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Tectonic evolution of western Amazonia from the assembly of Rodinia to its break-up

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Plate tectonic reconstructions of the late Mesoproterozoic–Neoproterozoic supercontinent of Rodinia juxtapose the western margin of Amazonia against eastern Laurentia based on palaeomagnetic, isotopic, and geological evidence. Mesoproterozoic ('Grenvillian') orogenic belts are of crucial importance to these reconstructions as they act as key tectonic tracers for Amazonia–Laurentia interactions. They include orogenic belts sited on Amazonia (such as the Sunsás Orogen), para-autochthonous elements such as the Mesoproterozoic metamorphic basement inliers within the Andean Belt (e.g. in the Colombian Andes), exotic terranes accreted to Amazonia during Rodinia assembly (such as the Arequipa Massif), and orphaned fragments of Amazonian basement in Central and North America. A review of the timing of Sunsás orogenesis demonstrates that it occurred from about 1.2–1 Ga in eastern Bolivia and the western Amazon region of Brazil. This is significantly older than the timing of metamorphism in Mesoproterozoic basement inliers of the Colombian Andes, which record a late metamorphic event between 0.9 and 1.0 Ga. Orphaned fragments of Amazonian basement in Laurentia (such as the Blue Ridge/Mars Hill terrane) suggest collision between southeastern Laurentia and Amazonia at ca. 1.15 Ga. The Arequipa Massif (and Antofalla Basement) most likely represents an exotic basement terrane that was caught up in the collision of southeastern Laurentia with western Amazonia. Recent palaeomagnetic data suggest that Amazonia moved northeastwards along the eastern Laurentian margin during Grenvillian collision. Amazonia evidently collided with southern Laurentia at ca. 1200 Ma and, as a result of progressive dextral transcurrent movement, it encountered the Labrador–Greenland sector of eastern Laurentia (and possibly Baltica) by 980 Ma. The timing of the Rodinia break-up in western Amazonia is poorly constrained. Evidence exists supporting a Neoproterozoic western Amazonian active margin, which would imply its at least partial separation from the conjugate rift margin of eastern Laurentia (i.e. formation of a proto-Iapetus Ocean) prior to ca. 650 Ma. This rifting event may be linked to A-type magmatism at ca. 770–690 Ma which is documented in both southeast Laurentia and western Amazonia. Final separation must have been completed by Early Cambrian times based on the unequivocal evidence for drift-related sedimentation on the Laurentian margin of the Iapetus Ocean.

Keywords: Amazonia; Rodinia; supercontinent; geochronology; palaeomagnetism

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Introduction

Plate tectonic reconstructions of the late Mesoproterozoic–Neoproterozoic supercontinent of Rodinia conventionally juxtapose the western margin of Amazonia against eastern Laurentia and Baltica (e.g. Dalziel 1991; Hoffman 1991; Weil *et al*. 1998; Torsvik 2003; Li *et al*. 2008). The link between western Amazonia and eastern Laurentia is based on palaeomagnetic data (e.g. Weil *et al*. 1998; Tohver *et al*. 2002; D'Agrella-Filho *et al*. 2008), isotopic data (e.g. Tohver *et al*. 2004a), and geological evidence (e.g. Hoffman 1991; Sadowski and Bettencourt 1996). The geological evidence is based mainly on the correlation of truncated Mesoproterozoic orogenic belts. For example, the western Amazonia–eastern Laurentia link was first suggested in the early 1980s based on the presence of 1.2–1.3 Ga high-grade metamorphic rocks in the Colombian Andes and metamorphic rocks of similar age in the Grenville Province of eastern Canada (Alvarez and Cordani 1980; Kroonenberg 1982), while the Amazonia–Laurentia link in the reconstruction of Hoffman (1991) was based on the presence of Grenvillian belts in eastern Laurentia and in southwest Amazonia (the Rondônia–Sunsás Belt, Figure 1A).

This contribution reviews the tectonic evolution of western Amazonia from the assembly of Rodinia through to the break-up of the supercontinent. Further reviews on aspects of the evolution of western Amazonia are also summarized in Teixeira *et al*. (2010), Ramos (2010), Cardona *et al*. (2010), Cordani *et al*. (2009), Mišković *et al*. (2009), Ramos (2008), Fuck *et al*. (2008), and Cordani and Teixeira (2007). Intra-plate tectonic events of 'Grenvillian' age within the Amazonian Craton are reviewed by Cordani *et al*. (2010). This review commences with a summary of the various tectonic elements of western Amazonia, with particular emphasis placed on Mesoproterozoic orogenic belts, which act as key tectonic tracers for Amazonia–Laurentia interactions. These include orogenic belts sited on the Amazonian Craton (such as the Sunsás Orogen), para-autochthonous elements such as the Mesoproterozoic metamorphic basement inliers within the Andean Belt, exotic terranes accreted to Amazonia during Rodinia assembly (such as the Arequipa Massif), and orphaned fragments of Amazonian basement in Central and North America (such as Oaxaquia). This is followed by a review of the various tectonic and palaeogeographic models for Amazonia–Laurentia interaction during Rodinia assembly and subsequent break-up.

Tectonic elements of western Amazonia

The Amazonian Craton and the Sunsás Orogen

The Amazonian Craton forms the nucleus of South America. It consists of two Archaean crustal blocks and five Proterozoic tectonic provinces (Figure 1A). These are the Central Amazonian (>2.6 Ga), the Maroni-Itacaiunas (2.25–2.05 Ga), Ventuari-Tapajós (1.98–1.81 Ga), Rio Negro-Jurena (1.78–1.55 Ga), Rondônian-San Ignacio (1.55–1.30 Ga), and Sunsás (1.28–0.95 Ga) provinces (Tassinari *et al*. 2000; Cordani and Teixeira 2007; Cordani *et al*. 2009; Bettencourt *et al*. 2010). Overall, the pattern of crustal growth implies a series of progressive crustal accretion events which result in a series of domains that young away from the old Amazonian core. As the Sunsás Province on the southwest Amazonian Craton (Figure 1A) is broadly of Grenvillian age, it forms one of the key western Amazonia–eastern Laurentia connections (e.g. Dalziel 1991; Hoffman 1991; Li *et al*. 2008). However, tectonic models, which are reviewed later, juxtapose the Sunsás Belt and southwest Amazonia against different regions of eastern Laurentia: the southwestern Llano segment (Tohver *et al*. 2002), the Grenville Province of eastern Canada

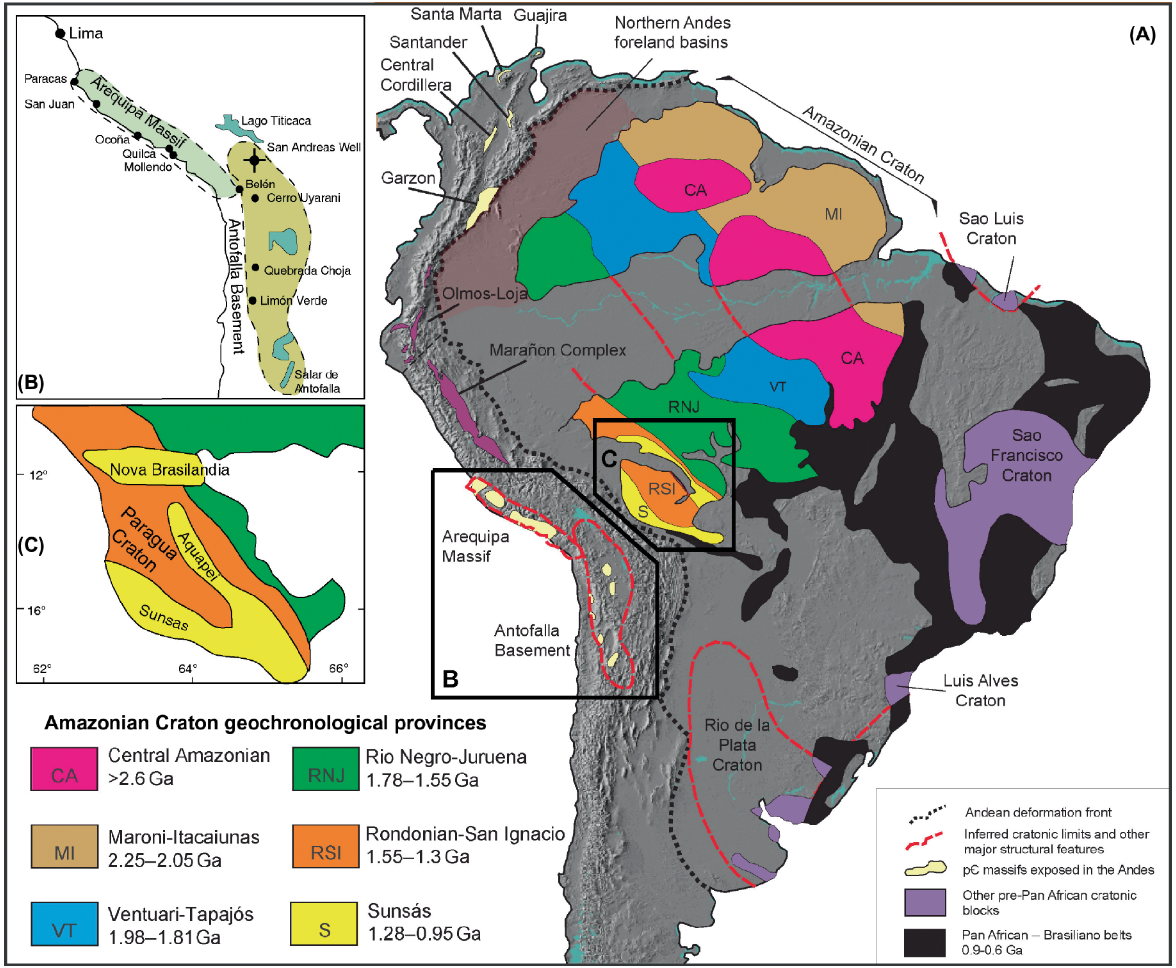

Figure 1. (A) Regional geological map of South America displaying the geochronological provinces of the Amazonian Craton and Andean basement inliers. Modified from Cordani *et al*. (2009) and Cordani *et al*. (2000). (B) Main basement inliers of the Central Andes (modified from Ramos 2008). (C) Map of the geology of southwest Amazonia from Elming *et al*. (2009).

(Sadowski and Bettencourt 1996), and the central and southern Appalachians (Loewy *et al*. 2003). An alternative model (Santos *et al*. 2008) suggests the two belts may not have been directly juxtaposed, but instead, that one may have been the extension of the other during the Mesoproterozoic.

Three Mesoproterozoic metamorphic belts of Grenvillian age are exposed on the southwest Amazonian Craton (Figure 1C). The Sunsás and Aguapeí belts are exposed in eastern Bolivia and along the Bolivia–Brazil border although different reconstructions have been proposed (Boger *et al*. 2005; Teixeira *et al*. 2010), while the Nova Brasilândia Belt forms the southern limit of the Rondônian province in southwest Brazil (Figure 1C). The Sunsás Belt comprises low- to medium-grade metamorphics of the Sunsás and Vibosi Groups. These groups comprise quartzitic and conglomeratic units that are believed to represent passive margin sedimentary sequences (Litherland *et al*. 1986; Sadowski and Bettencourt 1996) that were subsequently deformed during late (ca. 1 Ga) Sunsás orogenesis. The Sunsás and Vibosi Groups have been stratigraphically correlated with the Aguapeí Group of Brazil (Litherland *et al*. 1989). Deformation in the Sunsás Belt comprises a series of curvilinear, strike-slip shear zones. These mylonite zones are intruded by syn- to late-tectonic granitic suites, which were followed by the emplacement of post-tectonic plutons (Litherland *et al*. 1986, 1989). Boger *et al*. (2005) have constrained the main phase of Sunsás orogenesis at 1076 ± 18 Ma by a weighted mean $^{207}Pb/^{206}Pb$ SHRIMP zircon age on the undeformed post-tectonic Taperas granite in Bolivia. More recently, Teixeira *et al*. (2010), in a comprehensive review of the tectonic evolution of the Sunsás Belt, highlighted additional constraints on the syn- to post-tectonic Sunsás intrusive event. These include a U–Pb zircon age of 1105 ± 21 Ma from the slightly foliated Santa Teresa tonalite [pers. comm. of Matos quoted in Teixeira *et al*. (2010)], a U–Pb zircon age of 1014 ± 6 Ma for an orthogneiss in the Don Mario mineral district (Isla-Moreno 2009), and U–Pb LA-ICPMS zircon ages of 1092–1047 Ma for the Naranjito, Primavera, Taperas, and El Carmen plutons (Vargas-Mattos *et al*. 2009).

The Nova Brasilândia and Aguapeí belts (Figure 1C) represent rift basins that were affected by later transpression, crustal shortening, and granitic injections associated with Sunsás orogenesis (Cordani and Teixeira 2007). The Aguapeí Belt records only limited deformation at ca. 950 Ma with local development of greenschist-facies mylonites and minor S-type granite magmatism (Geraldes *et al*. 2001; Tohver *et al*. 2004b). K–Ar ages from the Aguapeí Belt range from 960 to 920 Ma (Geraldes *et al*. 1997). The metamorphic grade in the Nova Brasilândia Belt is significantly higher than that in the Aguapeí Belt, attaining amphibolite- to granulite-facies conditions. The syn-tectonic Rio Branco granite suite in the Nova Brasilândia Belt has yielded a U–Pb TIMS zircon age of 1098 ± 10 Ma and cogenetic gabbros have been dated by U–Pb zircon at 1110 ± 15 Ma (Rizzotto 1999). Deformed leucocratic melts injected along bedding planes in pelitic rocks have been dated by the U–Pb SHRIMP zircon method at 1122 ± 12 Ma, which may constrain metamorphism (Santos *et al*. 2000). Cooling from peak metamorphic conditions is recorded by U–Pb monazite ages of 1090 Ma, U–Pb titanite ages of ca. 1060 Ma, and ^{40}Ar–^{39}Ar hornblende and biotite ages of 970 Ma and 910 Ma respectively (Tohver *et al*. 2004b; 2006). Together with the age constraints quoted in Teixeira *et al*. (2010), these data suggest that *peak* Sunsás orogenesis occurred from about 1120 to 1000 Ma in eastern Bolivia and western Brazil. The onset of Sunsás orogenesis is less well constrained, but a maximum age can be derived from the age of cover sediments in the Sunsás, Aguapeí, and Nova Brasilândia belts. The maximum sedimentation age in the Nova Brasilândia Belt is inferred from the youngest U–Pb SHRIMP detrital zircons ages of 1215 ± 20 Ma (Rizzotto *et al.* 1999) and 1231 ± 14 Ma (Santos *et al*. 2000). The deposition age of the basal Huanchaca Group (a correlative of the Aguapeí cover sequence) is well constrained by Santos *et al.* (2005) and D'Agrella-Filho *et al*. (2008) to between 1167 ± 27 Ma (the youngest U–Pb detrital zircon age) and 1149 ± 7 Ma (a U–Pb xenotime age interpreted as dating diagenesis). Together these data suggest Sunsás orogenesis may have initiated as early as 1200 Ma.

Recent studies (Tohver *et al*. 2004b; Boger *et al*. 2005) consider an allochthonous evolution for the Sunsás Belt. They consider the Sunsás Belt to form the southern boundary of the Paragua Craton of Litherland *et al*. (1989) which represents an accreted micro-continental terrane. In this tectonic scenario, the deep-water metasedimentary rocks of the Nova Brasilândia Belt represent the southern margin of Amazonia prior to the Sunsás-age collision of the Paragua Craton. Santos *et al*. (2008) argues that exposed pre-Sunsás basement rocks in the Paragua Craton, along with inherited U–Pb zircon ages in plutonic rocks, can be correlated with the neighbouring Rio Negro-Jurena and Rondônian-San Ignacio provinces to the north. Coupled with the absence of fragments of older, Archean and Palaeoproterozoic Amazonian crust, this suggests that the Sunsás Belt is autochthonous to the Amazonian Craton (Santos *et al*. 2008).

The Arequipa Massif and the Antofalla Basement

The Arequipa Massif and the neighbouring Antofalla Basement (Figure 1B) on the Pacific Coast of the Central Andes were originally considered an integral part of the South American continent (Cobbing *et al*. 1977). Whereas the Amazonian Craton exhibits a simple pattern of crustal growth with an Archaean and Palaeoproterozoic core with progressively younger domains encountered towards the southwest (Figure 1A), the Arequipa Massif disrupts this polycyclic framework (Coira *et al.* 1982) as it has yielded components as old as Palaeoproterozoic (1.79–2.02 Ga) in age (Loewy *et al*. 2004). Although a para-autochthonous origin for the Arequipa Massif (e.g. Tosdal 1996) has been postulated, the anomalous position and crustal growth pattern of the Arequipa Massif and the Antofalla Basement has led most authors to propose that they are allochthonous to Amazonia (e.g. Ramos 1988; Dalziel 1994; Loewy *et al*. 2004).

The Arequipa Massif has a complex magmatic and metamorphic evolution which extends from early Proterozoic to early Palaeozoic times (Ramos 2008; Casquet *et al*. 2010). Basement outcrops stretch along the Pacific coast for 800 km and extend inland for approximately 100 km (Figure 1B). U–Pb zircon geochronology study demonstrates that the Arequipa Massif in southern Peru (in the regions of San Juan, Ocoña, Quilca, and Mollendo) and western Bolivia contains juvenile Palaeoproterozoic 2.02–1.79 Ga intrusions that were metamorphosed at 1.82–1.79 Ga and 1.05–0.93 Ga (Loewy *et al*. 2004). This younger Mesoproterozoic metamorphic event is constrained by a U–Pb zircon metamorphic age of 970 ± 23 Ma at Mollendo and an age of 1198 $^{+6}/_{-4}$ Ma at Quilca (Wasteneys *et al*. 1995). Martignole and Martelat (2003) document the presence of ultrahigh-temperature (UHT) conditions in migmatites, aluminous gneisses, mafic granulites, and granites of the Mollendo-Camana region. The breakdown of sillimanite-orthopyroxene to sapphirine-orthopyroxene-cordierite assemblages in migmatites is attributed to decompression from 1.1–1.2 to 0.8–0.9 GPa at temperatures in excess of 950°C. *In situ* Th–U–Pb microprobe chemical age determinations on monazite yield ages of ca. 1.0 Ga. The UHT metamorphism is of regional extent and has been attributed to hot asthenospheric mantle impinging on the base of over-thickened crust (Martignole and Martelat 2003). Casquet *et al.* (2010) present U–Pb SHRIMP zircon data from the Arequipa Massif in southwest Peru which demonstrate an early magmatic event at 1.89–2.1 Ga, UHT metamorphism at ca. 1.87 Ga, and late felsic magmatism at ca. 1.79 Ga. Grenville-age UHT metamorphism reworked the Palaeoproterozoic rocks and was diachronous: ca. 1040 Ma in the Quilca, Camaná, and San Juán Marcona regions, 940 ± 6 Ma in the Mollendo area, and between 1000 and 850 Ma in the Atico domain (Casquet *et al.* 2010).

The Antofalla Basement in northern Chile was originally considered to represent a single coherent basement block contiguous with the Arequipa Massif (Ramos 1988), but more recent studies suggest that the Antofalla Basement is a discrete entity (Ramos 2008). The southern portion of the Antofalla Basement from Limón Verde in northern Chile to Antofalla in the western Argentine Puna comprises juvenile material with Nd T_{DM} ages as young as 0.7–0.6 Ga. Magmatism and metamorphism are constrained to between 0.5 and 0.4 Ga (Loewy *et al*. 2004). The northernmost portions of the Antofalla Basement in Chile comprised foliated amphibolite and gneiss in the Belén region, felsic granulite, charnockite, and amphibolite near Cerro Uyarani (Wörner *et al*. 2000), and migmatitic quartz-biotite paragneiss and granodioritic orthogneiss in Quebrada Choja (Loewy *et al*. 2004). U–Pb TIMS zircon upper intercept ages from all three regions range from 1.70 to 2.02 Ga and demonstrate the presence of Palaeoproterozoic protoliths (Wörner *et al*. 2000; Loewy *et al*. 2004). Mesoproterozoic granulite-facies metamorphism in Cerro Uyarani is constrained by a U–Pb zircon lower intercept age of ca. 1.15 Ga, a 1008 ± 16 Ma Sm–Nd mineral isochron, and a 983 ± 2 Ma Ar–Ar hornblende age (Wörner *et al*. 2000). Calc-alkaline igneous rocks (granite, tonalite) in the Quebrada Choja yield U–Pb zircon ages of 1.03–1.07 Ga and are cut by dacitic dikes dated at 635 ± 5 Ma (Loewy *et al*. 2004). The data from Belén, Cerro Uyarani, and Quebrada Choja demonstrate the presence of Palaeoproterozoic protoliths with ages similar to the Arequipa Massif. Subsequent calc-alkaline magmatism and high-grade metamorphism yield Mesoproterozoic ('Grenvillian') ages which are also similar to those obtained from the Arequipa Massif (Wasteneys *et al*. 1995; Martignole and Martelat 2003).

Ramos (1988) considered that the Arequipa Massif and Antofalla Basement accreted to Amazonia during the early Palaeozoic Pampean Orogeny. Dalziel (1994) envisaged that the two basement blocks were transferred to Amazonia from the northeast corner of Laurentia during fragmentation of Rodinia. Loewy *et al*. (2003) refuted this correlation using whole-rock Pb isotopes and U–Pb geochronology, and suggested instead that the Arequipa Massif and Antofalla Basement represent a fragment of the Kalahari Craton caught up in Laurentia–western Amazonia collision at ca. 1.0 Ga. Chew *et al*. (2007a) presented U–Pb detrital zircon analyses from the Chiquerío Formation, a late Neoproterozoic glacial diamictite which unconformably overlies the Arequipa Massif in southern Peru. These data suggest the Chiquerío Formation was autochthonous with respect to the Amazonian Craton. This constrains the docking history of the underlying Arequipa Massif to late Neoproterozoic or older in age, probably colliding during the 1.2–1 Ga Sunsás Orogeny (Chew *et al*. 2007a), as first postulated by Loewy *et al*. (2004).

Mesoproterozoic basement inliers in the Andean Belt

Mesoproterozoic basement is encountered in several inliers in the Andean cordilleras of Colombia, Ecuador, and Peru and is also represented by a major detrital or inherited component within Neoproterozoic to Palaeozoic sedimentary and magmatic rocks. Alvarez and Cordani (1980) and Kroonenberg (1982) first considered this belt to represent the remnants of a continental collision of Grenvillian age between Laurentia and Amazonia.

Mesoproterozoic high-grade metamorphic rocks occur along the eastern segment of the Colombian Andes and in several massifs in the Caribbean region (Figure 1A). Recent reviews of the Mesoproterozoic geological evolution of these inliers are summarized in Cardona *et al*. (2010), Ordoñez-Carmona *et al.* (2006) and Cordani *et al*. (2005). Ordoñez-Carmona *et al*. (2006) subdivided the Mesoproterozoic basement inliers into autochthonous units such as the Garzón Massif, and para-autochthonous units such as the Santa Marta Massif, the Guajira Peninsula, the Santander Massif, and the Mesoproterozoic inliers on the eastern flank of the Central Cordillera (Toussaint 1993).

The Garzón Massif is the best exposed portion of Mesoproterozoic basement in the Colombian Andes. It is believed to be autochthonous to the northwestern Amazonian Craton based on its relationship with its unmetamorphosed Cambrian cover (Toussaint 1993; Ordoñez-Carmona *et al*. 2006). It is divided into three main lithological units (Jiménez-Mejía *et al*. 2006 and references therein): the Guapotón-Mancagua Gneiss in the west, and the Vergel granulites and Las Margaritas migmatites of the Garzon Complex in the east. The Guapotón Gneiss comprises predominantly hornblende-biotite augen gneisses whose magmatic precursor is constrained by a U–Pb TIMS zircon upper intercept age of 1088 ± 6 Ma age (Restrepo-Pace *et al*. 1997) and a U–Pb SHRIMP zircon age of 1158 ± 22 Ma (Cordani *et al*. 2005). Younger rim overgrowths reflect high-grade metamorphism and are dated at 1000 ± 25 Ma by the U–Pb SHRIMP zircon method (Cordani *et al*. 2005). A ^{40}Ar–^{39}Ar hornblende age of 911 ± 2 Ma records post-orogenic cooling (Restrepo-Pace *et al*. 1997).

The Vergel granulites (which consist of charnockite, mafic granulite, quartzo-feldspathic gneiss, and amphibolite) and the Las Margaritas migmatites (biotite-garnet sillimanite gneisses with a metasedimentary protolith) represent a metamorphosed volcano-sedimentary sequence (Kroonenberg 1982). *P–T* determinations for the Vergel granulites define a counterclockwise *P–T* path (5.3–7.2 kbar, 700–780°C) and a clockwise *P–T* path for Las Margaritas migmatites (6.3–9.0 kbar, 680–820°C) (Jiménez-Mejía *et al*. 2006). Based on the geochronological study of Cordani *et al*. (2005), leucosome generation in Las Margaritas migmatites is constrained by a U–Pb SHRIMP zircon age of 1015 ± 8 Ma, and Sm–Nd garnet–whole-rock isochrons from the same unit yield ages of 1034 ± 6 Ma and 990 ± 8 Ma. ^{40}Ar–^{39}Ar dating of five biotite and two amphibole samples yields ages from 950 to 1000 Ma, with one outlier at ca. 1030 Ma. U–Pb SHRIMP zircon data from the Vergel granulites suggest crystallization of the igneous protolith at ca. 1100 Ma and subsequent metamorphism at ca. 1000 Ma (Cordani *et al*. 2005). Sm–Nd garnet–whole-rock isochrons yield ages of 935 ± 5 Ma and 925 ± 7 Ma. Five biotites and one hornblende from the Vergel granulites yield ^{40}Ar–^{39}Ar plateau ages between 980 Ma and 860 Ma (Cordani *et al*. 2005).

All other Mesoproterozoic basement inliers in Colombia (the Santander Massif, the Santa Marta Massif, the Guajira Peninsula, and the basement inliers on the eastern flank of the Central Cordillera) are believed to represent a single, dismembered terrane. Recent palaeomagnetic data from their Mesozoic cover suggest that these basement inliers were displaced from southern latitudes (similar to present-day northern Peru and Ecuador) with most of the displacement probably occurring before the Early Cretaceous (Bayona *et al*. 2006). Cardona *et al*. (2010) thus favour a para-autochthonous origin for these basement units with respect to the Amazonian Craton. The major crystalline basement unit in the Santander Massif (Figure 1A) is the Bucaramanga Gneiss which comprises a metasedimentary sequence of pelitic paragneisses, with intercalations of marble, calc-silicate rock, and minor amphibolite. This unit has yielded a ^{207}Pb/^{206}Pb SHRIMP zircon age of 864 ± 66 Ma attributed to a late metamorphic event (Cordani *et al*. 2005). ^{40}Ar–^{39}Ar dating of feldspar yields Early Jurassic ages (ca. 200–175 Ma) due to a Mesozoic thermal event of regional extent, while ^{40}Ar–^{39}Ar dating of hornblende and biotite yields spectra consistent with cooling at ca. 900 Ma and a subsequent thermal overprint at 200 Ma

(Restrepo-Pace *et al*. 1997). The Dibulla Gneiss, a sequence of amphibolites and quartzo-feldspathic rocks, is exposed in the northeastern portion of the Santa Marta Massif. *P–T* conditions in the Dibulla Gneiss are 6.0–8.0 kbar and 760–810°C, and geochemical data from amphibolites imply a rift-related origin (Cardona-Molina *et al*. 2006). Based on U–Pb SHRIMP zircon data from a felsic Dibulla Gneiss sample, Cordani *et al*. (2005) attribute ages of ca. 1370 Ma to crystallization of the magmatic protolith and zircon ages of ca. 1140 Ma and ca. 1000 Ma to two subsequent metamorphic events. The younger ca. 1 Ga event is supported by a Sm–Nd garnet isochron from a granulite which yields an age of 971 ± 8 Ma (Ordoñez-Carmona *et al*. 2006). Other regions which may have undergone Mesoproterozoic tectonism in the northern Andes include the eastern segment of the Central Cordillera of Colombia and the adjacent northern San Lucas Serranía (Toussaint 1993). Mesoproterozoic basement has not yet been documented within the Venezuelan Andes, but the presence of inherited zircons within Neoproterozoic–early Palaeozoic granitoids argues for the assimilation of older (ca. 1.0 Ga) crust (Burkley 1976). Goldstein *et al*. (1997) report U–Pb SHRIMP zircon ages from modern-day river sand of the lower Orinoco River, Venezuela. The zircon ages range from 2.83 to 0.15 Ga, with discrete age groupings occurring at ca. 2.8, ca. 2.1, and ca. 1.1 Ga. The 1.1 Ga population makes up about a quarter of the total number of analyses and is likely derived from basement of Grenvillian age in the northern Andes.

Further south in the Andean foreland of Ecuador, several oil wells have sampled late Mesoproterozoic basement, although the amount of published information is limited (Litherland *et al*. 1994). In Peru, in addition to the previously discussed Arequipa Massif, Dalmayrac *et al*. (1988) report a U–Pb zircon multigrain TIMS lower intercept age of ca. 1160 Ma from granulite-facies rocks in the Amazon Basin within the Rio Picharí region east of Cuzco (Figure 1A). Metasedimentary basement crops out extensively in the Eastern Cordillera of Peru, where it is termed the Marañon Complex. Recent studies concerning the tectonic evolution of the Marañon Complex have shown that it consists of different units of Palaeozoic age and that it contains a major Mesoproterozoic (ca. 1.3–0.9 Ga) detrital zircon component (Cardona *et al.* 2007, 2009; Chew *et al*. 2007b, 2008; Mišković *et al*. 2009). These same studies also recognize a major inherited Mesoproterozoic (ca. 1.3–0.9 Ga) zircon component in magmatic rocks in the Eastern Cordillera of Peru. The study of Mišković *et al*. (2009) also demonstrates the presence of Mesoproterozoic basement in the Eastern Cordillera. They comprise partially foliated and compositionally diverse granitoids from the south–central cordilleran segment which yield U–Pb zircon crystallization ages of 985 ± 14 Ma, 1071 ± 23 Ma and 1123 ± 23 Ma (Mišković *et al*. 2009).

The tectonic history of the Mesoproterozoic basement inliers in the northern Andes clearly demonstrates the presence of a dismembered orogenic belt of ‘Grenvillian’ age. Although Phanerozoic tectonics may have redistributed some of the basement terranes, particularly in Colombia, they are still viewed as para-autochthonous domains that have remained in proximity to the margin of Amazonia (Cardona *et al*. 2010). This orogenic belt is thus likely to be continuous under the northern Andes (e.g. Litherland *et al*. 1985; Restrepo-Pace *et al*. 1997) and thus would have acted as a major sediment source for younger Palaeozoic sequences in the northern Andes, which contain abundant Mesoproterozoic detrital zircon. The metasedimentary and magmatic rocks of the Mesoproterozoic basement inliers of the Colombian Andes most likely formed on the margins of the Amazonian Craton based on their detrital/inherited zircon signature, Pb isotope signatures, and Nd model ages of 1.6–1.9 Ga (Restrepo-Pace *et al*. 1997; Ruiz *et al*. 1999; Cordani *et al*. 2005; Ordoñez-Carmona *et al*. 2006). The magmatic rocks in the Garzón and Santa Marta massifs probably represent subduction-related magmatism on an active margin. In the Garzón Massif, arc activity lasted until ca. 1.1 Ma (Cordani *et al*. 2005; Cardona-Molina *et al*. 2006). Additionally, all Colombian Mesoproterozoic basement inliers seem to record a common late metamorphic event between 0.9 and 1.0 Ga, which is significantly younger than Sunsás orogenesis (1.2–1 Ga) in eastern Bolivia and western Brazil. The relationship of this Mesoproterozoic orogenic belt on northwestern Amazonia to both the Sunsás Orogen in southwestern Amazonia and orphaned fragments of Amazonian basement in Central and North America is discussed later.

Orphaned fragments of Amazonian basement in Central and North America

Plate tectonic reconstructions of Rodinia usually juxtapose the western margin of Amazonia against southeastern Laurentia. During the subsequent break-up of Rodinia, the locus of rifting would be expected to approximate the collisional suture between the continents. If the axis of rifting was to deviate from the trace of the collisional suture, this could have resulted in fragments of the Amazonian Craton remaining attached to eastern Laurentia following Rodinia break-up.

Ruiz *et al*. (1999) demonstrate that the whole-rock Pb isotope compositions of Mesoproterozoic basement rocks in southern México are very similar to those of Mesoproterozoic basement rocks of the Santa Marta and Garzón massifs in the Colombian Andes, but distinct from those of Grenvillian rocks in eastern Laurentia. These southern México basement rocks, the Oaxaca and Guichicovi complexes, have been inferred to represent a single Mesoproterozoic terrane, 'Oaxaquia' (Ortega-Gutiérrez *et al*. 1995). The Pb isotopic data of Ruiz *et al*. (1999) demonstrate that Oaxaquia must consist of several discrete blocks. They suggest that the Mesoproterozoic basement rocks of southern México were likely transferred from their original position along the margin of Amazonia (probably from Colombia) during collisional orogenesis in the Palaeozoic or earlier.

Tohver *et al*. (2004a) also use whole-rock Pb isotope data to test the provenance of different blocks within western Amazonia–eastern Laurentia. They demonstrate that the southwestern Amazonia Craton shares a common Pb isotope composition (an elevated U/Pb ratio) and a similar magmatic history (based on U–Pb zircon crystallization ages) to a Mesoproterozoic basement inlier in the southern Appalachians, the Blue Ridge/Mars Hill terrane. This is in contrast to the Grenville Province of eastern Laurentia, which is characterized by a source region with a distinctly lower U/Pb ratio. Tohver *et al*. (2004a) propose that the Blue Ridge/Mars Hill terrane transferred from southwest Amazonia to southeast Laurentia during Grenvillian orogenesis after ca. 1.15 Ga.

Orphaned fragments of Amazonian basement in Central and North America such as Oaxaquia and the Blue Ridge/Mars Hill terrane serve as important tectonic tracers for Amazonia–Laurentia interaction. Using this information, the various tectonic and palaeogeographic models for Amazonia–Laurentia interaction during Rodinia assembly and subsequent break-up are now assessed. This synopsis also takes into account the previously described tectonic histories of the Mesoproterozoic basement domains on western Amazonia, along with recent constraints from palaeomagnetic data.

Discussion – tectonic models for western Amazonia during Rodinia assembly

Although there is a general consensus for the eastern Laurentia–western Amazonia connection within Rodinia, it should be noted that both these margins are more than 3000 km long. Hence identifying a collisional 'piercing point' common to both continental margins is key to any palaeogeographic reconstruction. This discussion commences with a brief synopsis of our understanding of the Grenville Province which formed the southeastern limit of Proterozoic Laurentia. Subduction–accretion and arc formation took place along southeastern Laurentia for >400 Ma from the late Palaeoproterozoic to the late Mesoproterozoic (Rivers 1997). Arc magmatism was accompanied by accretionary orogenesis during the Labradorian (ca. 1680–1660 Ma), Pinwarian (ca. 1500–1450 Ma), and Elzevirian (ca. 1250–1190 Ma) orogenies, with major crustal units tending to young towards the southeast of the Grenville Province. Rivers (1997) considers the continent–continent Grenvillian Orogeny took place between ca. 1190 and 980 Ma. Gower *et al*. (2008) define the Grenvillian Orogeny in eastern Laurentia to have occurred between 1085 and 985 Ma with events before ca. 1085 Ma considered as pre-Grenvillian.

The majority of the earlier reconstructions of Rodinia (e.g. Hoffman 1991; Weil *et al*. 1998; Torsvik 2003) in general place the Peruvian sector of western Amazonia against the central Grenville Province of Ontario and New York on eastern Laurentia. The reconstruction of Dalziel (1991) adopts a slightly different configuration, placing the Peruvian segment further north against the Labrador–Greenland–Scotland Promontory (LGSP) of Laurentia. The reconstruction of Dalziel (1994) places Amazonia even further north, juxtaposing the Arica embayment (Figure 1A) against the LGSP. This reconstruction was refuted by the study of Loewy *et al*. (2003), which demonstrated that the difference in Pb isotopic signatures between the LGSP basement and the Arequipa Massif–Antofalla Basement (which forms the crustal basement to the Arica embayment) precludes that particular palaeogeography. Instead they propose a Grenvillian connection between southern Laurentia (Llano) and Kalahari, and that Amazonia collided with a contiguous southeastern Laurentia/Kalahari margin at ca. 1.0 Ga.

This Amazonia–southern Appalachians linkage features prominently in recent constructions. Tohver *et al*. (2002) paired western Amazonia with southernmost Laurentia (the Llano segment in Figure 2) based on a low-latitude ca. 1200 Ma palaeomagnetic pole obtained from gabbroic sills and basalts of the Nova Floresta Formation of Rondônian–San Ignacio Province in southwest Amazonia. This represents the first palaeomagnetic pole for Amazonia in the 1200–600 Ma Rodinia time period. The Pb isotope study of Tohver *et al*. (2004a) demonstrated the Blue Ridge/Mars Hill terrane in southeast Laurentia represents an orphaned fragment of southwest Amazonian basement that was transferred during Grenvillian orogenesis after ca. 1.15 Ga. The presence of these Amazonian rocks in southeast Laurentia records the northward passage of the Amazon Craton along the Laurentian margin, following the original collision with southernmost Laurentia at ca. 1.2 Ga (Figure 2). This northward passage of Amazonia may have taken place along the whole Grenvillian margin of Laurentia (Tohver *et al*. 2004a) (Figure 2).

More recent palaeomagnetic data for the Amazonian Craton supports the tectonic scenario of Tohver *et al*. (2004a). D'Agrella-Filho *et al*. (2008) report a palaeomagnetic pole from an interstitial hematite cement from red beds of the Fortuna Formation in the Aguapeí Group of the southwest Amazonian Craton. The age of diagenesis of the red beds is well constrained by a 1149 ± 7 Ma U–Pb SHRIMP age of authigenic xenotime rims on detrital zircons. This palaeomagnetic pole fixes the palaeogeographic position of the Amazon Craton near the southeast Appalachians portion of North America at 1.15 Ga. Elming *et al*. (2009) obtained a palaeomagnetic pole for a basic sill in the Aguapeí Group which yielded a 981 ± 2 Ma ^{40}Ar–^{39}Ar plateau age. This palaeomagnetic pole places NW Amazonia against the LGSP of Laurentia and Baltica at 981 Ma. Elming *et al*. (2009) favour a scenario where Amazonia moved northeastwards along eastern Laurentia from ca. 1200 to 980 Ma, although the orientation of Amazonia at 980 Ma differs slightly from other recent reconstructions of Rodinia (e.g. Li *et al*. 2008), with northern Amazonia juxtaposed against the LGSP of Laurentia and Baltica. The palaeomagnetic data of Elming *et al*.

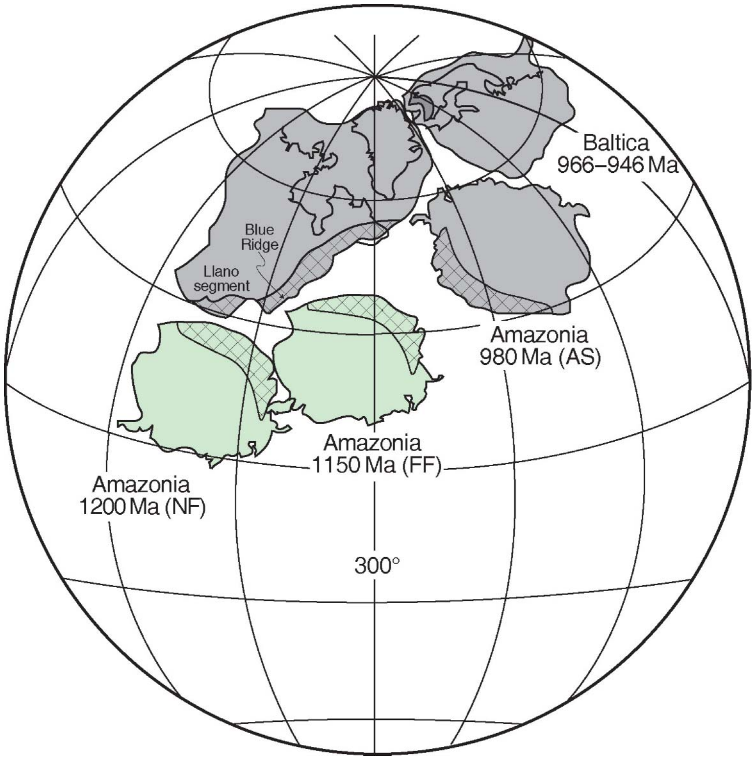

Figure 2. Tectonic reconstruction of the Amazon and Baltica cratons relative to Laurentia, with Laurentia shown in present coordinates. In the reconstruction at 980 Ma, the position of Amazonia is from Elming (2009). The position of Amazonia at 1150 Ma is from D'Agrella-Filho *et al*. (2008) and at 1200 Ma is from Tohver *et al*. (2002).

(2009), in particular the juxtaposition of NW Amazonia with Baltica, support the palaeogeographic reconstruction of Cardona *et al*. (2010). This model correlates the Colombian Mesoproterozoic basement inliers with the Oaxaquia terrane in Mexico based on the presence of a common late metamorphic event between 0.9 and 1.0 Ga. This late metamorphic event is significantly younger than Sunsás orogenesis (1.2–1 Ga) in eastern Bolivia and the western Amazon region of Brazil. Cardona *et al*. (2010) therefore favour interaction of the NW Amazonian Craton and Oaxaquia with the Sveconorwegian Province of Baltica while Sunsás (southwest Amazonia) was interacting with eastern Laurentia (Figure 2). The palaeogeographic reconstructions of Elming *et al*. (2009) and Cardona *et al*. (2010) are also supported by recent studies in eastern Laurentia. Gower *et al*. (2008) suggest that Grenvillian orogenesis in eastern Labrador involved a lateral-ramp regime accommodated by dextral strike-slip, in contrast to frontal-thrust ramp tectonics further west in central and western Labrador and adjacent Quebec. This lateral-ramp regime is interpreted as an indentor corner, possibly produced by the collision of northern Amazonia with eastern Laurentia (Gower *et al*. 2008 and figure 6 therein).

Discussion – tectonic models for western Amazonia during Rodinia break-up

The break-up history of the western Amazonian margin (which formed part of the larger western Gondwana margin during Rodinia break-up) is poorly understood compared to its conjugate margin of eastern Laurentia. On eastern Laurentia, two pulses of rifting activity are recognized. The earlier pulse (from ca. 760 to 680 Ma) is associated with A-type granite magmatism in the Blue Ridge of the southern Appalachians (e.g. Aleinikoff *et al*. 1995; Tollo *et al*. 2004). The later pulse ranges from 620 to 550 Ma (e.g. Kamo *et al*. 1995; Cawood *et al*. 2001) and is conventionally regarded to have resulted in the separation of eastern Laurentia and cratonic elements of west Gondwana and Baltica (e.g. Hoffman 1991; Bingen *et al*. 1998; Cawood *et al*. 2001; Li *et al*. 2008), followed by continued rifting along Laurentia's Iapetan margin, which generated a series of terranes at 540–535 Ma followed by Early Cambrian drift-related sedimentation (Williams and Hiscott 1987).

The magmatic record and the quantity of palaeomagnetic data for the western Amazonia margin during the late Neoproterozoic are restricted compared to eastern Laurentia. There is little evidence on the western Gondwanan margin for magmatism at this time, although this is at least partly due to the restricted amount of Precambrian basement exposed within the Andean Belt. A-type granitic plutonism has been documented (775–690 Ma) on western Gondwana in the Eastern Cordillera of Peru (Mišković *et al*. 2009) and in the Grenvillian basement of the Precordillera terrane (Baldo *et al*. 2006). Juvenile extensional magmatism (dacite dikes) has been dated at 635 ± 4 Ma in the Antofalla Basement of northern Chile (Loewy *et al*. 2004) while late Neoproterozoic extension-related volcanism (ultrapotassic mafic dikes and HFSE-enriched alkaline lava flows) related to Rodinia break-up has also been identified in the Puncoviscana Basin of northwestern Argentina (Omarini *et al.* 1999). Late Neoproterozoic extension in the Antofalla Basement and in the Puncoviscana fold belt probably resulted in the partial detachment of the Antofalla Basement, which was subsequently re-accreted during the early Palaeozoic Orogeny (Loewy *et al*. 2004; Ramos 2008; Mišković *et al*. 2009). By at least 530 Ma, segments of the western Gondwana margin were clearly a destructive margin (the onset of subduction-related plutonism in the Sierra Pampeanas, Rapela *et al*. 1998).

Chew *et al*. (2008) examined the U–Pb detrital zircon 'fingerprint' of autochthonous Neoproterozoic to early Phanerozoic sedimentary sequences from the northern and central segments of the western Amazonian margin. Detrital zircon populations are consistent with derivation from the western Amazonian margin, with the exception of a major peak at 550–650 Ma, which was attributed to a late Neoproterozoic magmatic arc buried beneath the modern Andean foreland. This belt may be contiguous with the Brasiliano-age Tucavaca Belt in Bolivia (Litherland *et al*. 1986). A Neoproterozoic active margin on western Amazonia would imply at least partial separation of west Gondwana from its conjugate rift margin of eastern Laurentia prior to ca. 650 Ma.

Separation of western Amazonia and eastern Laurentia may be linked to the ca. 770–690 Ma A-type magmatism found on southeastern Laurentia and in the Eastern Cordillera of Peru. Recent palaeomagnetic data (McCausland *et al*. 2007) suggests that Laurentia was already well separated from the Amazonian margin in late Neoproterozoic time, although palaeomagnetic data for Amazonia during the late Neoproterozoic are limited to the study of Trindade *et al*. (2003) of the Neoproterozoic Puga cap carbonate on the Amazonian Craton. A wide ocean separating Laurentia and Amazonia during the late Neoproterozoic would also imply that the unequivocal ca. 550 Ma rifting event that occurred along Laurentia's Iapetan margin did not represent the separation of true cratons from Laurentia, but rather that of peri-Laurentian Appalachian basement terranes (Cawood *et al*. 2001; Waldron and van Staal 2001; van Staal *et al*. 2007).

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