abundant and large clasts of up to more than 1 m in diameter followed by stratified diamictites, rhythmite sandstones and conglomerates. These Dwyka strata are mostly contained within valleys incised into bedrock (Visser and Kingsley, 1982; Cole, 1991; Von Brunn, 1996). This succession is followed by shales of the Pietermaritzburg Formation. The transition from the glaciomarine diamictite, to shallow water and deltaic sediments in the proximal coastal region, is an indication of regression most probably related to an isostatic rebound during and after the deglaciation. The relationship between isostatic uplift and eustatic sealevel rise is inverted in the deeper distal parts of the basin, where isostasy is not as sensitive to ice retreat. The distal deep

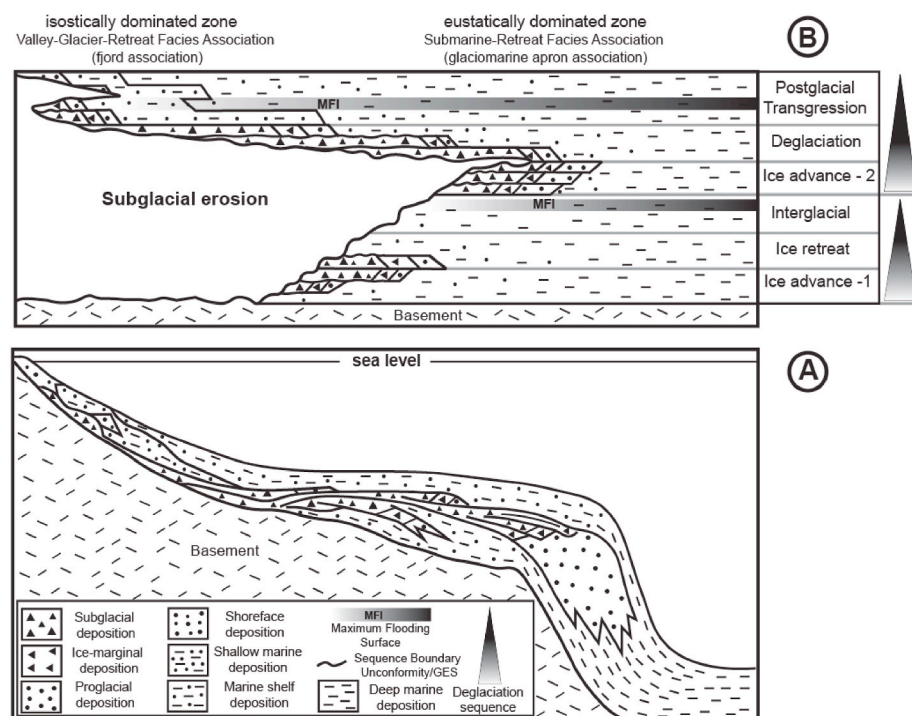


Fig. 14. Generalized (A) lithostratigraphic and (B) chronostratigraphic (Wheeler diagram) conceptual sections of glaciomarine depositional sequences, illustrating two phases of ice advance and retreat (phases 1 and 2). Note significant subglacial erosion of previous glacial and interglacial deposits along the basin margin. The missing section is visible on chronostratigraphic cross section B. Modified from Kurjanski et al. (2020). The lower deglaciation sequence (*sensu* Visser, 1997) is interglacial and the upper one culminates with a final deglaciation. The dip sections show two distinct domains (Boulton, 1990; Dietrich et al., 2018): a proximal, isostatically dominated zone with paleofjords (Valley-Glacier-Retreat Facies Association) and a more distal eustatically dominated zone (Submarine-Retreat Facies Association). The maximum flooding surface/interval (MFI) of the upper deglaciation sequence has better preservation potential of organic matter due to diminished clastic dilution.

region is characterized by a thick (up to 200 m) interval of homogeneous argillaceous glaciomarine diamictites with large regional extent, and blanketing character. These diamictites have clasts mainly of pebble/cobble size and very few boulders. They are overlain by stratified mudstones with dropstones, which in turn grade upwards to IRD-free marine shales of the Pietermaritzburg Formation (equivalent to the basal interval of the Prince Albert of the central and southwest parts of the basin). This open glaciomarine succession is equivalent to the deglaciation sequences defined by Visser (1997) in the areas of higher accommodation space of the basin.

In exceptional circumstances the glaciostatic sealevel rise and/or glacial front retreat were sufficiently rapid to completely overwhelm the sediment supply and create significant accommodation in which organic matter can be preserved and not be diluted by clastic input. This scenario is possible in linear or even in highly indented coastlines characterized by local and nearby source areas like in fjords and is associated with maximum flooding surfaces at the top of the postglacial TSTs.

The infill of early Permian erosional troughs in the Arckaringa Basin in south-central Australia consists of glaciomarine diamictites (Boorthanna Formation) succeeded by a marine mudstone-rich transgressive phase (Stuart Range Formation; Menpes et al., 2010; Menpez, 2013). An erosional surface, interpreted as a sequence boundary, cuts down into underlying glaciogenic sediments in the landward part of the trough but shows little evidence of erosion in the more seaward part. Seismic evidence indicates that the valleys were not completely filled with sediment when the sea transgressed, resulting in long and narrow bathymetrically-irregular fjords (Menpez, 2013). Abundant algal material (TOC values up to around 10%) recorded in the organic rich shale suggests density stratification of the water column with nutrient rich surface water and anoxic bottom water conditions. These conditions, also recorded in modern anoxia-influenced fjords (Skei, 1983; Syvitski and Shaw, 1995; Sohlenius et al., 2001), are associated with stagnant, brackish to restricted marine environments as evidenced by micro-paleontological assemblages.

The layers rich in *Tasmanites* cysts (named tasmanite beds) in the Woody Island Formation of the Tasmania Basin of southernmost Australia are laminated and occupy a consistent stratigraphic position ~ 10–30 m above the contact between the mudstone-prone postglacial

interval and the underlying glaciomarine lowermost Parmeener Super-group (Wynyard Formation) (Clarke, 1989; Domack et al., 1993; Hand 1993; Fielding et al., 2010). The glaciomarine section is overlain by the mudstone-rich Woody Island Formation deposited in an offshore or deep, quiet, poorly oxygenated fjordal setting (Fielding et al., 2010). The transition from glacial diamictite to siltstone is abrupt, suggesting rapid transgression and/or retreat of glaciers out of the fjords (Fielding et al., 2010). The Woody Island Formation siltstones have abundant *Tasmanites* cysts, in places accumulated as the thin and areally confined Tasmanite Oil Shale (Bacon et al., 2000). Domack et al. (1993) stated that the Tasmanites Oil Shale is a recurring facies, rather than a single bed. Pyrite nodules and dropstones are abundant in the tasmanite beds. TOC values for the tasmanite levels reach exceptionally up to 63%, with high (up to close to 1,000 mg HC/g TOC) hydrogen indexes (Revill et al., 1994; Bacon et al., 2000). In the central part of the basin conglomerates and sandstones in the mudstone facies associated with topographic highs triggered clastic dilution and the oil shale is absent and *Tasmanites* cysts are scarce to absent (Reid et al., 2003). Sulfur values, rarely exceeding 2%, suggesting that both marine sediments and fresh-water sediments occur, indicating conditions of variable salinity. The association of glendonite with the TOC-rich tasmanite levels suggests a marine origin and low temperatures (Shearman and Smith, 1985; Domack et al., 1993; Selleck et al., 2007). The organic-rich tasmanite beds in the Woody Island Formation and equivalent units represent condensed sections within a maximum flooding interval (Rogala et al., 2007).

7. Concluding remarks

The tripartite division of LPIA episodes that affected the southwestern margin of Gondwana has been refined in recent years to five shorter-lived discrete events (1. latest Devonian - earliest Tournaisian, 2. Tournaisian, 3. Visean, 4. Serpukhovian-early Bashkirian and 5. late Pennsylvanian-earliest Permian) with the help of new radiometric dates and biostratigraphic zonations. The event 1 (Latest Devonian - earliest Tournaisian) is present in basins adjacent to the Gondwanan margin (i. e., Rio Blanco Basin) and in intraplate settings (Paraná Basin). The event 2 (Tournaisian) appears to be confined to the Rio Blanco Basin. The event 3 (Visean) is present in the stratigraphic record of the Rio Blanco,

Calingasta-Uspallata, Tarija, Chaco-Paraná and Paraná basins. The event 4 (Serpukhovian-early Bashkirian) is the most widespread in all the basins of southwestern Gondwana and the latest glaciation recorded in the retroarc basins bordering the Gondwana margin (San Rafael, Calingasta-Uspallata, Río Blanco) and the Paganzo Basin. The glacial events are capped by postglacial transgressive deposits. This new chronostratigraphy of the LPIA for southwestern Gondwana is in concert with the long-term control of the supercontinent drift over the pole. It also confirms the persistence of glacial conditions in basins located farther from the Gondwanan margin, i.e., Chaco-Paraná Basin, VFB and Paraná Basin while postglacial conditions were evidenced by the presence of Moscovian coals in the Paganzo Basin and Bashkirian through latest Carboniferous (Gzhelian) shallow marine carbonates in the Madre de Dios Basin, Navidad-Arizaro and Calingasta-Uspallata basins.

In transitional to open marine settings the valley-glacier-retreat (fjord) facies association and the submarine-retreat facies association are proposed as the end members of a spectrum controlled by accommodation space (high in subsiding, open marine shelf areas or moderate to low along indented coastlines with valleys/fjords) and the spacing of points of sediment injection (linear or closely spaced source points or isolated source points confined to widely spaced valleys or fjords). A series of paleofjords (e.g., Talacasto, Las Lajas, Quebrada Grande), with dominantly northward flow, in the western Paganzo Basin exemplifies this facies association. In this sector of the basin the Protoprecordillera was a substantial topographic feature that, along with the Pie de Palo Arch and possibly the Famatina Arch, served as glacial centers.

The complex relationship between water depth during glaciation, position of the glacial front, sedimentation rates, eustasy, and isostatic adjustments has an important influence on the sequence stratigraphy of glacial successions. High sediment supply and a remarkable sensitivity of sealevel fluctuations are key aspects that influence significantly the sequence stratigraphy of glaciated margins.

Uplift from glacio-isostatic rebound as the thickness of the overlying ice sheet started to decrease is commonly reflected in the glacial stratigraphy of paleofjords. This decrease produced a glacio-isostatic rebound exceeding eustatic sea-level rise thereby leading to a fluvial-deltaic progradation during base-level fall, defining a configuration of forced regression. Concomitantly, steep walls on paleofjords may have triggered slope failures leading to deposition of wedges dominated by gravity flow deposits.

The proximal coastal region characterized by ice marginal deposition is contrasted with a distal deep region with open glaciomarine sedimentation characterized by the submarine-retreat facies association. This assemblage forms glaciomarine aprons with a general fining-upwards vertical stacking which reflect deposition mainly during the retreat of ice sheets, ice caps and alpine glaciers and successive deglaciation. The thickness of these deglaciation sequences is controlled by available accommodation space (the combined product of subsidence and sealevel changes), significant in basin centers and moderate to low along basin margins. Presence of glacial erosion surfaces (GSEs) of diverse hierarchy (scoured and polished surfaces on bedrock, inter- and intertill boulder pavements, striated soft sediment surfaces), mostly in the proximal domain, suggests different orders of glacial advance/retreat fluctuations.

A common tripartite motif, equivalent to second-order sequences with estimated durations of 10–80 Myr characterizes the late Paleozoic successions present along the southwestern margin of Gondwana. The lower section corresponds to glacial and glacially-influenced diamictites; the middle interval is initiated with widespread postglacial transgressions and, in some cases, punctuated by successive shorter-lived incursions (i.e., Paganzo and San Rafael basins). The lower and middle intervals correspond to the deglaciation sequences as described and identified in the Karoo and Paraná basins. Finally, the upper term includes coastal progradation, followed in some places by continentalization, which led in many sectors to aridization as evinced by eolian deposits. Examples of second-order sequences can be identified in

the thick late Paleozoic successions of the Paraná and Karoo basins and in the VFB. Thinner second order sequences can be identified in the Calingasta-Uspallata, Río Blanco, Paganzo and San Rafael basins. The basal sequence boundary is commonly an abrasion surface developed on bedrock.

We propose that the deglaciation sequences be assigned to third order sequences consisting, when present, of a thin lowstand system tract (LST) of subglacial deposits resting on glacially abraded surfaces followed by thick glaciomarine and glacially influenced sediments. These facies are part of a thick transgressive systems tract (TST) that culminates with marine shales that reflect interglacial or postglacial conditions during ice retreat. Thus, the deglaciation sequences are proposed to be third order sequences made up of LST-TST or exclusively TST.

Author statement

I attest that all authors contributed to the elaboration of this manuscript.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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