Invited Review

On the origins of the Iapetus ocean

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\begin{abstract}
The Iapetus Ocean opened during the fission of the supercontinent Rodinia, from the breakup of three of its core continental constituents: Laurentia, Baltica and Amazonia. The timing of Iapetus opening is still much debated, with estimates ranging from 700 to 550 Ma. Similarly debated is exactly how Laurentia, Baltica and Amazonia were positioned relative to each other immediately before their breakup. In this study, we reconsider the timing and framework of Iapetus opening by integrating the fragmentary mid-Neoproterozoic to early Cambrian observational records from these continents. We first demonstrate that paleomagnetic data, despite being both sparse and probably contaminated by some global-scale, non-uniformitarian process in Ediacaran time, support the existence of a wide ocean between these continents by 575 Ma. However, the paleomagnetic data alone are insufficient to allow the formulation of more specific conclusions concerning the timing and paleogeography of Iapetus opening. We therefore conduct an extensive review of the mid-Neoproterozoic to Cambrian geology of eastern Laurentia, western Baltica and western Amazonia which, jointly interpreted with the paleomagnetic constraints, allow us to construct a self-consistent and geodynamically feasible plate tectonic model. In this model, the breakup of Laurentia, Baltica and Amazonia was polyphase, involving the spalling of multiple marginal terranes from Laurentia and the successive opening of several oceans, including a composite ‘Iapetus Ocean’. The first phase of continental breakup occurred between eastern Laurentia and western Amazonia at 750–700 Ma, leading to the opening of the Puncoviscana Ocean. This was followed by the opening of the eastern branch of the Iapetus Ocean, between Laurentia and Baltica, at ~590 Ma, which may have been instigated by emplacement of the Central Iapetus Magmatic Province. The western branch of Iapetus subsequently opened at ~550 Ma by the detachment of marginal terranes from eastern Laurentia, following a protracted phase of rifting. We contend that our preferred scenario is the simplest solution given the presently available evidence but throughout this review we underline key outstanding questions and the attendant uncertainties in our preferred model.
\end{abstract}

1. Introduction

The Iapetus Ocean is irrevocably intertwined with plate tectonics. As a concept, the birth of the Iapetus can be traced to the predawn of the plate tectonic paradigm, when J. Tuzo Wilson posed the question ‘Did the Atlantic close and then re-open?’ (Wilson, 1966). The monumental leap forward that this question presented was not the recognition that the Atlantic had earlier opened—indeed several others had already presented reconstructions of the Atlantic-bordering continents prior to the birth of that basin (Wegener, 1912)—but rather the realization that some other ocean had closed before the Atlantic opened. This led directly to the broader understanding that ocean basins have cyclically opened and closed across geologic time (the ‘Wilson Cycle’), and further spurred the then-imminent plate tectonic revolution (Wilson et al., 2019). Wilson (1966) originally called this ocean which preceded the Atlantic the ‘proto-Atlantic’, but we now know it as the ‘Iapetus Ocean’, so named for the mythological figure who fathered the titan Atlas from which the Atlantic Ocean derived its name (Harland and Gayer, 1972). In fitting with the grandeur of being named after a Greek titan, the Iapetus Ocean has, ever-since Wilson posed his famous question, been recognized as one of the most important oceans of the Phanerozoic, and thus one of the keys to an understanding of global paleogeography.

And yet, despite that key historical role that the Iapetus Ocean has played in our understanding of tectonics, both as a concept and in the
sense of our practical knowledge of paleogeography, a clear understanding of the opening and early evolution of the Iapetus Ocean has eluded us. Conceptually, the opening of the Iapetus is often associated with the final disaggregation of the supercontinent Rodinia, via a three-way separation of the continents of Laurentia, Baltica and Amazonia, either in the latest Precambrian or at the dawn of the Phanerozoic. But the timing of the opening of that great ocean, the framework geography of the continents among which it opened from, and indeed even the identity of the continents that flanked the early opening of the Iapetus remain more speculative than demonstrated. Given the great importance that the opening of this ocean plays on our understanding of late Precambrian and early Phanerozoic geology, we consider it worthwhile to summarize our present state of knowledge and to attempt to answer the following outstanding questions: When and where did the Iapetus Ocean open? What were the geodynamic mechanisms responsible for its opening? What are the critical remaining uncertainties concerning the evolution of the Iapetus and what acquirable constraints can address them?

In the following sections, we first consider reconstructions of Rodinia—the paleogeography from which the Iapetus opened—and then review the existing paleomagnetic and geologic data available from the major continental blocks that are thought to have played a role in Iapetus opening. With these paleomagnetic and geologic constraints in hand, we then consider and evaluate alternative opening scenarios for the Iapetus, and highlight important opportunities for targeted future work that may significantly advance the lingering impasse about how it evolved.

2. Rodinia reconstructions in the literature

As the Iapetus Ocean is thought to have been born from the final breakup of the supercontinent Rodinia (Hoffman, 1991), any consideration of Iapetus opening necessarily requires a consideration of the geography of Rodinia. Rodinia was assembled in the late Mesoproterozoic through a series of continental collisions and major orogenic events, commonly thought to include the Grenville orogeny of Laurentia, the Sveconorwegian orogeny of Baltica and the Sunnás orogeny of Amazonia (e.g. Cawood and Pisarevsky, 2017). By the start of the Neoproterozoic, which remains paleomagnetically unconstrained within Rodinia during the 1.1–2 Ga, but Amazonia may not have had any connection with Laurentia in Rodinia (Merdith et al., 2017; Slagstad et al., 2019).

For example, alternative reconstructions of Amazonia in Rodinia have placed the western margin of Amazonia along the eastern margin of Laurentia in Pisarevsky et al. (2003), along northeast Laurentia (Greenland) in Dalziel (1992), and along the southeastern tip of Laurentia in Meert and Torsvik (2003). Others have suggested that Amazonia may not have had any connection with Laurentia in Rodinia (Evans, 2009). In the case of Baltica, its western margin is often reconstructed along the eastern margin of Greenland or Scotland-Ireland (Dalziel, 1997; Evans, 2009; Hoffman, 1991), but there are also alternative models where Baltica is ‘upside down’ in Rodinia (Hartz and Torsvik, 2002) or not part of the supercontinent (Slagstad et al., 2019).

Despite these uncertainties, one relative configuration of Laurentia, Amazonia and Baltica in Rodinia is clearly favored in the literature—that in which the western margin of Amazonia is restored against the eastern margin of Laurentia, and the western margin of Baltica is reconstructed along Scotland-Ireland, between the North American craton and Greenland (Fig. 1) (Hoffman, 1991; Li et al., 2008; Pisarevsky et al., 2003; Weil et al., 1998). That restoration of Baltica is supported by age similarities between the ~1 Ga Sveconorwegian orogen and the Grenville orogen, and pre-Grenvillian tectonic zones in Baltica and Amazonia, the ‘East Iapetus Ocean between Laurentia and Baltica, and highlighting important opportunities for targeted future work that may significantly advance the lingering impasse about how it evolved.

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Tonian-early Cryogenian opening around 750–700 Ma (Li et al., 2013), (2) a mid-Ediacaran opening around 600–570 Ma (Cawood et al., 2001), and (3) a latest Ediacaran-early Cambrian opening around 550–530 Ma (Dalziel, 1992; Pisarevsky et al., 2008). After reviewing the paleomagnetic constraints (section 3) and geological observations (section 4) that are potentially relevant to unraveling the opening of the Iapetus Ocean, we will return to entertain these various proposed breakup timings and elaborate on our own preferred scenario in section 5.

3. Paleomagnetic constraints

Neoproterozoic-Cambrian paleomagnetic data from Laurentia, Baltica and Amazonia are key to testing the different proposed scenarios of the Iapetus Ocean’s opening. In this section, we will first review and discuss the quality of the available paleomagnetic data from 750 to 500 Ma for these continental blocks, and then discuss their compatibility with the different proposed opening scenarios.

Reliable paleomagnetic data between 750 and 500 Ma are unevenly distributed in time, and are more numerous at the beginning of the Phanerozoic (541–500 Ma) than during the Precambrian. The paleomagnetic data of the Ediacaran (635–541 Ma) are particularly scattered and difficult to interpret and have long been debated in the literature, leading to an array of alternative continental reconstructions (e.g. Li et al., 2008; Pisarevsky et al., 2008). In particular, the Ediacaran data from multiple continents are associated with two distinct groups of poles of similar age that seem to offer contradictory (low vs. high) paleolatitude constraints (e.g. Abrajevitch and Van der Voo, 2010). The use of both groups implies plate velocities exceeding those that are commonly regarded as reasonable in the Phanerozoic (Torvik et al., 2012; Zahirovic et al., 2015). Various hypotheses have been proposed in the literature to explain these paleomagnetic data, namely: (1) some of the data represent remagnetizations (e.g. Bono and Tarduno, 2015), (2) Earth’s magnetic field was alternatively dominated by an axial dipole and a highly-inclined (equatorial) dipole (Abrajevitch and Van der Voo, 2010), (3) the magnetic field was reversing rapidly, leading to a frequent record of transitional paleomagnetic data (Halls et al., 2015) and (4) very fast episodes of true polar wander (TPW) occurred—wherein the entire lithosphere and mantle rotate relative to the spin-axis (e.g. Evans, 2003; Robert et al., 2017).

Robert et al. (2017) recently presented a series of new Ediacaran–Cambrian paleomagnetic poles from West Africa which display a large ‘shaped’ geographic distribution, further substantiating the velocity of the low and high latitude pole groups. As discussed by Robert et al. (2017), that this paleomagnetic oscillation is also apparent in the best paleomagnetic data from Laurentia and Baltica would suggest that these poles record a real, abrupt and global-scale event from 615 to 565 Ma. Together, those observations pose a significant challenge to the remagnetization hypothesis and support a dipolar geometry of the magnetic field at that time. Nevertheless, additional data are needed to better constrain the apparent polar wander paths (APWPs) of all those blocks, and to distinguish between the remaining hypotheses (e.g. an equatorial dipole vs. TPW vs. fast moving plates). It is beyond the scope of this study to extensively interrogate these competing hypotheses, and so here we only consider them insofar as it is necessary to understand what paleogeographic constraints can be confidently extracted from the data.

Our selection of paleomagnetic poles in the late Tonian and Cryogenian is based on a filtering of the Paleomagia database (Veikkolainen et al., 2017), as described below, which was queried in May 2020. For the Ediacaran and early Cambrian, our pole compilation is essentially that of Robert et al. (2017), whereas for younger times we use the running mean poles without inclination correction reported by Torvik et al. (2012). The pole selection was made according to the following criteria, slightly modified from the Q criteria of Van der Voo (1990): the age uncertainty of the pole must be <16 Ma (Q#1), the number of samples used to calculate the pole must be >24, and the associated A\(_{95}\) must be <16° (Q#2), progressive thermal and/or alternating field demagnetization techniques must have been applied to isolate the magnetic components (Q#3), and the tectonic coherence of the sampled block relative to the parent craton should be established (Q#5). Van der Voo (1990) included several additional criteria (Q#4, 6 and 7) that we do not strictly employ because a failure to meet them does not necessarily imply that a pole is not suitable for use in making continental reconstructions. Q#4 is fulfilled if a paleomagnetic result is supported by a field/stability test, but the absence of such does not necessarily imply remagnetization. However, we discard all poles associated with a negative field/stability test. Q#6 is fulfilled if there are reversals. The presence of reversals generally indicates that secular variation of the magnetic field is sufficiently averaged, and thereby improves confidence that the associated remanence could be primary. Nevertheless, we do not discard results lacking reversals because the reversal frequency is unknown for most of the Precambrian. Q#7 is fulfilled if a pole does not resemble younger poles, which again improves confidence that it could be primary. We do not adopt this criterion as the likelihood of having a Precambrian segment of an APWP crossing a younger segment is high (Pivarunas et al., 2018). However, in cases where resemblance to a younger pole is supported by a clear mechanism of remagnetization, such as an orogenic event, the pole is discarded. A detailed explanation of our pole selection is presented below; the selected poles are reported in Table 1 and the excluded poles are listed in Appendix A.

3.1. Selection of paleomagnetic poles

3.1.1. Laurentia

Because the late Tonian and Cryogenian (~750–635 Ma) dataset of Laurentia is very sparse, it is worth briefly discussing all the data available, including those which we reject. After querying the Paleomagia database and removing all poles that fail to meet Q#1 and/or Q#2, we are left with results from the Franklin dykes, sills and extrusives in northern Canada and Greenland (e.g. Denyszyn et al., 2009), the Kwagunt Formation (Fm.) in Arizona, USA (Weil et al., 2004; Elston et al., 2002), and the Mount Harper volcanics in Yukon, Canada (Eybler et al., 2017; Park et al., 1992). The Franklin dykes are well dated to 721–712 Ma by U-Pb on baddeleyite, yield robust paleomagnetic results supported by a baked contact test, and most likely carry a primary remanence (Denyszyn et al., 2009). Denyszyn et al. (2009) reported a combined pole incorporating both their new results and those of previous studies (Christie and Fahrig, 1983; Fahrig et al., 1971; Fahrig and Schwarz, 1973; Palmer et al., 1983; Palmer and Hayatsu, 1975; Park, 1994; Park, 1981; Robertson and Baragar, 1972) and we therefore consider this pole to supersede those earlier results, and so exclude them from our compilation. The paleomagnetic pole from the 739–716 Ma Mount Harper volcanics is 50° away from the Franklin dykes pole of Denyszyn et al. (2009), and Eybler et al. (2017) concluded that large rotations occurred between the Yukon block and cratonic Laurentia, and so this pole does not meet Q#5 and is therefore excluded from our selection. The pole from the Kwagunt Fm. is supported by a positive fold test constraining the age of magnetization as pre-middle Cambrian, and it is considered likely primary. An ash layer at the top of the Kwagunt shallow-marine sedimentary sequence has been dated to 742 ± 6 Ma (Karlstrom et al., 2000) and is taken as the best age estimate of the paleomagnetic pole. Our paleomagnetic compilation for Laurentia between 750 and 635 Ma thus comprises only two paleomagnetic poles (Table 1).

For the Ediacaran, our pole selection is the same as Robert et al. (2017), and consists of nine poles with ages spanning 615–550 Ma. For the Cambrian we include the ~532 Ma Mont Rigaud and Chatam-Grenville pole of McCausland et al. (2007), and from 510 Ma we adopt the mean poles reported by Torvik et al. (2012). The polarity of the Ediacaran poles has been assigned based on the APWPs defined by Robert et al. (2017), while the polarity of the late Tonian and Cryogenian poles was so chosen to minimize their distance from the
Table 1
Selected paleomagnetic poles from 750 to 520 Ma. Q = Modified paleomagnetic reliability criteria of Van der Voo (1990) (see text); Plat/Plon = paleopole latitude/longitude; A95 = radius of cone of 95% confidence about the pole (in degrees); Age/Error age = best-estimated age of pole / its age uncertainty (in Ma).

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<th>ID</th>
<th>Name</th>
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<th>Q details</th>
<th>Plat</th>
<th>Plon</th>
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3.1.2. Baltica

A large proportion of the late Tonian and Cryogenian paleomagnetic poles from Baltica listed in the Paleomagia database are ill-dated and cannot be used for paleogeographic reconstruction. We note that Torsvik et al. (1996) used three poles to define the position of Baltica at ~750 Ma, but the ages of those poles are mostly inferred and do not meet our criterion Q#1, and so we do not include them in our selection. Salminen et al. (2006) reported a paleomagnetic pole from the Jánisjárvi impactites, which were subsequently dated to $682 \pm 4$ Ma by $^{40}$Ar/$^{39}$Ar (Jourdan et al., 2008). This pole is constrained by only 21 samples, and thus fails to meet criterion Q#2. However, as it is otherwise a good quality pole supported by a positive contact test, we decided to provisionally retain this result.

For the Ediacaran, our compilation for Baltica follows that of Robert et al. (2017), which comprises ten paleopoles. Among the Ediacaran poles we dismiss, the Nyborg Fm. pole is well-defined and supported by a positive fold test (Torsvik et al., 1995), but its age is only loosely bracketed between ~635 and 580 Ma (Halverson et al., 2005). The Cambrian paleomagnetic dataset of Baltica is particularly ill-constrained, as the few available poles are either based on few samples (e.g. Andrarum limestone pole; Torsvik and Rehnström, 2001), or are derived from rocks located in the Caledonian thrust belt (e.g. Torne-skag/Dividal pole; Rehnström and Torsvik, 2003), or are not well-constrained in age (e.g. Bursov pole; Fedorova et al., 2016). Additionally, these poles are unconstrained by field tests and (with the exception of the Brusov pole) fall close to Permo-Triassic poles from Baltica, and so there is a suspicion that they are remagnetizations (Meert, 2014; Robert et al., 2017). Consequently, we include no Cambrian poles from Baltica in our compilation, and, following Meert (2014), the oldest Paleozoic data we use is the Early Ordovician (480 Ma) mean pole of Torsvik et al. (2012).

As for the Precambrian poles from Laurentia, the polarities of the Ediacaran poles from Baltica follow those assigned by Robert et al. (2017), while the polarity of the Cryogenian Jánisjárvi impactite pole was chosen in order to minimize its distance from the Ediacaran poles.

3.1.3. Amazonia and other blocks of Gondwana

The Precambrian and earliest Paleozoic paleomagnetic records from Amazonia are very sparse, but because that craton became part of the growing landmass of Gondwana during its amalgamation from about 750 to 520 Ma (Caby, 2003; Collins and Pisarevsky, 2005; Meert, 2003), we may also consider poles from blocks it became coheren with. Already before 750 Ma, the cratons of Amazonia and West Africa had probably united (hereafter together referred to as ‘West Gondwana’), as inferred from geological similarities (Torquato and Cordani, 1981; Vil- leneuve and Conne, 1994) and paleomagnetic constraints (D’Agrella-Filho et al., 2016; Nomade et al., 2003; Onstott and Hargraves, 1981) which specifically suggest those blocks united by ~2 Ga. On the other hand, Trindade et al. (2006) proposed that Amazonia was isolated from the rest of Gondwana (including West Africa) by the ‘Clymene Ocean’ until the Cambrian. Notably, the Gurupi belt located between Amazonia and West Africa recorded an extensional event around ~730 Ma, followed by greenschist to amphibolite facies metamorphism between 580 and 520 Ma (Klein et al., 2020). However, given the absence of late Neoproterozoic oceanic and arc-related rocks, Klein et al. (2020) attributed the Ediacaran-early Cambrian metamorphic event to intracontinental tectonics. Furthermore, more recent models of the Clymene Ocean (e.g. Tovver and Trindade, 2014) depict that ocean separating a united West Gondwana from blocks of Central Gondwana (see below), rather than bisecting West Gondwana. It is therefore reasonable to consider Amazonia and West Africa effectively coherent since before 750 Ma; however, because this turns out to be a critical assumption, we will note its implications several times in the following sections. The cratons of São Francisco and Congo (hereafter together called ‘Central Gondwana’) are also thought to have formed a single continental entity.
since the Paleoproterozoic (e.g. D’Agrella-Filho and Cordani, 2017). In turn, the two composite landmasses of West and Central Gondwana may have collided in the early Ediacaran (~630–600 Ma), leading to the formation of the West Gondwana orogen (e.g. Ganade de Araujo et al., 2014). However, a younger, early Cambrian age has alternatively been proposed for the final assembly of those landmasses (e.g. Tohver and Trindade, 2014), again associated with the closure of the Cymene Ocean. That younger age remains controversial, but if correct it has vast implications for the history of Iapetus opening, as will be discussed later. Another major collision event occurred in the late Ediacaran to early Cambrian, when the continental blocks of East Gondwana (Australia and East Antarctica) amalgamated with West-Central Gondwana (e.g. Meert, 2003). The position of Rio de la Plata in Rodinia, and the timing of its subsequent amalgamation with greater Gondwana remains a topic of active debate, but it is often assumed to have amalgamated with the Congo craton during the mid- to late Ediacaran (Li et al., 2008; Meredith et al., 2017). Thus, in addition to the paleomagnetic data derived from Amazonia itself, paleomagnetic poles that are relevant to consider in the following include those from West Africa from 750 Ma, plus those from São Francisco, Congo and Rio de la Plata after 630 Ma. However, because the specific amalgamation timing of these blocks is not strictly known, we report their Precambrian poles independently (Table 1). By the end of the early Cambrian, poles from East Gondwana can also be used to constrain the position of Amazonia.

For the late Tonian and Cryogenian, the Paleomagia database reports no paleomagnetic results from Amazonia and six results from West Africa. These six poles, which are from the Tin Akof carbonates, the Tin Dioulf cap carbonates, the Guinguettes sediments, the Digouéra sediments, the Banardougo sediments and the Tiara sediments (Boudzoumou et al., 2011), were determined from few samples and do not pass criterion Q#2. Moreover, these poles lie close to the Carboniferous and Permian segments of the APWP of Gondwana (Torsvik et al., 2012), and remagnetizations of such age have been repeatedly identified in the Paleozoic and Proterozoic rocks of West Africa (Perrin and Prévot, 1988; Robert et al., 2017). Therefore, these poles also fail criterion Q#7, except the pole from the Tin Dioulf cap carbonate, which is supported by a positive fold test where the folding is of Pan-African age. Nevertheless, the number of samples used to perform that stability test (n = 10) was very low, and the statistical result is questionable (Evans and Raub, 2011). Consequently, our compilation includes no paleomagnetic constraints from Amazonia during the late Tonian and Cryogenian.

For the Ediacaran, our compilation includes four poles from West Gondwana, one pole from Central Gondwana, and two poles from Rio de la Plata (Table 1). This selection is almost the same as the one provided by Robert et al. (2017), except that we added one pole from Amazonia, derived from the 605 Ma alkaline complex of Planalto da Serra (Garcia et al., 2013) and we did not retain the pole from the Puga cap carbonates (Trindade et al., 2003). The Planalto da Serra pole is based on a well-separated paleomagnetic direction and the host rocks are well-dated, fulfilling criteria Q#1–3. However, this pole lies close to the Cambrian segment of the APWP of Gondwana and there is no paleomagnetic test asserting its primary nature (i.e. Q#7 is not met). On the other hand, it is statistically indistinguishable from the coeval Adma Diorite pole of West Africa when the continents are reassembled into their relative positions in Gondwana. We therefore provisionally include this pole in our compilation. Regarding the pole from the Puga cap carbonates, the presence of reversals following stratigraphy tend to support a primary origin, but the pole is similar to the present-day field and so could represent a recent remagnetization. Because there is no paleomagnetic/field test that otherwise constrains the age of magnetization, we excluded this pole from our compilation. As a conservative measure, we also excluded an Ediacaran pole from the Arabia-Nubia Shield (the Dokhan Fm, pole) used by Robert et al. (2017) because that block lies far from Amazonia, and is separated from it by several orogenic zones that could have still been active in Ediacaran time (e.g. Sengör et al., 2020). Nevertheless, we note some paleogeographic implications that the use of this pole would have in the discussion section (section 5.1.3).

For the early Cambrian, our compilation includes only one pole from West Gondwana—the 536 Ma Djebel Boho volcanics pole of Robert et al. (2017). Starting at 520 Ma, we adopt the mean poles of Gondwana reported by Torsvik et al. (2012). We elected not to use the early Cambrian mean poles of Torsvik et al. (2012) because most of the constituent poles used to compute those means come from Australia, which may not have been totally amalgamated with the rest of Gondwana then (Robert et al., 2017). The compilation of Torsvik et al. (2012) also included the 547 Ma Sinya metabolerite pole from East Africa (Meert and Van Der Voo, 1996). This pole, usually considered as a key pole, is a remagnetization isolated from greenshist facies dikes. 40Ar/39Ar dating of the dikes yielded an age of 547 ± 4 Ma, and this is commonly assumed to represent the age of the remagnetization. Nevertheless, the age of this pole is ill-constrained as there is no demonstration that the isotopic age is associated with the remanence acquisition event, and we therefore exclude this pole from our compilation. As before, the polarity assignments of the Ediacaran poles follow those of Robert et al. (2017).

3.2. Paleogeographic constraints from the selected paleomagnetic data

3.2.1. The nature of paleomagnetic variations: individual plate motions vs. global mechanisms

Ideally, we would like to be able to make quantitative paleogeographic reconstructions of Laurentia, Baltica and Amazonia (or West Gondwana) through time according to their individual paleomagnetic data. However, such an approach assumes that paleomagnetic variations are effectively due only to individual plate motions, whereas the late Neoproterozoic dataset comprises large paleomagnetic variations that likely express a combination of continental drift and the occurrence of some global, non-uniformitarian phenomena (e.g. rapid TPW and/or deviations of the magnetic dipole) (Abrajевич and Van Der Voo, 2010; Robert et al., 2017). On the other hand, abrupt ‘globally’ coincident paleomagnetic changes could also manifest the rapid movement of a single supercontinent (i.e. Rodinia). Thus, a robust paleomagnetic test of the tectonic coherence of Rodinia through time necessitates some consideration of these different contributions to the observed paleomagnetic signal. Unfortunately, there is not presently enough data to rigorously distinguish these contributions, but we may draw some useful provisional inferences from a collective analysis of the compiled data. Because we don’t know the time-dependent paleogeography a priori, it is instructive to begin with a null hypothesis that Rodinia remained intact across the period of interest (~750–480 Ma) and that the paleomagnetic variations can be explained purely by global mechanisms such as TPW or magnetic dipole deviations, or simply the drift of the intact supercontinent itself. This provides us with a convenient frame of reference to explore and compare the paleomagnetic data from the different continents without first interpreting them paleogeographically beyond adoption of the null hypothesis. For this initial paleogeographic configuration, we chose to use the Rodinia fit of Li et al. (2008) (Fig. 2), but the amplitude of the late Neoproterozoic paleomagnetic variations are large enough that the use of an alternative Rodinia fit as a starting point would not change the outcome of the following analysis.

In examining our compiled paleomagnetic data from Laurentia, we see that they mostly fall into two groups of poles, as aforementioned: one located close to, or over Laurentia itself (Fig. 2); poles GD2, GDB, GDE, CC and CTA), and another approximately ~90° away (poles F, K, LR, SIA, SC, MR and LS10-480 poles). We will refer to these as the ‘high latitude’ and ‘low latitude’ groups, respectively (but note that these designations are only meant to discriminate between the pole groups in a relative sense and may not necessarily reflect absolute latitude). Two groups of poles are also recognizable in the compiled datasets from Baltica and Gondwana. When these blocks are reconstructed relative to Laurentia in the Rodinia fit of Li et al. (2008) (which again is maintained from 750 to 480 Ma as a frame of reference; Fig. 2a), these two groups of poles become coincident with the ‘high’ and ‘low latitude’ pole groups of
Laurentia (Fig. 2b). To further highlight the distinct pole groups, we define two distance markers: a circle of 35° centered on pole GD2, which encircles most of the 'high latitude' poles, and a circle of 90°, also centered on the GD2 pole, which passes through the distribution of the 'low latitude' poles (Fig. 2b). To facilitate consideration of the temporal relationships of these spatially-defined pole groups, we also show the data in time, expressed as arc distances with respect to pole GD2, which has a central position in the high latitude group (Fig. 3).

During the late Tonian and Cryogenian (Fig. 5a) the general sparsity of reliable paleomagnetic data and the absence of coeval data from multiple continental blocks in particular precludes a robust characterization of the paleomagnetic variations observed then. Nevertheless, the few available data do not seem to exhibit the abrupt directional changes that are characteristic of the Ediacaran. Specifically, from 750 to 700 Ma, the only poles in our compilation are from Laurentia (K and F poles), and they fall very close to one another (Figs. 2b, 3a). Between 700 and 635 Ma, the only pole available from Baltica is the 680 Ma Janisjarvi pole, which plots about 90° from the 720 Ma Franklin pole from Laurentia. Given the ~40 Ma age difference between those data, our null hypothesis would imply their separation can be explained by the motion of Rodinia at a rate of ~25 cm/yr—which is fast but may not be implausible for the drift of an individual plate. Because coeval paleomagnetic data are not available from multiple blocks during this interval, it is not possible to distinguish any relative plate motions between Laurentia, Baltica and Amazonia in the late Tonian-Cryogenian.

The Ediacaran presents the first opportunity to compare paleomagnetic data of the same age from the different continents, starting at ~615 Ma. At that time, the reconstructed poles from each continental block are located in the ‘low-latitude’ group (arc distances to GD2 pole around 90°) (Figs. 2b, 3b). However, by the mid-Ediacaran (590–575 Ma), the paleomagnetic poles from all three of those continental blocks shift to the ‘high latitude’ group (with arc distances to GD2 mostly <35°), before returning to the ‘low latitude’ group by ~565 Ma (Figs. 2b, 3b) (Robert et al., 2017). Although the data are sparse, we note that the selected poles from Rio de la Plata and Congo (represented respectively by triangles and a square in Fig. 3b) also appear to be part of the ‘high latitude’ group around 590–575 Ma, together with the result from West Gondwana. These abrupt, large magnitude paleomagnetic swings from 615 to 590 Ma and 575 to 565 Ma appear coeval and of similar amplitude and are consistent with our null hypothesis that they manifest either some global phenomena or the rapid drift of an intact Rodinia. However, interpreting the paleomagnetic swings as pure plate motion (of an intact Rodinia) would imply drift rates of ~30 to 180 cm/yr—significantly faster than the plate ‘speed limit’ inferred from Phanerozoic motions (Zahirovic et al., 2015). We therefore contend that these paleomagnetic swings are more likely a manifestation of rapid TPW or some instability of the geomatic field.

The latest Ediacaran to early Cambrian dataset is especially curious, because while poles from Laurentia and West Gondwana remain in the ‘low latitude’ group after 565 Ma, the poles from Baltica exhibit another abrupt shift, returning to the ‘high latitude’ group by ~550 Ma (Figs. 2b, 3b). Given the sparse data from Laurentia and West Gondwana at this time, it is conceivable that the dataset from Baltica expresses another ‘global’ signal that has not yet been identified in the other two datasets, but in the absence of further evidence of that, we rather interpret these differences to mark significant relative plate motions between those continents. Thus, our null hypothesis fails to explain the rapid shift observed in the data from Baltica at this time.

Previous studies also proposed a ‘high-latitude’ position for the Gondwana poles at 550 Ma (Cawood et al., 2001; McCausland and Hodych, 1998), much closer to the latest Ediacaran poles of Baltica (e.g. G550 in Figs. 2b, 3b). However, as pointed out by Robert et al. (2017), the suitability of the poles used to build the APWP of Gondwana from 550 to 530 Ma (G550-530; Torsvik et al., 2012) is uncertain. Indeed, as detailed in section 3.1.3., the age of the Sinyai metadolerite pole is not detailed in section 3.1.3., the age of the Sinyai metadolerite pole is not

Summarizing the insights drawn from this simple analysis: we
distinguish two time intervals where the paleomagnetic data from Laurentia, Baltica and West Gondwana exhibit very rapid changes that are approximately coincident (both spatially and temporally) when the continents are reassembled into a Rodinia fit. We consider it highly likely that these two intervals, 615 to 590 Ma and 575 to 565 Ma, manifest some non-uniformitarian process, and so it is unclear if/how these data express individual plate motions with respect to the planetary spin-axis (given that the signal may be dominated by other contributions). Nevertheless, the apparent internal consistency of these data would seem to suggest that we may still be able to infer something about relative plate motions through instantaneous comparisons and, fascinatingly, they may provide a pathway toward constraining relative paleolatitude. This is explored in the next section.

3.2.2. Paleomagnetic tests for Iapetus Ocean opening

If the grossest paleomagnetic variations observed in the late Precambrian are indeed due to some global, non-uniformitarian phenomena rather than individual plate motions, the true time-dependent paleogeographic locations of the continents are also obscured. However, the relative positions of the continents may still be determinable through instantaneous comparisons of their coeval paleomagnetic data at a few key times. Because the initial positions of Laurentia, Amazonia and Baltica in the supercontinent Rodinia are still debated, we consider a series of alternative fits of Rodinia and test their compatibility with the late Neoproterozoic paleomagnetic dataset. Although we refrain from ascribing a particular mechanism to the global paleomagnetic variations, in the following we maintain a reference frame that treats all paleomagnetic data as derived from an axial dipolar field for convenience.

As aforementioned, the very sparse late Tonian and Cryogenian paleomagnetic data (750–635 Ma) are too few to discern any relative motions between Laurentia, Baltica and Amazonia, and so the early
Ediacaran (~615 Ma) presents the first opportunity to directly compare paleomagnetic poles of the same age from these continents (Fig. 3). To do this succinctly, we consider a range of published reconstructions that we consider approximately representative of the different ‘families’ of reconstructions in the literature (Fig. 4a-g). The first family (Fig. 4a-c) represents a collection of reconstructions similar to the ‘conventional’ positioning of Laurentia, Baltica and Amazonia in Rodinia, and can be thought of as alternative early Ediacaran states evolved from the ~750 Ma reconstruction of Li et al. (2008). In the simplest variation (Fig. 4a), the relative positions of Laurentia, Baltica and West Gondwana are exactly the same as in the 750 Ma fit of Li et al. (2008), illustrating a case where the Iapetus Ocean has not yet started to open. In another variation, based on the model of Trindade et al. (2006) (Fig. 4b), the same relative positions of Laurentia, Baltica and Amazonia are maintained but West Africa is separated from Amazonia by a wide ocean (Clymene Ocean), such that its paleomagnetic data cannot be used to constrain the location of Amazonia. In a third variation, based on the model of Robert et al. (2020) (Fig. 4c), Baltica and West Gondwana are rotated counterclockwise with respect to Laurentia about an Euler pole near eastern Greenland. This variation illustrates a case where Amazonia rifted from eastern Laurentia before 615 Ma (opening the West Iapetus), but the East Iapetus and Tornquist oceans have not yet opened. In all these reconstructions, the 95% confidence ellipses of the 615 Ma paleomagnetic data from Laurentia, Baltica, Amazonia and West Africa (poles LR, ED, PS and AD; see Table 1) intersect the geographic pole (the AD pole is actually distinct from the geographic pole in pan a but this could be ameliorated by a minor rotation of Amazonia), and so all these configurations (and minor variations thereof) can be considered paleomagnetically viable.

In contrast to those variations on the ‘conventional’ fit, we also consider several alternative families of configurations where Baltica and West Gondwana are positioned differently with respect to Laurentia. In one alternative family, Laurentia, Baltica and West Gondwana remain connected and have approximately the same azimuthal orientations as in the ‘conventional’ fit, but Baltica and West Gondwana are significantly shifted along the Laurentian margin. As one example, we show a reconstruction based on the models of Dalziel (1992) and Torsvik et al. (1996) (Fig. 4d). As before, the corresponding reconstruction of the paleomagnetic data indicates that this reconstruction (and thus similar minor variations) is paleomagnetically viable. In another alternative family, the orientation of West Gondwana and Laurentia remains the same, but Baltica is either azimuthally inverted, as in Hartz and Torsvik (2002) (Fig. 4e) or is otherwise independent, as in Slagstad et al. (2019) (Fig. 4f). In the latter variant, the relative paleolatitude and orientation of Baltica can be made identical to that of the earlier families (just with a shift in longitude) and so it is paleomagnetically permissible. In contrast, the strong relative rotation of Baltica in the former variant is clearly inconsistent with the location of the ED pole (Fig. 4e). In yet another alternate family, Laurentia and Baltica remain in the same relative position, but West Gondwana represents an independent block, as in Evans (2009) (Fig. 4g). As in the case of an independent Baltica, an independent West Gondwana (or Amazonia) can maintain the same relative paleolatitude and orientation as in the earlier configurations (just with a paleolongitude shift) and so cannot be excluded on paleomagnetic grounds. Collectively, these comparisons reveal that while the early Ediacaran paleomagnetic data can exclude some of the more exotic reconstructions proposed, they are not sufficiently well-resolved to discriminate between models that involve rifting (and migration) between Laurentia, Baltica and Amazonia (± West Africa) by this time, and those that do not; nor are they especially sensitive to the specific relative configurations of those continents.

As shown in the previous section, the mid-to-late Ediacaran paleomagnetic data from Laurentia, Baltica and West Gondwana exhibit coincident, large-magnitude changes that appear too rapid to be due to plate motions, and are likely rather dominated by some global phenomena (e.g. TPW or magnetic field perturbations; Abrajevitch and Van der Voo, 2010; Robert et al., 2017). Crucially, if the excursions indeed record either rapid TPW or the transient dominance of an equatorial dipole—and the corresponding paleomagnetic changes occurred at rates much greater than those of contemporaneous relative plate motions—then the collection of paleomagnetic poles can be used to fully define the relative positions of the continents (i.e. including relative paleolongitude). In Fig. 4h-j we present reconstructions of Laurentia, Baltica and West-Central Gondwana at 575 and 565 Ma that were built following this logic: the continents are positioned such that all the constituent poles of the two age groups are internally coincident, even though the groups themselves are ~90° apart. Notably, the validity of this approach is dependent on how accurately the large paleomagnetic excursions have been recorded, and in particular on the assumption that the available data adequately capture the end-points (‘tips’) of the true track of the global excursions. Given that the 590–575 and 565 Ma poles from multiple continents (including Congo and Rio de la Plata) are similarly separated by about 90°, we contend that it is reasonable to assume that they approximate the tips of the true paleomagnetic excursion track, although clearly more data are needed to better resolve it.

The reconstructed positions of Laurentia, Baltica and large parts of West-Central Gondwana (West Africa, Congo and Rio de la Plata) derived from this approach suggest that Laurentia was separated from both Baltica and West-Central Gondwana by substantial oceans by 575 Ma. However, the specific identity of those oceans (e.g. whether the West and East Iapetus or Clymene and Tornquist oceans) is not determinable from these data, as there are no mid-Ediacaran constraints on Amazonia. Thus several paleogeographic ‘families’ are again able to explain the data available. One straightforward solution is to assume that Amazonia remained coherent with West Africa since 615 Ma (i.e. following from reconstructions in Fig. 4a, c or 4d), which would imply that by 575 Ma, Amazonia and Laurentia were separated by a wide West Iapetus Ocean (Fig. 4h) (Robert et al., 2017). Alternatively, it is equally permissible by the paleomagnetic data (or rather the absence of it) to maintain a connection between Laurentia and Amazonia at 575–565 Ma (i.e. following from reconstruction Fig. 4b), so that the ocean between Laurentia and the other Gondwana blocks would have been the Clymene Ocean (Fig. 4i) (Trindade et al., 2006). A variation on this latter reconstruction was recently presented by Wen et al. (2020), who restored Amazonia beside Laurentia at 575 Ma, but specifically placed it along the Scotland-Ireland segment (Fig. 4j). Notably, their restoration of Amazonia was based on paleomagnetic data from West Avalonia, which we do not consider here. To accommodate this alternative position of Amazonia, Wen et al. (2020) also moved Baltica to relatively higher latitudes, claiming the Bakaevoo and Kurgashlya poles (BK and KF) to be poorly dated. As in the previous configuration (Fig. 4i), this reconstruction also places a large ocean between Amazonia and the remaining blocks of West-Central Gondwana (the Clymene Ocean), but now the eastern margin of Laurentia also faces an open ocean (which, for simplicity, we still refer to as the ‘West Iapetus Ocean’ despite Amazonia not being the conjugate margin). Unlike the previous two solutions, this reconstruction requires more than just an ocean opening event to be linked back to one of the 615 Ma reconstructions where Laurentia and Baltica were united; some strike slip between Laurentia and Amazonia would be required to make room for the restoration of Baltica.

A final set of comparative reconstructions from Ediacaran paleomagnetic data can be made at ~550 Ma, when data are available from Laurentia, Baltica and West Africa (Fig. 3h). As at 575–565 Ma, the paleolatitudes imposed by these data suggest that an ocean separated Laurentia and West Africa in latest Ediacaran time, but because the position of Amazonia is again unconstrained, that ocean could either be the West Iapetus (if Amazonia was affixed to West Africa; Fig. 4k) or the Clymene Ocean (if Amazonia remained attached to Laurentia; Fig. 4l). With reference to the mid-Ediacaran model of Wen et al. (2020) (Fig. 4j), Amazonia could also be restored to the Scotland-Ireland segment of the Laurentian margin at this time, in which case the ‘West Iapetus Ocean’
Alternative paleogeographic reconstructions at 615 Ma (a-g), 575 Ma (h-j) and 550 Ma (k-m) and their compatibility with paleomagnetic data. The alternative 615 Ma reconstructions present various states that could have evolved from different restorations of Rodinia (i.e. different pre-Ediacaran starting points). Panels a-c depict possible 615 Ma derivatives from the Rodinia fit of Li et al. (2008); panel d derives from the combined Rodinia fit of Dalziel (1992) and Torsvik et al. (1996); panels e-g show possible 615 Ma derivatives from more controversial Rodinia reconstructions (from Hartz and Torsvik (2002), Slagstad et al. (2019) and Evans (2009), respectively). The same 615 Ma paleomagnetic poles (Table 1) are restored according to the paleogeography in each panel. In panel c, the black dot highlights the pole of relative rotation of West Gondwana and Baltica with respect to Laurentia implied by the evolution from the starting Rodinia fit of Li et al.
For the 575 Ma reconstructions (panels h-j), the relative positions of the colored continents (all except Amazonia and São Francisco) are restored by fitting the two paleomagnetic directional groups simultaneously (lower panels; see main text). The alternative states show various positions of Amazonia (paleomagnetically unconstrained at this time) and an alternative positioning of Baltica (in panel j). The lower panels show the correspondingly reconstructed paleomagnetic data (Table 1). The dashed arrow highlights the rapid apparent polar shift between 575 and 565 Ma. The alternative 550 Ma reconstructions (panels k-m) show different states that could have evolved from the 575 Ma reconstructions (panels h-j), that are also consistent with paleomagnetic data. Lower panels again show the correspondingly reconstructed paleomagnetic data (Table 1). Continent abbreviations in all panels are the same as in Fig. 2 and pole abbreviations are as in Table 1. The color of each pole’s 95% confidence ellipse corresponds to the color of its host continent; continents in grey have no paleomagnetic constraints at the given time.
and Clymene Oceans would have effectively represented the same basin (Fig. 4m). Another observation, independent of the position of Amazonia, is that the ~550 Ma paleomagnetic data from Baltica restore it to an orientation that is strongly rotated with respect to the relative framework of the major continents (Laurentia, Baltica and West Gondwana) at 565 Ma. Given that the associated paleomagnetic excursion appears unique to Baltica (unlike the previous two excursions), this could express a rapid azimuthal rotation of Baltica alone. On the other hand, it is curious to note that the orientation of Baltica at 575 Ma is highly similar to that at 550 Ma, which could alternatively imply that this last excursion is again a global effect that has not yet been recognized elsewhere; we will return to this possibility in the discussion section.

In summary, the paleomagnetic data are sparse and consistent with several possible configurations of the supercontinent Rodinia and a range of timings for the opening of the West and East Iapetus oceans. The paleomagnetic constraints are strongest from Laurentia and Baltica, and together they indicate that the East Iapetus Ocean may not have opened before 615 Ma but also that it was wide by 575 Ma. Paleomagnetic inferences on the opening timing of the West Iapetus Ocean are strongly dependent on whether the cratons of West Africa and Amazonia were linked during the Ediacaran, as there are no reliable paleomagnetic data from Amazonia then. If those continents were coherent, the West Iapetus Ocean must have opened earlier than 575 Ma, and possibly before 615 Ma. If Amazonia was not attached to West Africa in the Ediacaran, an opening of the West Iapetus Ocean as young as early Cambrian is paleomagnetically permissible. A key observation is thus that the paleomagnetic data alone are not sufficiently informative to allow robust conclusions to be drawn about the nature and timing of Iapetus Ocean opening. We therefore need to turn to the geologic record to look for additional information.

4. Geologic constraints

In this section, we present a review of the Neoproterozoic to Cambrian geology of the main continental margins thought to be involved in the opening of the Iapetus Ocean (Fig. 5). These comprise the eastern margin of Laurentia (eastern North America and Scotland-Ireland; section 4.1), the western margin of Baltica (section 4.2), the western margin of Amazonia (section 4.3) and Rio de la Plata (section 4.4), in addition to some potentially key exotic/displaced terranes (section 4.5). It is important to note that this selection of margins follows from the insights drawn from the previous sections, but is not exhaustive; other margins (e.g. eastern Greenland, eastern Baltica) could also have played a significant role in the evolution of the Iapetan system according to a minority of the reconstructions considered above (Fig. 4). Similarly, while peripheral tectonic systems (e.g. Tornquist and Clymene oceans) are also relevant to a fuller understanding of the Iapetus, we have elected to maintain our focus on those margins thought to be most central to the Iapetus system itself. An integrated analysis of the surrounding regions therefore remains an important opportunity for future work. In the following (main text), we present brief summaries of the most salient geological observations from each of the selected margins, together with our associated interpretations of them—but much more comprehensive descriptions, more detailed stratigraphic illustrations and many additional references are provided in Appendix B.

4.1. Laurentia and peri-Laurentian terranes

The Iapetan margin of eastern Laurentia developed upon the late Mesoproterozoic/early Neoproterozoic Grenville orogen and was later reworked during multiplePaleozoic orogenic events (e.g. Taconic, Acadian, Alleghanian orogenies). In the context of Iapetan tectonics, three principal realms can be defined in the Appalachian-Caledonian orogen (Fig. 6) (Hibbard et al., 2007): (1) a ‘Laurentian realm’ which represents parautochthonous rocks initially deposited on the Laurentian craton and later thrust during subsequent Paleozoic orogenesis, (2) a ‘peri-Gondwana realm’ that comprises rocks formed along or close to the Gondwanan margin that were later accreted to Laurentia in the Paleozoic, and (3) an ‘Iapetan realm’ that comprises units likely formed in the Iapetus Ocean and then caught between Laurentia and the exotic (Gondwanan) terranes colliding with it during polyphase Paleozoic orogenesis. Vestiges of the Iapetan margin of Laurentia are thus found in the Laurentian realm, and so it will be our focus in the following. To consider any important lateral distinctions along the extensive Iapetan margin of eastern Laurentia, and to ease our review of its Neoproterozoic to Cambrian geology, we break the margin into four main geographic segments (Fig. 6): the northern Appalachians (‘Northern Segment’, 4.1.1), the central and southern Appalachians (‘Central Segment’, 4.1.2), the Ouachita orogen (‘Southern Segment’, 4.1.3) and Scotland-Ireland (4.1.4). A final sub-section (4.1.5) is dedicated to a hypothesized peri-Laurentian ribbon terrane which geographically may have spanned most of the Iapetan margin of eastern Laurentia. Expanded descriptions of each of these segments are provided in section 1 of Appendix B.

4.1.1. The Northern Segment (northern Appalachians)
The Northern Segment runs from Labrador and Newfoundland in the northeast, through Quebec and Vermont, to southern New York and northern Pennsylvania/New Jersey in the southwest. In this segment, vestiges of the Iapetan margin of Laurentia are found locally on the Laurentian autochthon, but thicker, more complete successions are preserved in the Appalachian orogen to the east, i.e. in the Laurentian parautochthon, and in Taconic allochthons emplaced structurally on top of it (Humber Zone; Fig. 6). Both the parautochthon and the Taconic allochthons consist of late Neoproterozoic to Ordovician successions that were originally deposited on the Laurentian continental shelf and slope, but were later thrust over the craton during the Taconic orogeny. Those late Neoproterozoic to Ordovician sedimentary successions have conventionally been described as two main assemblages unconformably overlying the ‘Grenvillian’ Laurentian basement (e.g., Williams and Hiscott, 1987; Appendix B). The lower assemblage is characterized by coarse siliciclastic units that exhibit abrupt lateral changes in both
Fig. 6. Map of the eastern margin of North America and the British Isles (restored to their approximate relative position before Atlantic opening) showing the Iapetan rifted-margin of eastern Laurentia. Names and ages (select ages listed in Ma) of magmatic data points detailed in Appendix B. Abbreviations: CF = Catoctin Fm; CS = Central Segment; GD = Grenville Dykes; GF = Grenville front; LRD = Long Range dykes; M = Marathon volcanic clasts; NS = Northern Segment; SI = Scotland-Ireland; SOM = Southern Oklahoma magmatism; TF = Tayvallich Fm. North America figure modified from Thomas (2006), Hatcher et al. (2007), McCausland et al. (2011), Tremblay and Pinet (2016), Hanson et al. (2016), Domeier (2016) and Thomas et al. (2017); British Isles figure modified from Oliver et al. (2008), Fettes et al. (2011) and Domeier (2016).

Fig. 7. Simplified stratigraphic/magmatic/tectonic framework of the Iapetan margins of Laurentia. Subcolumns 1, 2 and 3 in each segment of the margin depict: schematic stratigraphy (1), magmatic activity (2) and interpreted phases of rifting and drifting (3). Abbreviations: AG = Argyll Gr; CF = Catoctin Fm; CG = Chilhowee Gr; FLS = Fleur de Lys Supergroup; GD = Grenville Dike Swarm; GG = Grampian Gr; LRD = Long Range Dikes; NDA = Notre Dame arc; OS = Ocoee Supergroup; SH = Southern Highlands Gr; TG = Trossachs Gr; TF = Tayvallich Fm.
facies and unit thickness, and which are interbedded with volcanic rocks and intruded by mafic dikes and plutons. They are interpreted as syn-rift deposits, and isotopic ages derived from the volcanic and plutonic rocks range from ~615 Ma to 550 Ma (Fig. B1) (Cawood et al., 2001; Kamo et al., 1989; Williams, 1995). Overlying those syn-rift rocks, the upper assemblage consists of laterally uniform sequences of shallow marine limestones, shales and mature sandstones that are neither interbedded with volcanic rocks, nor intruded by the dikes and plutons that invade the lower assemblage. This upper assemblage is interpreted as a passive margin (drift) sequence following rifting (Cawood et al., 2001; Williams, 1995). Fossils from some of the lower units of the upper assemblage indicate that its deposition began at least by late early Cambrian time, and passive margin sedimentation continued into the Ordovician. The boundary separating the two assemblages, which is thus interpreted as a rift-to-drift transition marking a continental breakup event, is placed in early Cambrian time, at ~540–530 Ma (Figs. 7, B1).

The first appearance of magmatism in this segment, at ~615 Ma, occurred contemporaneously with regional faulting, as reflected by the formation of pseudotachylytes along basement faults in the Quebec salient (O’Brien and van der Pluijm, 2012). The earliest phase of magmatism mostly occurred as mafic dikes (e.g. Long Range Dikes) and minor plutons restricted to Newfoundland and Labrador, but after ~590 Ma plutonism became more widespread along the Northern Segment (Figs. 6, 7, B1) (Kamo et al., 1995; McCausland et al., 2011; Miller and Barr, 2004; Volkert et al., 2015). After ~570 Ma, bimodal volcanism appeared and continued until the latest Ediacaran (Walsh and Aleinikoff, 1999). Collectively, these Ediacaran magmatic rocks have been described as a magmatic province, referred to as the ‘Central Iapetan Magmatic Province’ (CIMP; Puffer, 2002; Tegner et al., 2019; Ernst and Buchan, 1997).

Regionally, the orientation of the Ediacaran dike swarms appear to converge at a point near the Sutton Mountains in the Quebec salient (Fig. 6), and that pattern has been interpreted as the relic of a plume-related triple junction (Burke and Dewey, 1973; Seymour and Kumarapeli, 1995). According to that interpretation, two of the rift arms were later exploited during Iapetus Ocean opening in the Northern Segment, and the Ottawa graben represents the third, failed rift axis. Geochemically, Puffer (2002) classified the Ediacaran magmatic rocks into two groups: a generally older group (~615–565 Ma) resembles continental flood basalts and is ascribed to a mantle plume source mixed with substantial contributions from the subcontinental lithosphere. A younger (~555–550 Ma) and temporally more restricted group is of ocean island basalt (OIB) affinity, and is interpreted to have been derived directly from a mantle plume source (Puffer, 2002; Volkert et al., 2015). That geochemical evolution could be explained by the action of a single plume—with initial flood basalt magmatism involving abundant lithospheric melting before the plume was able to reach shallow crustal levels directly.

Thus, one permissible interpretation is that an early Ediacaran mantle plume impinged upon and weakened the lithosphere along the Northern Segment, ultimately facilitating the continental breakup that evidently occurred there in earliest Cambrian time (Puffer, 2002). As it will be relevant to our discussion later, we note that there are no indications that a passive margin formed along this segment prior to early Cambrian time (~540–530 Ma).

4.1.2. The Central Segment (southern Appalachians)

The Central Segment spans from Pennsylvania in the northeast to northeastern Alabama in the southwest (Fig. 6). In this segment, the Allegheny-Cumberland Plateau represents the Laurentian autochthon, flanked to the east by the parautochthonous Valley and Ridge Province (foreland fold-and-thrust belt) and the Blue Ridge Province. The Blue Ridge Province consists of a west-vergent stack of thrust sheets that contain crystalline basement massifs and multiply deformed late Neoproterozoic to Paleozoic meta-sedimentary and igneous rocks. Rocks of the western Blue Ridge Province (WBR) are thought to be parautochthonous with respect to Laurentia, whereas the eastern Blue Ridge Province and the Inner Piedmont further east may have been dislodged from Laurentia in the latest Neoproterozoic–Cambrian and later re-accreted to it in the Ordovician (Hibbard et al., 2007; Merschat, 2009; section 4.1.5 and Appendix B).

In contrast to the Northern Segment, relics of the Iapetan margin along the Central Segment extend backward into the early Neoproterozoic (Figs. 7, B2). The oldest of those rocks are late Tonian and Cryogenian plutonic rocks and volcaniclastic successions that occur in the WBR. In the central WBR (in northern North Carolina and southern Virginia), late Tonian plutonic rocks include A-type granitoids and mafic dikes that range in age from ~765 to 730 Ma (Figs. 6, 7, B2) (Fetter and Goldberg, 1995; Ownty et al., 2004; Su et al., 1994). Broadly contemporaneous volcanic-siliciclastic successions occurring in the same area include bimodal volcanic rocks that have been dated to ~760–740 Ma, but with volcanic clasts among interbedded sediments that extend back to ~780 Ma (Mccelland and Holm-Denoma, 2014). The clastic components of those successions are characterized by coarse siliciclastic rocks that exhibit abrupt lateral changes in facies and age, and are interpreted as syn-rift deposits (Mccelland and Gazel, 2014). Geochemically, Mccelland and Gazel (2014) identified the bimodal volcanic rocks as products of intraplate magmatism derived from an enriched mantle, suggesting that rifting and anorogenic magmatism may have been related to the arrival of a mantle plume. To the north (in the northern WBR of central Virginia to Maryland), anorogenic magmatism appeared later in the Tonian (at ~745 Ma) and continued into the late Cryogenian (~653 Ma), although it was regionally most voluminous between ~750 and 700 Ma (Figs. 7, B2) (Fokin, 2003; Tollo et al., 2004).

After a hiatus of ~100 Myr, a second pulse of Neoproterozoic magmatism occurred in the northern WBR at ~572–563 Ma, with the emplacement of extensive volcanics and dikes of the Catoctin Fm. (Figs. 6, 7, B2) (Aleinikoff et al., 1995). Emplacement of the Catoctin Fm. is thought to have been related to a regional rifting event, which in the southern WBR was associated with the deposition of the thick, synrift Ocoee Supergroup. The start of that rifting, as defined by the base of the Ocoee Supergroup, is only loosely temporally constrained and could be as old as Cryogenian, but it certainly continued into the latest Ediacaran (~653 Ma), although it was regionally most voluminous between ~750 and 700 Ma (Figs. 7, B2) (Fokin, 2003; Tollo et al., 2004).

The Ediacaran to Cambrian evolution of the Central Segment strongly resembles that observed in the Northern Segment: Ediacaran rifting followed by an early Cambrian (~540–520 Ma) rift-to-drift transition and the subsequent establishment of a late early Cambrian to Ordovician passive margin (Fig. 7). Puffer (2002) moreover included the late Ediacaran Catoctin Fm. as part of his ‘older’ flood basalt group (together with dike swarms from the Northern Segment), which he attributed to a mantle plume. However, in contrast to that common Ediacaran–Cambrian evolution, the earlier phase of late Tonian to Cryogenian rifting and magmatism (mostly concentrated between ~750–700 Ma) of the Central Segment does not have a known analog in the north. The apparent absence of that earlier phase in the Northern Segment has led to a prevailing interpretation that that earlier rifting failed to proceed, to dislocate the opening of an ocean basin (Aleinikoff et al., 1995). Although the late Tonian–Cryogenian intraplate magmatic rocks have also been interpreted to relate to a mantle plume, the break in magmatism between the late Tonian–Cryogenian and late Ediacaran pulses (as well as the total length of time they span together) makes it challenging to link them to a common source.

4.1.3. The Southern Segment (Ouachita orogen)

The Southern Segment of Laurentia’s Iapetan margin runs from
Alaska to the Marathon region of west Texas, and is covered by late Paleozoic Ouachita-Appalachian allochthons and post-orogenic Atlantic-Gulf passive-margin deposits. Structurally, the Ouachita and southern Appalachian orogens trace out a conspicuous first-order architecture with prominent salients and recesses that Thomas (2011) has interpreted to reflect the underlying morphology of the Laurentian margin, inherited from Iapetus rifting. This includes a pair of principal NE–SW-trending rift zones separated by large-scale, NW–SE-trending transforms. Several large, intracratonic fault systems that are thought to relate to that rifting framework are also observed to branch from it: namely the NE–SW-oriented (rift parallel) Birmingham and Mississippi Valley grabens and the NW–SE-oriented (transform parallel) Southern Oklahoma fault system (Fig. 6) (Thomas, 1991).

The earliest clues of post-Grenvillian rifting along the Southern Segment are indirect, and gleaned from volcanic clasts contained in Ordovician turbidites found in thrust sheets of the Ouachita fold-and-thrust belt in western Texas (Hanson et al., 2016). The clasts are dated to ~706 Ma and have been geochemically identified as intraplate, alkaline products derived from an OIB-type source (Figs. 6, B3). Although the data are sparse, they invite a tentative correlation to the ~750–700 Ma anorogenic magmatic pulse observed in the Central Segment (Fig. 7) (Appendix B).

Following that poorly-resolved Cryogenian event, magmatism reappeared in this region in the late Ediacaran–early Cambrian (~580–530 Ma) in the Southern Oklahoma fault system (Figs. 6, 7, B3). Magmatism was associated with the intrusion of a large tholeiitic layered mafic complex that was subsequently invaded and covered by a type bimodal plutonic and volcanic rocks (Brueseke et al., 2016; Hanson et al., 2013). Rifting evidently occurred contemporaneously with that magmatic phase, as the late Ediacaran layered mafic complex was tilted prior to its invasion by earliest Cambrian dikes. Although the details remain vague, a possibly similarly-aged sequence of early(?)? Cambrian volcaniclastic rocks has been reported from a drill-core in the Marathon region of southwestern Texas, which may testify to a wider distribution of early Cambrian magmatism than is presently known (Nicholas and Rozendal, 1975; Rodriguez et al., 2017).

Unlike in the Central Segment, the rift-to-drift transition has not been identified anywhere along the Southern Segment. Across most of the margin, the oldest passive margin successions are middle to late Cambrian in age, although in the east, late early Cambrian sequences are recognized in isolated thrust blocks and inferred from boreholes (Fig. B3) (Mack, 1980; Thomas, 1991). This is consistent with an early Cambrian rift-to-drift transition as seen in the Central Segment (Fig. 7). However, a brief incursion of red beds and evaporites into the late early Cambrian sequence has been suggested to mark a distinct extensional event that is furthermore reflected by the continuation of faulting in the Mississippi Valley and Birmingham grabens into the middle Cambrian (Fig. B3) (Thomas, 2011; Read and Repetski, 2012). That extensional event has been ascribed to the rifting of the Precordillera terrane—which now lies in southwestern South America (section 4.5.3; Appendix B)—from the Ouachita embayment (Thomas, 2011; Thomas and Astini, 1999). That interpretation is supported by the occurrence of syn-rift successions in the Precordillera terrane that resemble those of the Birmingham graben (Thomas, 2011; Thomas and Astini, 2003), as well as by Cambrian faunas and paleomagnetic data from the Precordillera terrane (Benedetto, 2004; Rapalini and Astini, 1998). The Precordillera terrane (as it can be delineated at present-day) furthermore fits into the Ouachita embayment when the margin is palinspastically reconstructed.

4.1.4. Scotland-Ireland

The Iapetan margin that developed along the eastern margin of Laurentia in the Neoproterozoic to Cambrian extended beyond North America to Scotland-Ireland, which was attached to North America prior to the opening of the Atlantic Ocean (Torsvik et al., 2012). Vestiges of the Iapetan rifted margin are found in the Hebridean, Northern Highlands and Grampian terranes that are located in northwestern Ireland, Northern Ireland and northern Scotland (Fig. 6) (Appendix B). These terranes are generally considered to have been proximal to one another since the Archean–Paleoproterozoic as they likely share a similar base- ment of that age (Chew and Strachan, 2014; Dewey et al., 2015), but they could also have assembled later, during the Grenville orogeny (Strachan et al., 2020). This segment of the Iapetan rifted margin of Laurentia was later reworked during the Caledonian orogeny, which marked the closure of the Iapetus Ocean (McKerrow et al., 2000).

Details of the Iapetan rifting history in Scotland and Ireland are mainly preserved in the Grampian terrane, where they are recorded in the Dalradian Supergroup, which comprises thick mid-Neoproterozoic to Early Ordovician sequences of mainly clastic sediments (Stephenson et al., 2013). Deposition of the supergroup started after 800 Ma (Cawood et al., 2015; Robertson and Smith, 1999), and it recorded two possibly distinct phases of rifting (Figs. 7, B4). The first phase, recognized by an abrupt change in sedimentary facies from fluviatile/shallow-marine to deep-water turbidites (Stephenson et al., 2013), probably occurred in the late Tonian, but is only poorly bracketed in age between ~800 and 720 Ma. However, that upper age constraint is based on controversial correlations between stratigraphically distinct glacial deposits in the Dalradian Supergroup and global-scale Neoproterozoic glaciations (Appendix B); alternative correlation schemes could allow the first phase of rifting to be as young as late Cryogenian (Figs. 7, B4) (Fairchild et al., 2018; Prave et al., 2009; Rooney et al., 2011; Stephenson et al., 2013). After an interval of quiet shelf deposition following the first rifting phase, the second phase of rifting was also characterized by an abrupt change to deep-water facies and was accompanied by widespread magmatism (Fig. B4) (Fettes et al., 2011). On the basis of sparse radiometric ages from lavas and intrusions associated with rifting (Dempster et al., 2002; Halliday et al., 1989), as well as additional glacial correlations (Appendix B), that younger phase of rifting is estimated to have occurred during the early to mid-Ediacaran (~635–580 Ma) (Fig. 7).

Magmatism in the Dalradian Supergroup was mainly expressed as subaqueous basaltic flows and mafic sills with N-MORB to E-MORB geochemical affinities (Fettes et al., 2011), and as rift-related, A-type granitoids which also invaded the Northern Highlands terrane (Tanner et al., 2006). Magmatism possibly started at ~650 Ma and became more widespread by ~635 Ma (Fettes et al., 2011), but the most voluminous pulse occurred at ~600–590 Ma (Dempster et al., 2002; Halliday et al., 1989), when granitoids were also intruding the Northern Highlands terrane (Figs. 6, B4) (Oliver et al., 2008). Magmatism is thought to have ceased by ~570 Ma (Fettes et al., 2011), although clear upper age constraints are lacking.

The rift-drift transition in Scotland-Ireland is generally proposed to have occurred when magmatism culminated at ~600–590 Ma (Stephenson et al., 2013). That was followed by the deposition of a thick succession of deep-water turbidites that characterize the upper Dalradian Supergroup and which are interpreted as slope and rise deposits formed along the passive margin of Laurentia (Figs. 7, B4) (Tanner and Sutherland, 2007; Appendix B). The upper Dalradian successions pass up into strata of Cambrian age that contain deep water trilobites, which demonstrate their time-equivalence with Cambrian successions of the Caledonian foreland found in the Hebridean terrane (Rushon et al., 2011). The latter successions are shallow-marine shelf sediments which were deposited further inboard the margin. Note that although there are many clear similarities between this history and that of the eastern margin of North America, the rift-drift transition estimated here is distinctly older (Fig. 7).

4.1.5. Taconic ribbon continent

To the east of its late Neoproterozoic-early Paleozoic autochthon and paraautochthon, the eastern margin of Laurentia is rimmed by a discontinuous series of continental slivers that are generally presumed to be of Laurentian derivation, and which may have spilled off Laurentia in the late Ediacaran to early Cambrian and then re-accreted to it in the Ordovician (Taconic-Grampian orogenies) (Cawood et al., 2001;
Hibbard et al., 2007; Waldron and van Staal, 2001). These terranes broadly include the Dashwoods terrane of Newfoundland, scattered masses in the ‘Piedmont terrane’ of the southern Appalachians and continental fragments south of the Grampian terrane in Scotland-Ireland (Fig. 6) (Appendix B). Here we use the term ‘Taconic ribbon continent’ to refer to these continental slivers collectively, but whether those terranes were actually part of a single coherent continental fragment or were a loose collection of independent blocks is not clear.

The Dashwoods terrane is inferred to be floored by Laurentian basement, but is characterized by medium to high-grade metasedimentary rocks whose protoliths are thought to be of Ediacaran-Cambrian age and correlative with sedimentary successions on the Laurentian parautochthon (Van Staal et al., 2007; Whalen et al., 1997; Williams, 1995; Cawood et al., 1995). However, in contrast to the Laurentian margin of the Northern Segment, which appears to have remained passive until ~475 Ma, the metasedimentary cover of the Dashwoods terrane was metamorphosed and overthrusted by an oceanic tract before the start of the Ordovician, and the terrane was invaded by subduction-related arc plutons from ~490 Ma (Figs. 7, B5) (Waldron and van Staal, 2001). To reconcile this discrepancy, Waldron and van Staal (2001) proposed that the Dashwoods terrane rifted from Laurentia in the latest Ediacaran-early Cambrian, opening a marginal seaway (here called the ‘Taconic seaway’). Shortly thereafter, that marginal seaway was subducted beneath the Dashwoods terrane, resulting in the late Cambrian to Ordovician arc magmatism observed there, and leading to its collapse back against the Laurentian margin during the Taconic orogeny. Beyond Newfoundland, a similar tectonic history can be inferred from masses of the ‘Piedmont terrane’ in the southern Appalachians that are also thought to be floored by Laurentian basement but include Early Ordovician arc magmatic rocks (Fig. B5) (Hibbard et al., 2007; Horton et al., 2010; Merschat, 2009; Appendix B). Likewise in Scotland-Ireland, continental fragments along the south of the Grampian terrane may have rifted from the Laurentian margin in the late Ediacaran to Cambrian, collided with an intraoceanic arc system in the Cambro-Ordovician and then re-accreted to the Laurentian margin in late Early-Middle Ordovician time (Chew et al., 2008a, 2008b; Chew and Strachan, 2014; Hollis et al., 2013; Appendix B). Additionally, the discovery of a juvenile, late Cambrian calc-alkaline igneous complex within thrust nappes of northern Scotland suggest that part of the Grampian and Northern Highlands terranes themselves may represent a displaced continental fragment and possible Taconic ribbon continent correlative (Dunk et al., 2019).

One possible interpretation of those observations is thus that the early Cambrian rift-drift transition and middle Cambrian-Early Ordovician passive margin succession observed along the Laurentian (para)autochthon marked the opening of the Taconic seaway and not the Iapetus Ocean. However, assuming that the main Iapetus Ocean opened prior to that seaway, this raises an important question: why would subduction related arc plutons of ~490 Ma be found here? To reconcile this discrepancy, Waldron and van Staal (2001) proposed that the Dashwoods terrane rifted from Laurentia in the latest Ediacaran-early Cambrian, opening a marginal seaway (here called the ‘Taconic seaway’). Shortly thereafter, that marginal seaway was subducted beneath the Dashwoods terrane, resulting in the late Cambrian to Ordovician arc magmatism observed there, and leading to its collapse back against the Laurentian margin during the Taconic orogeny. Beyond Newfoundland, a similar tectonic history can be inferred from masses of the ‘Piedmont terrane’ in the southern Appalachians that are also thought to be floored by Laurentian basement but include Early Ordovician arc magmatic rocks (Fig. B5) (Hibbard et al., 2007; Horton et al., 2010; Merschat, 2009; Appendix B). Likewise in Scotland-Ireland, continental fragments along the south of the Grampian terrane may have rifted from the Laurentian margin in the late Ediacaran to Cambrian, collided with an intraoceanic arc system in the Cambro-Ordovician and then re-accreted to the Laurentian margin in late Early-Middle Ordovician time (Chew et al., 2008a, 2008b; Chew and Strachan, 2014; Hollis et al., 2013; Appendix B). Additionally, the discovery of a juvenile, late Cambrian calc-alkaline igneous complex within thrust nappes of northern Scotland suggest that part of the Grampian and Northern Highlands terranes themselves may represent a displaced continental fragment and possible Taconic ribbon continent correlative (Dunk et al., 2019).

In the most general terms, the Scandinavian Caledonides can be divided into three segments (‘Southern’, ‘Central’ and ‘Northern’) along the strike of the orogen (Fig. 8), which are defined by prominent changes in the architecture of the nappe stack. The Scandinavian allochthons include several Proterozoic continental-derived metasedimentary complexes, which carry most of the rock units interpreted to have witnessed the formation of the Iapetus margin, whereas rift-related rocks in the Baltic autochthon are scarce. It is also important to keep in mind that some of the nappe complexes that are interpreted to include rocks from the Iapetan rifted margin are commonly but not universally accepted to be of Baltic origin (e.g. Corfu et al., 2007; Kirkland et al., 2007). Because of the lack of unequivocal evidence for an exotic origin, we prefer to interpret these units as having originated on Baltica.

### 4.2. Baltica

Baltica is composed of several Archean–Paleoproterozoic cratonic blocks that in the west are wrapped by mostly Meso–Neoproterozoic magmatic and metamorphic domains, i.e. the domains of the Sveco-norwegian orogen. These domains underlie much of today’s Scandinavia and in the west have been thoroughly reworked during the Silurian–Devonian Scandian Orogeny of the Caledonian orogenic cycle. The rocks in Scandinavia that have been affected by or were emplaced over the Baltic margin during the Scandian Orogeny are commonly referred to as the Scandinavian Caledonides.

In the most general terms, the Scandinavian Caledonides can be divided into a (para)autochthonous basement with a cover and an intricate stack of nappes that has been thrust over the western Baltic cratonic margin (Fig. 8) during the Scandian Orogeny. The autochthon underlining the nappe stack experienced a westward increase in Scandinavian deformation and metamorphism from the poorly affected rocks in the foreland towards the (UK)HP metamorphic domains (e.g. the Western Gneiss Region) in the hinterland. The autochthon and allochthon of the Scandinavian Caledonides were subsequently reworked by Devonian and Mesozoic extensional tectonics and the opening of the Atlantic.

The Scandinavian Caledonides can further be divided into three segments (‘Southern’, ‘Central’ and ‘Northern’) along the strike of the orogen (Fig. 8), which are defined by prominent changes in the architecture of the nappe stack. The Scandinavian allochthons include several Proterozoic continental-derived metasedimentary complexes, which carry most of the rock units interpreted to have witnessed the formation of the Iapetus margin, whereas rift-related rocks in the Baltic autochthon are scarce. It is also important to keep in mind that some of the nappe complexes that are interpreted to include rocks from the Iapetan rifted margin are commonly but not universally accepted to be of Baltic origin (e.g. Corfu et al., 2007; Kirkland et al., 2007). Because of the lack of unequivocal evidence for an exotic origin, we prefer to interpret these units as having originated on Baltica.

#### 4.2.1. South Scandinavian Caledonides

The autochthon in the Southern Segment consists of mostly Meso-proterozoic ortho- and paragneisses that locally have been intruded by granitoïds ranging in age between ~980–910 Ma (Figs. 8, B6) (Bingen et al., 2008; Bingen and Solli, 2013). In southern Norway, the autochthon was intruded by the mafic Hunnedalen dyke complex at ~855 Ma (Walderhaug et al., 1999) and by ultramafic lamprophyre dykes at ~686 Ma (Zozulya et al., 2020). Magmatic rocks interpreted to be directly related to the opening of the Iapetus Ocean are scarce in the autochthon but include the ~616 Ma mafic Egersund dyke swarm in southern Norway (Bingen et al., 1998) and the ~584 Ma Fen and Alnö carbonatite complexes in Norway and Sweden, respectively (Figs. 8, 9, B6) (Meert et al., 2007; Meert et al., 1998). The geochronological transition to alkali Fersunds dyke swarm in southern Norway (Bingen et al., 1998) and the ~584 Ma Fen and Alnö carbonatite complexes in Norway and Sweden, respectively (Figs. 8, 9, B6) (Meert et al., 2007; Meert et al., 1998). The geochemically transitional to alkali Fersunds dyke swarm in particular have been linked to the Central Iapetus Magmatic Province (CIMP) (Tegner et al., 2019) and their emplacement is inferred to have occurred under a regime of sinistral transtension on the basis of their magnetic fabrics (Montalbano et al., 2016). Neoproterozoic metasediments that were unconformably deposited on the Baltic basement include basal glacial deposits and are referred to as the Moelv tillite (e.g. Nystuen et al., 2008). These glacial deposits are overlain by shales and shallow-marine sandstones within which the Ediacaran-Cambrian boundary is preserved (Fig. 9, B6).

In the allochthons, the lowest allochthonous levels preserve Neoproterozoic–early Paleozoic metasediments. These successions typically have kilometre-thick, fluvial–alluvial, continentally derived, coarse-
selected crystallization ages of Proterozoic rocks in the Baltic autochthon and in the Scandinavian nappe complexes are also shown. Abbreviations: C = Scandinavian Dyke Complex at Corrovarre (Kjell et al., 2019a, 2019b), granitic bodies at Corrovarre (Kjell et al., 2019a) and Rhapvsvarri granitic gneiss (Corfu et al., 2007); ED = Egersund Dykes (Bingen et al., 1998); EG = Edvdvøgdei Granite (Kirkland et al., 2008); EL = Leucosome in Edvdvøgdei Gneiss (Corfu et al., 2007); FCC = Fen Carbonatite Complex (Walderhaug et al., 1999); GG = Grimstad Granite (Kullerud and Machado, 1991); HD = Hunnedalen Dykes (Walderhaug et al., 1999); HG = Havlo Granite (Corfu, 1980); KD = Scandinavian Dyke Complex at Kebnekaise (Baird et al., 2014; Paulsson and Andræsson, 2002); OD = Scandinavian Dyke Complex at Otlfjølet (Kumpulaieni et al., 2021); OG = Ormsfall Granodiorite (Tucker et al., 1990); RG = Repvøg granite (Kirkland et al., 2006); SA = Scandinavian Dyke Complex at Sarek (a, Root and Corfu, 2012, Svenningsen, 2001); SIP = Seland Igneous Province (Pedersen et al., 1989; Roberts et al., 2010); SL = Leucosomes in Snøroy succession (Gasser et al., 2015); TG = Tovdal Granite (Andersen et al., 2002); UML = ultramafic lamprophyre at Vinoren (Zozulya, 2020); VMC = Vistas Magmatic Complex (Paulsson and Andræsson, 2002).

Fig. 8. Map of the Scandinavian Caledonides showing units interpreted to preserve rocks of the Iapetus margin (see main text and Appendix B). Selected crystallization ages of Proterozoic rocks in the Baltic autochthon and in the Scandinavian nappe complexes are also shown. Abbreviations: C = Scandinavian Dyke Complex at Corrovarre (Kjell et al., 2019a, 2019b), granitic bodies at Corrovarre (Kjell et al., 2019a) and Rhapvsvarri granitic gneiss (Corfu et al., 2007); ED = Egersund Dykes (Bingen et al., 1998); EG = Edvdvøgdei Granite (Kirkland et al., 2008); EL = Leucosome in Edvdvøgdei Gneiss (Corfu et al., 2007); FCC = Fen Carbonatite Complex (Walderhaug et al., 1999); GG = Grimstad Granite (Kullerud and Machado, 1991); HD = Hunnedalen Dykes (Walderhaug et al., 1999); HG = Havlo Granite (Corfu, 1980); KD = Scandinavian Dyke Complex at Kebnekaise (Baird et al., 2014; Paulsson and Andræsson, 2002); OD = Scandinavian Dyke Complex at Otlfjølet (Kumpulainen et al., 2021); OG = Ormsfall Granodiorite (Tucker et al., 1990); RG = Repvøg granite (Kirkland et al., 2006); SA = Scandinavian Dyke Complex at Sarek (a, Root and Corfu, 2012, Svenningsen, 2001); SIP = Seland Igneous Province (Pedersen et al., 1989; Roberts et al., 2010); SL = Leucosomes in Snøroy succession (Gasser et al., 2015); TG = Tovdal Granite (Andersen et al., 2002); UML = ultramafic lamprophyre at Vinoren (Zozulya, 2020); VMC = Vistas Magmatic Complex (Paulsson and Andræsson, 2002).

grained, silicilastic deposits at the base, which traditionally have been referred to as Sparagmites (Figs. 8, 9, B6) (Kumpulainen and Nystuen, 1985; Nystuen, 1987; Bingen et al., 2011; Bingen et al., 2005; Nystuen et al., 2008; Esmark, 1829). Detrital zircon age data indicate that these sediments have been sourced from the Baltic craton (e.g. Bingen et al., 2011) and may also indicate an Ediacaran maximum depositional age (Bingen et al., 2005; Lamminen et al., 2015). However, except for the 616 Ma Egersund dykes, there is a paucity of Ediacaran magmatism in the autochthon of the southern Scandinavian Caledonides and thus the source of the youngest detrital zircons in the Sparagmites remains uncertain. Locally, the successions are cut by mafic dykes and, at least in one location, the silicilastic deposits are overlain by a lava flow (Fig. B6) (Andresen and Gabrielsen, 1979; Furnes et al., 1983; Saether and Nystuen, 1981). Geochemically, the mafic rocks resemble continental tholeiites and have collectively been referred to as Sparagmite Basalts (Furnes et al., 1983). The kilometre-thick coarse-grained and locally dyked successions are commonly interpreted to represent a phase of major crustal thinning during the opening of the Iapetus Ocean (Nystuen et al., 2008). Locally, near the top of the successions are shallow marine carbonate and shale successions that interfinger with marine conglomerates. The conglomerates contain clasts derived from the underlying sediments, clasts of mafic magmatic rocks that are reminiscent of the Sparagmite Basalts, and clasts derived from other extra-basinal sources. Some clasts, derived from the underlying shales and carbonates, contain a rich, post-Marinoan, Ediacaran acritarch fauna (Adamson, 2016). Glacial deposits of the Moelv Fm. unconformably overlie the carbonate and shale formations as well as the deep-water sediments. The glacial deposits, in turn, are overlain by fluvial and shallow-marine sandstones and shales, in which the Ediacaran–Cambrian boundary is preserved (Nystuen et al., 2008; Saether and Nystuen, 1981). The Ediacaran–Cambrian shallow-marine sedimentary successions are interpreted as deposits of the rifting–drift transition. The available evidence (see also Appendix B) suggests that the onset of the rift-drift transition occurred after the Marinoan glaciation (~635 Ma) and before the deposition of the Moelv Tillite that is here interpreted to have been deposited contemporaneous with the Gaskiers glaciation (~580 Ma; Fig. 9, B6).

4.2.2. Central Scandinavian Caledonides

The autochthon in the Central Segment is mostly of Meso- to Paleoproterozoic age. Notably, there is a paucity of Neoproterozoic granitoid intrusions akin to the late Sveconorwegian granites in the Southern Segment. Unconformably overlying the basement gneisses are Ediacaran–early Palaeozoic silicilastic successions (Fig. 9) that preserve a transgressive sequence interpreted to have been deposited following ~580 Ma glaciations, and which are commonly referred to as the
Evidence for the formation of the Iapetus rifted margin in the Central Segment is preserved within two nappe complexes (Fig. 8). These are (from structurally lower to higher level): (1) Neoproterozoic–early Paleozoic coarse-elastic sedimentary successions that are virtually devoid of Ediacaran intrusives, e.g. the Risbåck Group, and (2) Cryogenian–Ediacaran coarse-elastic shallow-marine deposits that were intruded by mafic dykes in the Ediacaran (Fig. 8, 9), i.e. the Seve and Sárv nappe complexes (Baird et al., 2014; Gee et al., 2017; Gee et al., 2014; Kirsch and Svenningsen, 2016; Kjoll, 2020; Kjoll et al., 2019a; Svenningsen, 2001, 2007; Svenningsen, 1995; Svenningsen, 1994). The Neoproterozoic–early Paleozoic successions consist of extensive quartzites and meta-arkose units that are overlain by glacial deposits (e. g. Långmarkberg Fm.) and finer-grained sediments, e.g. the Sjoutavlen Gr. The latter contains the Ediacaran-Cambrian boundary (Gee and Stephens, 2020a; Gee et al., 1974; Nystuen et al., 2008; Kumpulainen and Greiling, 2011; Appendix B). These successions have been correlated with those in the Southern Segment (e.g. Kumpulainen and Nystuen, 1985) and are interpreted as syn-rift and rift-drift deposits. The Cryogenian–Ediacaran metasedimentary complexes occur predominantly in the Central Segment of the Scandinavian Caledonides (Gee et al., 1985; Gee and Stephens, 2020b; Jakob et al., 2019; Kjoll, 2020; Törnebom, 1896; Appendix B). The top and bottom of these successions are truncated by thrust faults and so the depositional substratum of the sedimentary successions is largely unknown and sediments at the top of the successions indicative of, e.g., a rift–rift succession are commonly not preserved. Locally, the Cryogenian–Ediacaran successions include a lower and an upper monotonous, shallow marine, sandstone–metaarkose unit, which are separated from each other by a carbonate and shale formation (Fig. 9, B6). The latter is locally associated with a detritic that is interpreted to have been deposited during the Marinoan glaciation (~635 Ma) (Nystuen et al., 2008; Kjoll, 2020; Svenningsen, 2007; Svenningsen, 1994). Detrital zircons from the siliciclastic successions above and below the diamicmites are no younger than ~700 Ma (Kjoll, 2020).

The Cryogenian–Ediacaran successions were intruded by the Scandinavian Dyke Complex (SDC) at about 608–596 Ma (Andréasson, 1994; Baird et al., 2014; Gee et al., 2017; Kjoll et al., 2019a, 2019b; Kumpulainen et al., 2021; Svenningsen, 2001; Tegner et al., 2019). Estimates of the pressure and temperature during the emplacement of the SDC throughout the Central Segment indicate that the dykes invaded the Cryogenian successions at a mid-crustal level, i.e. at pressures between 2.5 and 4.5 kbar (Kjoll et al., 2019a). Locally, granitic melts sourced from the metasediments have been dated to 612 Ma, indicating that the geotherm was already elevated before the final emplacement of the mafic dykes. The source of the mafic melts is suggested to have been linked to the arrival of a mantle plume, i.e. GIMP (Tegner et al., 2019), which also may have aided eventual continental break-up. The timing of the emplacement of the SDC is therefore commonly interpreted to coincide with the formation of the magma-rich rifted margin bordering Iapetus (Fig. 9) (Kjoll et al., 2019a; Svenningsen, 2001; Svenningsen, 1994; Tegner et al., 2019).

4.2.3. North Scandinavian Caledonides

Autochthonous Baltic basement gneisses in the Northern Segment are mostly Archean–Paleoproterozoic in age and are locally intruded by early Mesoproterozoic bodies (Bingen et al., 2016; Krill et al., 1985; Levchenkov et al., 1993; Slagstad et al., 2016; Bergh et al., 2007; Corfu et al., 2005; Kullerud et al., 2005; Melezhik et al., 2015; Myhre et al., 2013; Zouлыa et al., 2009). Noonian–Ediacaran intrusives are reported from the autochthon. However, crystallization ages of white mica recovered from fault gouges of normal faults in the autochthon indicate that the basement experienced episodes of extension in the latest Mesoproterozoic (~1050 ± 15 Ma) to mid-Mesoproterozoic (~825–810 ± 18 Ma; Koehl et al., 2018). Locally, the basement gneisses are unconformably covered by tec tally little disrupted Proterozoic sedimentary successions that extend from the Tonian(?) to the middle Cambrian

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**Fig. 9.** Simplified stratigraphic/magmatic/tectonic framework in the Baltican autochthon and the Scandinavian nappe complexes. Subcolumns 1, 2 and 3 in both the autochthon and nappe complexes (NCs) depict: schematic stratigraphy (1), magmatic activity (2) and interpreted phases of rifting and drifting (3). Note that a Gaskiers age for the Moelv Formation is adopted here. Abbreviations: ED = Egersund Dykes (Bingen et al., 1993); EG = Eidvågeid Granite (Kirkland et al., 2001); FCC = Fen Carbonatite Complex (Meert et al., 1998); HD = Hurnedalen Dykes (Walterhaug et al., 1999); RG = Repvåg Granite (Kirkland et al., 2006); SDC = Scandinavian Dyke Complex (e.g. Svenningsen, 2001; Kjoll et al., 2019a, 2019b; Baird et al., 2014; Root and Corfu, 2012; Paulson and Andréasson, 2002; Kumpulainen et al., 2021); SL = late Sveco-norwegian granites (e.g. Bingen et al., 2008); SIP = Sella Land Igneous Province (Pedersen et al., 1989; Roberts et al., 2010); SG = Leucosomes in Søsvy succession (e.g., Gasser et al., 2015); UML = ultramafic lamprophyre at Vinoren (Zouлыa et al., 2020).

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Dividalen Gr. (e.g. Andresen et al., 2014; Kirkland et al., 2011). The Dividalen Gr. starts with conglomerates and fluvial sandstones at the base, which are overlain by shelfal and platform deposits (Tornetrask Fm.). The latter contains the Ediacaran-Cambrian boundary. The Tornetrask Fm., in turn, is overlain by the middle Cambrian Alum Shales (Kirkland et al., 2011; Andresen et al., 2014; Kirkland et al., 2011; Banks, 1973; Jensen and Grant, 1998; Thelander, 1982; Thickpenny, 1984). Locally, the transgressive sequence is underlain by discontinuous remnants of tillite (Winchester, 1988) that may correlate with the Moelv Fm. The Dividalen Gr can be traced along the Scandinavian erosional front over large distances and is suggested to link up with the autochthonous successions in southern Norway (Winchester, 1988).
(Figs. 8, 9, B6) (Andresen et al., 2014; Kirkland et al., 2011; Nystuen et al., 2008). Two glacial deposits have been reported from these parautochthonous successions, i.e. the Smallfjord and Mortensnes formations, which are interpreted to be Marinoan and Gaskiers in age, respectively (Halverson et al., 2005; Nystuen et al., 2008; Rice et al., 2011; Appendix B). The two glacial formations are separated by sandstones, shales and carbonates of the Nyborg Fm (Fig. B6). Carbonates at the base of the Nyborg Fm. are interpreted to include Marinoan cap carbonates (Rice et al., 2011). The younger of the two glacial deposits is overlain by shallow-marine deposits of the Verstertana Gr., in which the carbonates (Rice et al., 2011). The younger of the two glacial deposits is interpreted to include Marinoan cap carbonates (Rice et al., 2011). The younger of the two glacial deposits is overlain by shallow-marine deposits of the Verstertana Gr., in which the carbonates (Rice et al., 2011).

The tectono-stratigraphically lowest allochthons in the Northern Segment are composed of Neoproterozoic sedimentary successions that are interpreted to largely repeat the Precambrian of the parautochthon (Siedlecka and Siedlecki, 1971; Townsend et al., 1989). Similar to the Precambrian successions in the Southern and Central segments, there is a paucity of mafic magmatic rocks within this lowest structural level. Structurally overlying these Neoproterozoic successions are Tonian metasedimentary complexes, commonly referred to as the Kalak Nappe Complex. The Tonian metasedimentary complexes are predominantly composed of monotonous, shallow-marine, siliciclastic successions, which have been episodically metamorphosed and were also intruded by granites. Two successions can be distinguished: an older succession that was deposited between ~1030 and 980 Ma, and a younger succession that was deposited between 910 and 840 Ma (Figs. 9, B6) (Kirkland et al., 2007). In both cases, the estimated depositional ages are bracketed by the youngest detrital zircons among the successions and the age of the oldest intrusions cutting them. The 980 Ma and 840 Ma plutons are bimodal but mainly comprise granitoids that could have formed via partial melting of the surrounding metasediments in a continental rift (Andréasson et al., 2018) or developed in a volcanic arc setting (Kirkland et al., 2006). The Tonian metasedimentary complexes experienced
episodic metamorphism and some partial melting in the Cryogenian between ~710 and 680 Ma (Corfu et al., 2007; Gasser et al., 2015; Kirkland et al., 2006). The metamorphic mineral assemblage associated with deformation and leucosome formation, dated at 702 Ma, indicates that the rocks were buried to a depth corresponding to pressures of ~8 kbar at that time (Gasser et al., 2015).

Locally, extensive mafic–ultramafic magmas intruded the Tonian metasedimentary complexes at 580–560 Ma, i.e. the Seiland Igneous Province (SIP; Figs. 8, 9, B6). That Ediacaran magmatic phase was followed by a ~30 Ma magmatic hiatus before a second, subordinate phase of carbonatite and syenite magmatism occurred in the early Cambrian at ~530–520 Ma (Larsen et al., 2018; Pedersen et al., 1989; Roberts et al., 2006; Roberts et al., 2010), for which no analogue in the autochthon is known (Fig. 9). However, more than 90% of the mafic–ultramafic rocks of the SIP were emplaced within a short time span of ~4 Ma between 570 and 560 Ma (Roberts et al., 2010, 2006). Estimations of the ambient pressures during the 580–560 Ma magmatism (Larsen et al., 2018) suggest that the Tonian metasedimentary complexes remained at depths corresponding to ~8 kbar throughout the Cryogenian and Ediacaran. These magmatic rocks are interpreted as the remnants of a deep-seated magmatic conduit system that intruded the Tonian metasedimentary complexes at lower crustal levels during the formation of the CIMP and the magma-rich Iapetan rifted margin (Larsen et al., 2018).

4.3. Amazonia

The Amazon craton of northern South America is exposed as two shields—the Guiana Shield in the north and the Brazilian Shield in the south—separated by the Amazon basin (Fig. 10). The craton is mostly composed of Paleoproterozoic and Mesoproterozoic components, and notably includes the ~1.3–0.9 Ga Sisnsás Province that reflects the participation of Amazonia in Rodinia’s assembly (Gordani et al., 2009; Teixeira et al., 2010). The western margin of Amazonia is widely assumed to represent a former Iapetan margin that developed following the breakup of Amazonia and Laurentia. However, as will become clear from the descriptions below, relics of that Iapetan margin are almost entirely absent. Indirect inferences therefore become important.

4.3.1. Northwest Amazonia

Northwest of the Guiana Shield exposures, Precambrian basement inliers have been identified in the Andes of Colombia and Venezuela (Figs. 10, B7). Very broadly, they are characterized by ~1.1–0.9 Ga medium to high-grade metamorphic rocks which bear strong geochronological, petrological and geochemical similarities to Mesoproterozoic rocks of the Amazon Craton (Cardona et al., 2010; Ibanez-Meja et al., 2011; Kroonenberg, 2019). Although paleomagnetic data have suggested that some of those inliers may have been translated northward along the margin during the Phaner zoic (Bayona et al., 2010), they are nevertheless generally considered to be parutochthonous to the craton.

Post-Grenvillian rocks of demonstrated Neoproterozoic age are virtually unknown to the northwestern margin of the Amazon craton, although they have been inferred to occur among high-grade metamorphic rocks of the Colombian and Venezuelan Andes on the basis of contested age data (van der Lelij et al., 2016). However, there are two notable exceptions: the first is a well-dated nepheline syenite from an inlier of the Llanos Basin (Colombia) that has yielded a U-Pb crystallization age of 577.8 ± 6.3/–9 Ma (Figs. 10, 11) (Meija et al., 2012). The second, the Huarguallul Gabbro unit, comprises structural slices of metagabbro and peridotite along the Peltetec fault zone of the western Cordillera Real of Ecuador, and has yielded imprecise Ar-Ar dates of ~582–566 Ma (Spikings et al., 2021). Given those ages and the subalkalic composition of the mafic rocks, Spikings et al. (2021) postulated that they were associated with CIMP (Tegner et al., 2019) and may thus have heralded continental breakup. In addition to those meager in situ relics, detrital zircons from some Paleozoic metasedimentary and metamorphic rocks of the Venezuelan, Colombian and Ecuadorian Andes yield significant late Neoproterozoic and early Cambrian age fractions (Fig. 11) (Mantilla-Figueroa et al., 2016; Martens et al., 2014; Spikings et al., 2021; Suhr et al., 2019; van der Lelij et al., 2016). The provenance of those late Neoproterozoic detrital zircons is unclear, and could be complicated by younger, margin-parallel terrane translations of their host units (Bayona et al., 2010). However, as will be shown later, further consideration of the possible nature and implications of this age fraction is illuminating; we will return to this in the next section (section 4.3.2) and in the discussion.

The oldest Phanerozoic rocks of this region are Cambrian in age, but their absolute ages are generally poorly constrained. In the west, they occur as high-grade metamorphic rocks in the Northern Andes, whereas to the east they are mostly unmetamorphosed but are extensively buried in the foreland basins (Fig. B7) (Feo-Codecido et al., 1984; Tazzo-Rangel et al., 2019; van der Lelij et al., 2016). The eastern successions comprise continental to shallow marine rocks that have conventionally been interpreted as passive margin sequences (Feo-Codecido et al., 1984), but they have also been proposed to represent syn-rift deposits possibly associated with the opening of the Rheic Ocean in the latest Cambrian–early Ordovician (Bartok, 1993; Rodriguez Milano et al., 2016). Widespread magmatism appeared in this region at least by the late Cambrian, around 500 Ma (van der Lelij et al., 2016), but became much more extensive during the Ordovician (Figs. 11, B7). That magmatism is ascribed to the Famatinian active margin that probably spanned most of

Fig. 11. Simplified stratigraphic/magmatic/tectonic framework along the northwest, southwest and southern margins of Amazonia and of the Rio de la Plata Craton. Subcolumns 1, 2 and 3 for each region depict: schematic stratigraphy (1), magmatic activity (2) and interpreted phases of rifting and drifting (3). Abbreviations: AC = Alto Caapucú block; CB = Chilla beds.
the margin of west Gondwana then (Fig. 10) (Ramos, 2018). Famatinian orogenesis in the Northern Andes was furthermore marked by widespread Ordovician metamorphism and deformation (Cardona et al., 2016; Mantilla-Figueroa et al., 2016; van der Lelij et al., 2016).

4.3.2. Southwest Amazonia

Southwest of the Amazon Craton, Precambrian rocks are known from two coast-parallel belts: an eastern belt in the Eastern Cordillera of Peru, and a western belt along the offshore Outer Shelf High (Fig. 10). The latter is likely a continuation of the Arecipa Massif exposed along the southwest coast of Peru, and is discussed in section 4.5.2. In the Eastern Cordillera, Precambrian exposures are really limited; the oldest known rocks are late Neoproterozoic (~1.12–0.98 Ga) subduction-related granitoids that Miskovik et al. (2009) have suggested to be correlative with the Sunis Province of Amazonia, implying the Eastern Cordillera to be, at least partly, floorored by basement parautochthonous to Amazonia.

Post-Grenvillian rocks of unequivocal Neoproterozoic age are restricted to a few, small middle Neoproterozoic granitoids (~750–690 Ma) and isolated, metamorphosed ultramafic bodies (718 ± 47 Ma) in the Eastern Cordillera (Figs. 10, B7) (Miskovik et al., 2009; Tassinari et al., 2011). Miskovik et al. (2009) have classified the former as A-type, anorogenic granitoids and suggested them to relate to continental breakup. The ultramafic bodies, which occur in nappes, have been interpreted as a dismembered ophiolite that was partly subducted and then exhumed in the Ordovician, before being thrust over unmetamorphosed Carboniferous sediments overlying the basement of the Eastern Cordillera (Castroviejo et al., 2009; Tassinari et al., 2011; Willner et al., 2014). Although the details remain sparse, those A-type granitoids and ophiolitic rocks in the Eastern Cordillera could reflect a phase of rifting and continental breakup along western Amazonia in the late Tonian–early Cryogenian (Fig. 11).

Beyond the Eastern Cordillera of Peru, a small exposure of post-Grenvillian Neoproterozoic rocks (Chilla beds) has been reported from an isolated pop-up structure in the Bolivian Altiplano (Fig. 10). The Chilla beds comprise metabasalts and tuffs interbedded with conglomerates and poorly sorted, turbidite sandstones (Bahlburg et al., 2020). Geochemically, the volcanics resemble tholeiitic, intraplate basalts, and the succession is interpreted to reflect an extensional setting (Bahlburg et al., 2020). Detrital zircons provide a Tonian (925 ± 12 Ma) maximum depositional age for the sequence, but an upper age constraint is ill-defined. Bahlburg et al. (2020) consider the absence of any post-Tonian age fractions to be significant given that nearly all Ordovician and younger sedimentary rocks of the region exhibit a pronounced late Cryogenian–Ediacaran detrital zircon age fraction (see below). They thus infer the depositional age of the Chilla beds to predate that age fraction, and specifically prefer a Tonian depositional age. Notably, their argumentation equally permits the Chilla beds to be early Cryogenian, and thus potentially a further relic of late Tonian–early Cryogenian rifting along western Amazonia (Figs. 11, B7).

The bulk of the metamorphic rocks of the Eastern Cordillera, collectively referred to as the Maranon Complex (Fig. 10, B7), were previously thought to represent exposures of Precambrian basement, but detrital zircon studies have indicated they were mostly deposited during the Ordovician and Carboniferous (Cardona et al., 2009; Chew et al., 2007a, 2007b; Miskovik et al., 2009). Significantly, the detrital zircon populations of those metamorphic rocks and other Paleozoic sedimentary rocks of the Eastern Cordillera have a characteristic age distribution with prominent Precambrian peaks at ~1.2–1.0 Ga and ~650–550 Ma (Fig. 11) (Chew et al., 2008a, 2008b), which are furthermore observed in modern river sediment of the central Andes (Pepper et al., 2016). The first of those two age fractions has been interpreted as a signal of ‘Grenvillian’ provenance, for which the Sunis Province, or the local late Mesoproterozoic granitoids could potentially have been a source. However, the second peak is unexpected, as it has no obvious source in western Amazonia. Chew et al., 2008a, 2008b considered three alternatives to explain that late Cryogenian–Ediacaran age fraction: 1) a local Neoproterozoic active margin, 2) local rift-related magmatism, and/or 3) derivation from the Brasiliano orogens on the far (eastern) side of Amazonia. The first alternative is problematic in that no in situ evidence of a local arc of such age has been recognized in western Amazonia. The second alternative is similarly problematic because there is little evidence of rift-related magmatism of that age from the margin, and because compositionally basic products do not produce abundant zircon (Cawood et al., 2012). The third alternative was deemed unlikely by Chew et al., 2008a, 2008b because there is only a minor signal of the Amazon Craton among the detrital zircon age fractions, despite the fact that sediment transportation from the Brasiliano orogens would likely necessitate traversing wide swaths of it. We will return to this conundrum in the discussion section, but here note that we consider the provocative hypothesis of a hidden, local late Neoproterozoic arc to be a compelling scenario (Robert et al., 2020).

4.3.3. South Amazonia + Rio Apa block

In contrast to the obscured Neoproterozoic history of its western margins, along the south of Amazonia, Neoproterozoic rocks unconformably overlie the craton directly. There, Neoproterozoic rocks occur in the northern Paraguay belt of southwest Brazil, the Tucavaca belt of northeastern Bolivia, and in the Corumbá area where those belts meet (Fig. 10, B8). To the south of the Corumbá area, similar Neoproterozoic rocks are found in two belts unconformably overlaying the east and southwest flanks of a distinct Precambrian cratonic block, Rio Apa. Rio Apa comprises a north–south elongate exposure of Paleoproterozoic (~2.0–1.7 Ga) basement that was metamorphosed in the early Mesoproterozoic (1.35–1.30 Ga) and invaded by gabbros and diabase dikes at ~1.1 Ga (Cordani et al., 2010; Teixeira et al., 2019). The Neoproterozoic rocks unconformably covering the flanks of that block exhibit strong lithostratigraphic similarities to those of the North Paraguay and Tucavaca belts, and they are generally interpreted to reflect a common, regional evolution (Appendix B). This implies that Rio Apa was proximal to or coherent with southern Amazonia by the time deposition in those belts began.

The broadly-common evolution displayed by those Neoproterozoic belts began with siliciclastic sedimentation that started at least by late Cryogenian time (but possibly significantly earlier), and which may have been instigated by an extensional event (Figs. 11, B8) (Freitas et al., 2011; Litherland, 1986; Piacentini et al., 2013; Trompette et al., 1998). However, the details and causes of that extensional episode remain unclear. Proposed explanations include the impingement of a mantle plume beneath the region (with the Tucavaca, North and South Paraguay belts forming a rift-rift-rift triple junction) (Jones, 1985), tectonic disruption between Amazonia and neighboring crustal blocks (Cordani et al., 2009; Ramos et al., 2010), and extension driven by foreland flexure or extrusion associated with the developing Brasiliano system to the southeast (Trompette et al., 1998). A regime of prevailing extension may have locally persisted into the early Ediacaran along the east and southwest flanks of Rio Apa, but by mid-to-late Ediacaran time the southern margin of Amazonia and Rio Apa were both characterized by stable carbonate platforms. During the course of the late Ediacaran, carbonate deposition was supplanted by siliciclastic basinal deposits that also recorded an apparent shift in regional sedimentary provenance. In the latest Ediacaran to Cambrian, the Neoproterozoic successions of the region were variably deformed, and in the eastern part of the Paraguay belt, syn- to post-orogenic granitoids were intruded between ~550 and 500 Ma (Figs. 11, B8) (Godoy et al., 2010; McGee et al., 2012; Tolver and Trindade, 2014). The origin of that latest Ediacaran–Cambrian tectono-magmatic episode remains controversial, and has been proposed to relate either to a continent-continent collision between Amazonia and Rio de la Plata (and other cratons of central Gondwana) (Tolver et al., 2010; Tolver and Trindade, 2014), or to displacement along a continental megashear (the Transamazonian lineament) (Cordani et al., 2013, 2014). The former scenario requires that an ocean
existed between Amazonia and Rio de la Plata prior to latest Ediacaran–Cambrian time: the Clymene Ocean (Tohver and Trindade, 2014; Trindade et al., 2006). Although a detailed assessment of those competing hypotheses is beyond the scope of our focus herein, a further exploration of their implications for models of Iapetus Ocean opening is presented in the discussion.

4.4. Rio de la Plata

The Rio de la Plata Craton (RPC; Figs. 10, B9) to the south of Amazonia represents one of the major cratons of South America, and may have been involved in the final breakup of Rodinia alongside Amazonia (e.g. Li et al., 2008). The RPC is characterized by Paleoproterozoic (~2.2–2.1 Ga) basement that avoided Neoproterozoic reworking that is otherwise very prominent in the Brasiliano belts to the north and northeast, and in the Eastern Sierras Pampeanas to the west (Chyntico, Ciabus et al., 2018; Rapela et al., 2011). The basement and Neo-proterozoic–Cambrian cover of the craton is largely buried, with unambiguous exposures effectively restricted to southwestern Uruguay and the Buenos Aires Province of Argentina.

In the north, the extent of the RPC is debated (Appendix B), but it may include the Alto Caupúcú block of southwestern Paraguay (Figs. 10, B9), which includes Paleoproterozoic basement (~2.0 Ga) that was deformed and intruded by granodiorite in the early Ediacaran (~625–620 Ma; Cordani et al., 2001; Cubas et al., 1998). Ediacaran volcano-sedimentary rocks were then deposited in the Alto Caupucú block before a second phase of deformation affected the region and it was invaded by late Ediacaran (~565–560 Ma), A-type felsic rocks interpreted to reflect an extensional regime. A small, A-type granite also invaded the eastern margin of the RPC (Piedra Alta terrane) at ~587 Ma (Figs. 11, B9) (Cubas et al., 1998; Leite et al., 2018).

In the south of the RPC, the Precambrian basement and its unconformably overlying Neoproterozoic cover is exposed in the Tandilia belt of the Buenos Aires Province (Fig. 10). The Neoproterozoic successions there comprise carbonate and siliciclastic sequences that may span most of the Cryogenian and Ediacaran, and they are unconformably overlain by Ordovician-Silurian sandstones and shales (Figs. 11, B9) (Arrory et al., 2019; Gómez-Peral et al., 2017). The detrital zircon spectra of the Neoproterozoic rocks are dominated by a Paleoproterozoic age fraction and lack Neoproterozoic grains, suggesting they have been derived from the local RPC basement (Cingolani, 2011). Together, the Neoproterozoic-Early Paleozoic rocks are generally interpreted as stable platform successions, although structural evidence reveals episodes of late Ediacaran-early Paleozoic(? shortening (Cingolani, 2011; Hernández et al., 2017). To the southeast of the Tandilia belt, the Sierra de la Ventana system exhibits a contrasting Neoproterozoic history with multiple episodes of calc-alkaline and alkaline magmatism between late Tonian and early Cambrian time (Justiniano et al., 2020), and so was likely distinct from the RPC until late Neoproterozoic or Cambrian time (Appendix B).

In the west, the RPC is entirely buried beneath the Chaco-Paraná basin, but the continuation of the craton beneath those basins has been demonstrated from the recovery of ~2.2–2.1 Ga basement rocks from boreholes in the Córdoba Province of central Argentina (Rapela et al., 2007). The specific location of the western boundary of the craton has also been inferred from geophysical data, which have revealed a sharp lineament against the Eastern Sierras Pampeanas (Pavetto et al., 2015; Peri et al., 2013; Ramé and Miro, 2011; Rapela et al., 2011). In the Chaco-Paraná boreholes, the Paleoproterozoic rocks are unconformably overlain by late Paleozoic sedimentary rocks, so no Neoproterozoic-early Palaeozoic record is preserved. However, on the basis of HF- and Nd-model ages from early Paleozoic igneous rocks near the south and western margins of the RPC, and from the detrital zircon records of its Neoproterozoic and younger cover in the east, Chernicoff et al. (2012) suggested that the western margin of the RPC may trace a Grenville-aged Mesoproterozoic orogen extending from the Sunsás Province of Amazona.

4.5. Smaller cratonic blocks and ‘suspect’ terranes of South and Central America

In addition to the small basement inliers described from the Andean ranges to the northwest and southwest of Amazonia that may be parautochthonous to that craton (section 4.3), there are also a series of basement blocks further outboard in southwestern Peru, southwest Bolivia, western Argentina and Chile that have been considered distinct terranes that may have been tectonically independent of Amazonia and Rio de la Plata for part or all of the late Neoproterozoic-Cambrian time frame that we are concerned with here. Because the interaction of these various blocks and suspect terranes has important implications for any model of the region across this time frame, we succinctly review and discuss some of the prevailing observations and concepts concerning them below. Importantly, the concept, delineation and affinity of these various terranes are almost invariably debated, and we do not strive to present a comprehensive review of all the interpretations that have been put forward, but rather attempt to provide a concise picture of what we consider to be the presently-prevailing ones. The order of their presentation roughly proceeds from inboard to outboard (or approximately east to west; Figs. 10, B10): the Eastern Sierras Pampeanas, the Western Sierras Pampeanas and Arequipa-Antofalla block, the Precordillerata terrane, and then finally the Oxaquia and Maya blocks of present-day Mexico that were likely positioned along the northwest of Amazonia in the late Neoproterozoic-early Paleozoic. As before, expanded descriptions of each of these are provided in Appendix B.

4.5.1. Eastern Sierras Pampeanas

The Sierras Pampeanas is a region of northwestern Argentina characterized by north-south oriented basement uplifts. The region is flanked to the east by the Paleoproterozoic Rio de la Plata Craton (section 4.4), and to the west it is bordered by the Precordillera terrane (section 4.5.3 and Fig. 10). Geologically, the Sierras Pampeanas can be divided into a latest Precambrian-early Paleozoic metasedimentary succession of potential Gondwana affinity—the Eastern Sierras Pampeanas (ESP)—and a possibly exotic Paleoproterozoic basement block in the west—the Western Sierras Pampeanas (WSP; section 4.5.2). The ESP itself can be further subdivided into an eastern belt of late Neoproterozoic to Cambrian low- to high-grade metasedimentary rocks affected by Ediacaran–early Cambrian deformation, metamorphism and magmatism—the Pampean belt—and a western belt of metasedimentary rocks that were affected by late Cambrian to Ordovician deformation, metamorphism and magmatism—the Famatinian belt (Fig. B10) (Appendix B). However, the former belt is not exclusive to the latter, and rocks that exhibit signs of having experienced both orogenic episodes lie at their interface; further west, in the Famatinian belt, relics of the Pampean phase may have been entirely overprinted by Famatinian orogenesis.

Late Neoproterozoic to early Cambrian rocks of the Pampean belt are characterized by monotonous tracts of turbidites that are referred to as the Puncoviscana 'Formation' (Figs. 12, B10) (Acanolaza and Acanolaza, 2007; Jeze et al., 1985; Omarini et al., 1999; Zimmermann, 2018). The base and basement of that complex is nowhere exposed, and laterally its stratigraphy has been obscured by deformation and metamorphism. However, in the Eastern Cordillera, where the rocks are of low grade, sedimentological structures seem to reflect a generally westward-deepening trend in the depositional environment (Acanolaza and Acanolaza, 2007). The detrital zircon age spectra of the turbidites are generally characterized by a late Mesoproterozoic peak (~1.1–1.0 Ga) and a late Neoproterozoic peak (~0.65–0.55 Ga; Adams et al., 2011, Adams et al., 2008; Drobe et al., 2011; Escayola et al., 2011; González et al., 2014; Hauser et al., 2011; Weinberg et al., 2018). The zircons mostly have negative εHf values (Casquet et al., 2018; Hauser et al., 2011)—suggesting a continental provenance—and paleocurrent indicators reveal that sediment was transported from the east and...
Fig. 12. Simplified stratigraphic/magmatic/tectonic framework of displaced terranes in South and Central America. Subcolumns 1, 2 and 3 for each region depict: schematic stratigraphy (1), magmatic activity (2) and interpreted phases of rifting and drifting (3). Abbreviations: A.A.-W.S.P. = Arequipa-Antofalla and Western Sierras Pampeanas; PV = Puncoviscana Fm.

southeast (Adams et al., 2008; Ježek et al., 1985). Together, those observations might suggest that the neighboring Rio de la Plata Craton was the primary source of these Precambrian sediments, yet ages of ~2.0–2.3 Ga, which dominate and characterize that craton, are scarce in the turbidite detrital zircon age distribution (Escayola et al., 2007). Other proposed (but debated) sources of the Mesoproterozoic and Neoproterozoic sediments include the Western Sierras Pampeanas (Collo et al., 2009; Escayola et al., 2007), the Brasiliano Belt, and the Kalahari Craton (Casquet et al., 2018; Hauser et al., 2011; Rapela et al., 2016). The youngest (latest Ediacaran-earliest Cambrian) detrital zircon grains are associated with euhedral forms (Adams et al., 2011; Adams et al., 2008; Hauser et al., 2011) and the abundance of this first-cycle detritus suggests that deposition of the turbidites occurred in close proximity to an active arc (Cawood et al., 2012).

Pampean metamagism, which spanned at least from ~555 to 520 Ma but could have started earlier, was characterized by the emplacement of calc-alkaline, I-type, metaluminous to slightly peraluminous granitoids and the eruption of rhyolites, dacites and tuffs (Figs. 12, B10) (Casquet et al., 2018; Escayola et al., 2011; Iannizzotto et al., 2013; Luchi et al., 2018; von Gosen et al., 2014). The major and trace element geochemistry of these rocks commonly display subduction-related signatures, suggesting that they comprised an arc along an active continental margin (Casquet et al., 2018; Escayola et al., 2011). Mafic-ultramafic rocks are also reported from the Pampean belt, but their ages and tectonic affinity remain unclear. Some alkaline basic rocks that were previously assigned Neoproterozoic ages and interpreted to mark continental extension (Omarini et al., 1999; Ramos, 2008) are now recognized to be post-Middle Ordovician in age (Hauser et al., 2011). However, some mafic-ultramafic rocks are interlayered with Puncoviscana Fm. equivalents and so must be mid-Cambrian in age or older. Escayola et al. (2007) reported a Sm-Nd isochron age of 647 ± 77 Ma from some of those mafic-ultramafic rocks and interpreted them as ophiolitic remnants (see also Anzil et al., 2012; Casquet et al., 2018), possibly marking a relict backarc or terrane suture. However, Tibaldi et al. (2008) interpreted other mafic-ultramafic bodies of the same region as Cambrian, OIB-like magmas possibly associated with ridge-subduction, and Weinberg et al. (2018) have alternatively suggested the mafic-ultramafic rocks to be products of hyper-extension. The cessation of Pampean arc magmatism was preceded by regional deformation and metamorphism that peaked at ~530–520 Ma and which has been ascribed to various processes including ridge subduction (Schwartz et al., 2008), but most often the result of collision between the ESP and the WSP (Casquet et al., 2018; Escayola et al., 2011; Rapela et al., 2016; Weinberg et al., 2019). Following a perhaps ~20 Ma arc magmatic hiatus, arc magmatism associated with the Famatinian cycle began by ~500 Ma (Fig. 12).

4.5.2. Western Sierras Pampeanas and Arequipa-Antofalla

Among the basement uplifts of the WSP, the oldest components are Proterozoic rocks that were metamorphosed between ~1.2 and 1.0 Ga and intruded by massif-type anorhthosites at ~1.09 to 1.07 Ga (Casquet et al., 2006; Casquet et al., 2008a, 2008b; Rapela et al., 2010; Varela et al., 2011). Those Mesoproterozoic rocks are juxtaposed with Neoproterozoic metasedimentary successions, and are locally intruded by ~845 and ~774 Ma A-type granitoids and a late Ediacaran (~570 Ma) syenite-carbonatite (Figs. 12, B10) (Baldo et al., 2006; Casquet et al., 2008a, 2008b; Colombo et al., 2009). Some of the Neoproterozoic successions have been correlated with the Precordilleran, whereas others have been linked with units in the ESP, and so there may be some cryptic terrane boundary associated with them (Galinondo et al., 2004; Mulcahy et al., 2013; Murra et al., 2016; Rapela et al., 2016; Appendix B). During the Famatinian cycle in the Ordovician and Silurian, both the Mesoproterozoic basement and Neoproterozoic cover were metamorphosed, deformed and invaded by peraluminous granitoids (~481–470 Ma).

To the northwest of the WSP lies a much larger continental block, the Arequipa-Antofalla block (AA), which the WSP may represent the southeast extension of (hereafter WSP-AA; Appendix B). The AA comprises Precambrian metamorphic basement inliers in southern Peru, west Bolivia and northern Chile (Fig. 10). The oldest rocks occur in the north of the block and comprise Paleoproterozoic (~2.0–1.8 Ga) intrusions that were metamorphosed at ~1.8 Ga (Loewy et al., 2004). In the south of the block the oldest rocks include sediments, volcanics and intrusives of mid-late Mesoproterozoic age that, together with the northern part of the block, experienced metamorphism between ~1.2 and 0.9 Ga. Following that, Neoproterozoic glacial tillites and then carbonates (possibly cap carbonates) were deposited in the north, whereas clastic sediments were deposited in the south (Chew et al., 2007a, 2007b; Loewy et al., 2004; Pankhurst et al., 2016). Neoproterozoic volcanic-volcaniclastic rocks are known from one area in the south (Quebrada Choja), where the succession is intruded by ~635 Ma dacite dikes (Figs. 12, B10) (Loewy et al., 2004). The entire block was then affected by another metamorphic event in the early Paleozoic (~470–440 Ma), associated with the invasion of subduction-related granitoids of the Famatinian arc (Ramos, 2008). Detrital zircons from Mesozoic and Cenozoic sedimentary successions in the northern part of the block exhibit age peaks at ~1.3–1.0 Ga and ~700–500 Ma; the former being recognized as ‘Greenvillian’ and the latter as ‘Pampean-Brasiliano’ (Wotzlaw et al., 2011).

The Precambrian relationship between the AA and the Amazon Craton, as well as between the WSP and the ESP/Rio de la Plata, remain uncertain. WSP-AA must have been proximal to western Gondwana at least by the early Paleozoic because the Ordovician-Silurian Famatinian arc crosses through both the Antofalla and Arequipa Massifs (Fig. 10).
Conversely, on the basis of differing Paleoproterozoic histories, Loewy et al. (2004) argued that WSP-AA was not affiliated with Amazonia at ~2.0–1.8 Ga. A common interpretation therefore links the contemporaneous ‘Grenvillian’ (~1.2–1.0 Ga) metamorphic episodes in WSP-AA and the Sunsia Province, and contends that they collided together at that time (Loewy et al., 2004; Ramos, 2008). Importantly, however, these inferences do not exclude the possibility that WSP-AA could have been separated from Amazonia in the late Neoproterozoic before being re-assembled with it before Famatinian orogenesis.

4.5.3. Precordillera

The Precordillera (or Cuyanía) terrane was first defined by a succession of early Paleozoic platform carbonates exposed in the Precordillera of central west Argentina (Fig. 10) that bear a striking resemblance to coeval successions in southeast Laurentia, and which formed the basis for a now prevailing hypothesis that those rocks are exotic to west Gondwana (Astini et al., 1995; Ramos, 2004). However, the boundaries of the Precordillera terrane beyond that early Paleozoic platform succession, as well as the nature of the basement beneath it, remain poorly defined and controversial (Fig. B10) (Appendix B). The basement beneath the early Paleozoic platform rocks in the Precordillera is not exposed, but Miocene volcanics in the southeast of the Precordillera contain xenoliths of highly deformed, amphibibole to granulate facies metamorphic mafic and silicic rocks that are thought to be sampled from it (Kay et al., 1996). U-Pb dating has determined the age of the xenoliths to be late Mesoproterozoic (~1.10–1.16 Ga) and their whole rock Nd and Pb compositions have been argued to reflect a Laurentian Grenvillian affinity (Kay et al., 1996; Rapela et al., 2010).

Deformation and metamorphism of those high-grade basement rocks must have occurred prior to the deposition of the early Paleozoic platform succession, and may be reflected by U-Pb ages of ~1.06–1.08 Ga from zircons in the xenoliths that are associated with lower U and Pb concentrations and complex textures (Kay et al., 1996; Rapela et al., 2010; Varela et al., 2011).

The oldest exposed rocks in the Precordillera proper are late early Cambrian red beds, evaporites and carbonates that have been interpreted as syn-rift deposits (Figs. 12, B10) (Gomez and Astini, 2015; Thomas and Astini, 2003). Those syn-rift rocks are overlain by a thick sequence of late early to middle Cambrian deep-water black shales that first appeared in the late Early Ordovician as forearc clastic wedge onto the Precordillera as it collided with Gondwana. These inliers that bear a similar composition and were metamorphosed to granulate facies at ~1.0 Ga are thought to represent exposures of a common, largely buried Grenvillian terrane (‘Oaxaquía’) (Keppie and Ortega-Gutiérrez, 2010; Ortega-Gutiérrez et al., 1995; Appendix B). The oldest sedimentary cover atop those basement rocks are latest Cambrian-Ordovician shallow marine carbonates and clastic rocks that provide faunal links to Gondwana, and Oaxaquia is commonly reconstructed to the north/northwest margin of Amazonia then (Boucot et al., 1997; Keppie et al., 2008; Landing et al., 2007; Streng et al., 2011). Having participated in Grenvillian orogenesis and being present along the northwest margin of Gondwana at least by the Cambrian, it is possible that Oaxaquia remained along the northwest margin of Gondwana throughout the Neoproterozoic (although importantly, as with WSP-AA, this does not preclude the possibility that Oaxaquia became temporarily independent from Amazonia during the late Neoproterozoic). Following ~1.0 Ga metamorphism, exhumation and cooling of Oaxaquia mostly occurred during the early Neoproterozoic (Keppie and Ortega-Gutiérrez, 2010), and a mafic dike swarm intruded the Novillo basement intruder at 619 ± 9 Ma (Fig. 12) (Weber and Schmitt, 2019). On the basis of their geochemistry and similarity in age to the Long Range and Egersund dikes, Weber and Schmitt (2019) interpreted the dike swarm to be plume-related and associated with the CIMP.

To the southeast of Oaxaquia, the Maya Block of southeast Mexico, Guatemala and Belize, also exhibits evidence of a Mesoproterozoic basement (at least in the south), including ~1.0 Ga felsic orthogneisses (Cisneros de León et al., 2017; González-Guzmán et al., 2016; Appendix B). The oldest cover of the Maya Block is a metasedimentary sequence of micaschists, calc-silicate rocks and marbles that are constrained by detrital zircons to be younger than ~0.9 Ga, but are inferred to be ~600–580 Ma on the basis of stable isotopes (González-Guzmán et al., 2016; Weber et al., 2008). That sequence also contains abundant amphibibole layers that are interpreted as mafic intrusions (dikes and sills) and could be further CIMP correlatives (González-Guzmán et al., 2016; Weber et al., 2008). Supportingly, Cisneros de León et al. (2017) inferred that a tectonothermal event affected the area at ~600 Ma on the basis of zircon U-Pb ages, and associated that event to the emplacement of a mafic dike swarm. In the northern part of the Maya Block, latest Ediacaran (~546 Ma) granitic basement fragments have been identified in the 65 Ma Chicxulub impact breccia, and have been suggested to relate to a magmatic arc (Keppie et al., 2011; Krogh et al., 1993). The Neoproterozoic-early Paleozoic tectonic affinity of the Maya Block is not clearly established. It has generally been associated to Oaxaquia on the basis of their similar Mesoproterozoic basement, and linked to the north margin of Amazonia, but the Ediacaran-Cambrian granites from the north of the block have also been used to link it to Avalonia or the Suwannee terrane (González-Guzmán et al., 2016; Keppie et al., 2011; Martens et al., 2010). It is also possible that the block is composite (Keppie and Keppie, 2014).

5. Discussion

To reconstruct the opening of the Iapetus Ocean requires knowledge of the pre-rifting continental configuration of Laurentia, Baltica and Amazonia and the timing of their continental breakup—both of which ideally can be retrieved from paleomagnetic and geologic constraints. However, a key deduction from our review of late Neoproterozoic paleomagnetic data (section 3) is that while the sparse available constraints provide critical insights, they are generally too few to allow the formulation of robust paleogeographic conclusions. Concerning the pre-breakup starting point, for example, our quality-filtered dataset is in agreement with several families of Rodinia fits (Li et al., 2008; Slagstad et al., 2019; Torvik et al., 1996; Dalziel, 1992), and so, apart from eliminating some exotic alternatives (e.g. Hartz and Torvik, 2002), these data cannot isolate the specific initial configuration of Laurentia, Amazonia and Baltica before the opening of Iapetus. The timing of Iapetus opening is similarly ill-constrained by the paleomagnetic data. For example, in the context of the three main opening times proposed in the literature for the ‘West Iapetus Ocean’ (~700 Ma vs. ~500–570 Ma
vs. ~550–530 Ma), the very small late Tonian to Cryogenian (750–635 Ma) quality filtered paleomagnetic dataset alone offers no means to evaluate the earliest option. The paleomagnetic data alone are therefore insufficient to reconstruct the opening history of the Iapetus Ocean. Likewise, the available geologic constraints, when considered in isolation, are often ambiguous, as evident by the variety of pre-breakup reconstructions and alternative interpretations of the same lithological assemblages.

What is clearly needed then, is an integrated consideration of these individually incomplete but strongly complementary paleomagnetic and geologic datasets, and a frank discussion of their joint uncertainties. In the following section, we attempt to present such an integrated discussion in the context of different possible scenarios of Iapetus opening. To ease the comparison between the different continental segments involved in the opening of Iapetus, we defined two domains, the western and eastern Iapetan domains, which comprise the ocean basins and the related rifted margins of Laurentia, Amazonia and Baltica (Fig. 13). In these domains, we also assign the segments capital letters (A, A', B, ...). The correspondence between the conjugate segments depends on the Rodinia fit chosen. For example, the default reconstruction in Fig. 13 displays the pre-rift position of B' (southwest Amazonia) adjacent to B (the central segment of Laurentia), while B' is adjacent to D (Scotland-Ireland) in the reconstruction of Dalziel (1992). First, we discuss the position of the major cratons (Laurentia, Baltic, West Gondwana) based on our review of the paleomagnetic and geologic constraints (section 5.1.1.). We then integrate geological constraints provided by the small terranes that evolved within the Iapetan system (section 5.1.2.), which we consider separately because their exact reconstructed locations are generally more uncertain. In the absence of extensive paleomagnetic data and unambiguous geologic constraints, the detrital zircon record may offer valuable insights into Iapetan tectonics; but it also represents a notoriously fickle tool. We therefore consider potential interpretations of the detrital zircon record and their implications for Iapetus ocean opening independently in section 5.1.3. Finally, in section 5.2 we present and describe our preferred interpretation of Iapetus opening in the form of a plate kinematic model.

5.1. An integrated perspective on Iapetan rifting phases

5.1.1. The view from the major cratons

Combining the paleomagnetic and geologic constraints reviewed above, we may now consider relative motions of the Iapetan margins from an integrated perspective. We start by noting, from the collected geological observations (section 4 and Appendix B), that two principal phases of rifting appear to have been widespread in the Iapetus realm: a first phase spanned ~750–700 Ma, followed by a largely quiescent interval in the Cryogenian, and then a second phase of rifting extended from ~635 to 520 Ma (Fig. 14). The eastern margin of Laurentia (eastern North America and Scotland-Ireland) was evidently affected by both phases. Relics of the first phase are observed in Scotland-Ireland, the Central Segment of North America and perhaps also the Southern Segment, but not in the Northern Segment. The second phase appears to have been much more widespread, having affected the entire eastern margin of Laurentia. Neoproterozoic rocks are sparse along western Amazonia, but there is nevertheless some (weak) evidence of the first rifting phase along the southwest margin of that craton, and similarly sparse evidence of the second phase in its northwest (Fig. 14). Likewise, the evolution of the western margin of Baltica is poorly understood for the time interval corresponding to the first rifting phase, but some indications of tectonics have nevertheless been observed in both its autochthonous and allochthonous domains. However, both of those domains exhibit much clearer evidence of the second rifting phase.

The interpretation of how these rifting phases relate and whether they were successful (i.e. leading to seafloor spreading) or aborted is dependent on the initial position of Laurentia, Baltica and Amazonia in the supercontinent Rodinia (e.g. Fig. 13). In section 3.2.2 we considered several distinct ‘families’ of reconstructions of Amazonia and Baltica relative to Laurentia, including some which considered Amazonia or Baltica as independent from Laurentia in Rodinia (Fig. 4f, g). In these cases the apparent contemporaneity of some of these rifting phases would be purely coincidental, as the margins would not have been conjugate. We also considered two families where Amazonia, Baltica and Laurentia were all adjacent: one in which Amazonia is positioned along the eastern margin of North America while Baltica lies along Scotland-Ireland (fit of Li et al., 2008; Fig. 4a-c), and another where Amazonia is adjacent to Scotland-Ireland and Baltica is opposite Greenland (fit of Dalziel, 1992; Fig. 4d). Recall that the late Tonian and

![Fig. 13. Distribution of the continental blocks involved in the opening of the Iapetus Ocean following the slightly modified Rodinia fit of Li et al. (2008). The segments A-A', B-B', C-C' and D-D' define the smaller scale rifting systems studied in this paper. The thick lines indicate the location of the ocean basins and seaway that successively open in the Iapetan system. The dashed lines refer to the suture of the Clymene Ocean (Trindade et al., 2006), and the Transbrasiliano lineament (Cordani et al., 2013). Abbreviations: AA = Arequipa-Antofalla; CIMP = Central Iapetus Magmatic Province; ED = Egersund dykes; ESP = Eastern Sierras Pampeanas; GD = Grenville dykes; LN/MV = Lough Nafooey arc/Midland Valley terrane; LR = Long Range dykes; MD = Mexican mafic dykes (in Novillo basement inlier); O&M = Oaxaquia and Maya blocks; Pc = Precordillera terrane; WSP = Western Sierras Pampeanas;]{fig13.png}
Cryogenian paleomagnetic data are extremely sparse and so cannot
distinguish between these reconstructions (Fig. 3a). Both re
constructions are also permissible with the relatively more numerous
paleomagnetic data at ~615 Ma, although, importantly, those data do
not rule out that those blocks were in fact already separated by an ocean
at that time (Fig. 4; see below). Thus, paleomagnetic data do not allow
us to uniquely identify conjugate margins before Iapetus opening.

The first rifting phase inferred in southwest Amazonia (B’ in Fig. 14)
is mirrored by rifting in both the Central Segment of Laurentia (B) and
Scotland-Ireland (D), and so a cursory look at the geological record
would suggest that either of those reconstructions could be viable. A
notable difference between those regions, however, is that post-rift
sediments are unknown from the Central Segment of Laurentia,
whereas in Scotland-Ireland, sedimentation continued until the second
rifting phase started at around ~635 Ma. The lack of post-rift sediments
in the Central Segment has often led to the interpretation that rifting was
aborted there (e.g. Aleinikoff et al., 1995), although curiously a broadly
similar narrative is inferred in Scotland-Ireland despite the enduring
accumulation of sediments (Stephenson et al., 2013). Unfortunately, a
comparison of these different marginal histories against the potentially
conjugate margin of western Amazonia is complicated by the fact that
Cryogenian relics are almost entirely unknown from the latter. The
assessment of which parts of the Laurentian and Baltican margins were
conjugate at this time (i.e. Scotland-Ireland or East Greenland) is
rendered similarly difficult by the dearth of records of this age from
western Baltic. It is therefore not straightforward to discriminate be
 tween these two reconstructions on the basis of only the geological re
cords of the first phase of rifting.

At the regional scale, the second rifting phase appears to have been
more protracted, extending from 635 to 520 Ma, although many of the
individual margins exhibit a narrower temporal range of activity
(Fig. 14). At this scale there is also a conspicuous first-order northeast
southwest younging trend across the eastern margin of Laurentia. While
rifting spanned from 635 to 580 Ma in Scotland-Ireland, it continued
into the latest Ediacaran/early Cambrian in the eastern segments of
North America, where the youngest rifting activity is recorded in the
Southern Segment (Fig. 14). Given this diachroneity, we observe that the
timing of rifting in western Baltic matches that of Scotland-Ireland very
well (Fig. 14; D-D’), and rift-related magmatism reached its paroxysm in
both around 600–580 Ma. Subsequently, passive platforms developed
along both margins, consistent with the opening of an ocean between
them. In contrast, while the onset of rifting in both the Northern and
Central segments of Laurentia is similar to that observed in western
Baltic (Fig. 14; B, C vs. D’), the rift-drift transition along the former

occurred later, in the early Cambrian (notably, we are also not aware of any proposed reconstruction of Baltica along the eastern margin of North America in Rodinia). Furthermore, ~575 Ma paleomagnetic data from Baltica and Laurentia indicate that those continents were separated by some 20° by that time (Fig. 4), revealing that a substantial ocean existed between them already then and thus precluding a later (Cambrian) breakup of those continents.

In Amazonia, small intrusions of ambiguous nature along the northwestern margin (in Ecuador and Colombia) present the only geological clues that the craton may have experienced the second phase of rifting. The Oaxaquia block (and possibly the Maya block), which was possibly adjacent to the northwestern margin of Amazonia in late Neoproterozoic time, was also invaded by a mafic dike swarm at ~619 Ma, suggesting it too experienced a phase of rifting then (Fig. 14) (Weber and Schmitt, 2019). Given the broadly similar age and composition of the Ediacaran mafic intrusive rocks of Oaxaquia-Maya and Ecuador to those of the Long Range dikes of Laurentia and the Egersund dikes of Baltica, they may all represent correlative expressions of the same magmatic event, i.e. the CIMP (Fig. 13 and 14) (Spikings et al., 2021; Tegner et al., 2019; Weber and Schmitt, 2019). The CIMP has been ascribed to a mantle plume and the distribution of its relics suggest that it impinged the lithosphere near the triple junction of Laurentia, Baltica and Amazonia (e.g. Piasarevsky et al., 2008; Fig. 13). As the emplacement of CIMP shortly pre-dated the apparent breakup of Scotland-Ireland and Baltica as just discussed (Fig. 14; D–D'), a straightforward interpretation is that the CIMP instigated or at least facilitated that breakup. But if northwestern Amazonia also exhibits relics of CIMP, do they likewise mark the breakup of Amazonia from Laurentia and Baltica? In other words, did CIMP induce or coincide with the margin-wide opening of three oceans (the West Iapetus, East Iapetus and Tornquist) or only some subset of them?

Relics of the second phase of rifting are not known from the southwestern margin of Amazonia, in contrast to the entire eastern margin of Laurentia against which Amazonia is usually reconstructed in Rodinia (Fig. 13). One potential explanation of that disparity is that Amazonia was never connected to Laurentia in Rodinia (e.g. Fig. 4g), but this would imply that the indications of the first rifting phase in the southwest of Amazonia and the proposed CIMP correlative in its northwest are simply coincidental. An alternative possibility is that while the northwestern part of Amazonia remained connected to Baltica and perhaps proximal to northeastern Laurentia, the southwestern part of Amazonia broke away from eastern Laurentia earlier, during the first rifting phase (Escayola et al., 2011; Robert et al., 2020). Amazonia itself has only one ambiguous paleomagnetic pole from this time interval, precluding a rigorous test of this hypothesis, but a richer selection of Ediacaran paleomagnetic data are available from West Africa which was probably coherent with Amazonia by that time (section 3.1.3). If indeed representative of Amazonia, these data allow Amazonia to have been located up to ~2500 km from Laurentia at 575 Ma, which would require a substantial ocean between them (Fig. 4h). Significantly, the data from earlier Ediacaran time (615 Ma), while permissive of a reconstruction of southwestern Amazonia against eastern Laurentia (i.e. no intervening ocean; Fig. 4a), are also compatible with alternative models where an intervening ocean was already open then (Fig. 4c) (Escayola et al., 2011; Robert et al., 2020). Clearly, additional paleomagnetic constraints are badly needed here. It is also important to underscore that all of these comparisons are predicated on the assumption that the West African data are used to indicate the orientation of Amazonia: if those blocks were separated in late Ediacaran time by an ocean (the Clymene Ocean in the sense of Trindade et al., 2006), the paleomagnetic constraints would not be applicable to Amazonia and the presence of that ocean could rather imply that Amazonia remained much closer (or still connected) to eastern Laurentia in the late Ediacaran (Fig. 4i, j). Importantly, in more recent models of the Clymene Ocean (e.g. Tohver and Trindade, 2014) that ocean does not pass between Amazonia and West Africa, but only separates West Gondwana from Central Gondwana. In these later models the Clymene Ocean does not pose any challenges to our reconstructions of Amazonia on the basis of data from West Africa; it does, however, pose challenges for the interpretation of other data, as will be discussed in section 5.1.3.

Further confounding an understanding of the Cryogenian-Ediacaran relationship between Laurentia and Amazonia is the fact that the southern margin of Amazonia, in contrast to its southwest margin, preserves evidence of mid-to-late Cryogenian-Ediacaran rifting. Accordingly, one interpretation of those relics along the southern margin of Amazonia could be that they represent the counterpart of the marginal rift successions in eastern North America. We note, however, that while poorly constrained in time, the rifting of the southern margin of Amazonia appears to mostly pre-date rifting along eastern Laurentia, and during most of the Ediacaran it was characterized by the deposition of platform carbonates (Fig. 11). The rifting in southern Amazonia could alternatively be paired with the more temporally coincident rifting in Scotland-Ireland (Dalziel, 1994), but that restoration would preclude the otherwise compelling links between western Baltica and Scotland-Ireland presented above. Intriguingly, an episode of Ediacaran rifting temporally coincident with that observed along the eastern margin of Laurentia seems to have affected the northern part of the Rio de la Plata Craton (Alto Caupacú block; Fig. 11). Given that contemporaneity, it is tempting to associate those rifting episodes, but ~575 Ma paleomagnetic data from Laurentia and Rio de la Plata rather suggest that those continents were significantly separated by then (Fig. 4h). As we discuss further below, geological observations from the Eastern Sierras Pampeanas, which likely flanked the west margin of Rio de la Plata by late Ediacaran time, likewise present problems for the interpretation that Rio de la Plata did not rift from eastern Laurentia until the mid-to-late Ediacaran. It is thus more likely that the late Cryogenian and Ediacaran tectonism observed along southern Amazonia and northern Rio de la Plata was associated with relative movements between those blocks, possibly linked to the assembly of West and Central Gondwana (e.g. Cordani et al., 2013).

All along the eastern margin of North America, the end of the second rifting phase was marked by an early Cambrian rift-to-drift transition and the commencement of a passive margin (Fig. 14 A, B, C). This clear and widespread event has thus been proposed to reflect the opening of the Iapetus Ocean (e.g. Williams and Hiscott, 1987). However, in the context of the aforementioned paleomagnetic data—namely the 575 Ma results that indicate that Baltica and (most likely) Amazonia were already separated from eastern Laurentia by a substantial ocean then (Fig. 4h)—an early Cambrian opening of the Iapetus is clearly not tenable. Moreover, the geological records from western Baltica and western Amazonia show no signs of early Cambrian breakup (Fig. 14 A–D'). Thus, given the available paleomagnetic data and the geologic constraints from the major continents, the Iapetus seems to have opened at least by mid-Ediacaran time. More specifically, the combined paleomagnetic data and the correlative early Ediacaran (CIMP) magmatic products in Laurentia, Baltica and Amazonia together present a compelling case for an early-mid Ediacaran opening of the East Iapetus. The sparse paleomagnetic data and geologic ties between eastern Laurentia and southwest Amazonia similarly imply that those two continents separated at least by early-mid Ediacaran time, but likely earlier, during the first rifting phase. Does this mean that the younger, early Cambrian rift-to-drift transition observed all along eastern North America reflects the opening of a distinct, younger ocean basin?

5.1.2. Insights from displaced terranes

Given the fragmentary late Neoproterozoic geologic and paleomagnetic constraints from Laurentia, Baltica and Amazonia, consideration of the coeval relics from the allochthonous blocks that may have lain between those continents is essential to unraveling the history of the oceans that opened among them. Evidence of the first phase of rifting is, to our knowledge, not apparent from any of these blocks, but relics of the second phase are known from Oaxaquia-Maya (as noted in the previous
section), Arequipa-Antofalla (AA), the Western Sierras Pampeanas (WSP) and the Precordillerà terrane (Fig. 12) (Appendix B). The contact between the AA and Amazonia is hidden below a thick pile of younger sediments, so it is not known whether the terrane remained attached to Amazonia during the opening of the Iapetus, or whether it rifted from Laurentia independently. Unfortunately, it is therefore unclear whether the rifting relics observed in the AA relate to relative movements between the AA and Laurentia, or between it and Amazonia (or both).

To the south, the AA may be continuous with the WSP (Fig. 10), which shares a similar basement and likewise records evidence of Ediacaran rifting (Appendix B). Although tenuous, this potential continuation of the AA southward into the WSP is an important observation, and another area in need of further research and scrutiny. This is because the WSP is now juxtaposed with the Eastern Sierras Pampeanas (ESP; Fig. 10), which was clearly flanked by a subduction zone from at least 555 Ma to 520 Ma (but possibly earlier; Fig. 14), and which is commonly assumed to have bordered the western Rio de la Plata Craton by late Ediacaran time. The termination of the active margin (Pampean arc) along the ESP has been interpreted to mark the arrival and accretion of the WSP in the early Cambrian (Casquet et al., 2018; Escayola et al., 2011). If, in late Ediacaran time, the AA was continuous with the WSP and Rio de la Plata was adjacent to Amazonia, the collision between the WSP and ESP would imply that the AA and Amazonia were independent in earlier times. In such a case, Amazonia could have rifted from eastern Laurentia in Tonian–early Cryogenian time, and WSP-AA could have rifted from Laurentia later, during the Ediacaran or early Cambrian. This would accordingly imply that two distinct basins opened in the mid-to-late Neoproterozoic: one between Amazonia and WSP-AA and another between WSP-AA and Laurentia. As an aside, we note that this introduces some terminological ambiguity: which of these basins was ‘Iapetus’? In drawing parallels between such a scenario and the paleogeographic framework of the Tethys system, we previously referred to these basins as the ‘PaleoIapetus’ and ‘NeoIapetus’, respectively (Robert et al., 2020). However, it is important to acknowledge that a similar model was earlier presented by Escayola et al. (2011), who named the ocean between Amazonia and WSP-AA the ‘Puncoviscana Ocean’ and that between WSP-AA and eastern Laurentia the ‘Iapetus Ocean’. Both to avoid further complicating the literature and to recognize the precedence of the names used by Escayola et al. (2011), we adopt their naming scheme herein.

A significant caveat with this two-ocean (Puncoviscana-Iapetus) model is that the connections inferred between 1) the AA and the WSP, 2) the ESP and Rio de la Plata, and 3) Rio de la Plata and Amazonia are all tenuous for Ediacaran time. For example, paleomagnetic tests of the coherency of the ESP and Rio de la Plata in Cambrian time have been obscured by the effects of younger tectonic rotations (Franceschinis et al., 2020; Franceschinis et al., 2016). Likewise, the Clymene Ocean model contends that Rio de la Plata and Amazonia were not amalgamated until Cambrian time, but this hypothesis was originally submitted on the basis of paleomagnetic data (see Puga cap carbonate pole in section 3.1.3 and Appendix A; Trindade et al., 2003, 2006) which have since been questioned (D’Agrilla-Filho and Cordani, 2017; Pisarevsky et al., 2008). Thus, given these uncertainties, we could alternatively assume that the AA and the WSP were coherent with Amazonia since their breakout from Rodinia. In this case, the Ediacaran rifting of WSP-AA from Laurentia would have also marked the breakup of Amazonia and Laurentia, and only one ocean (‘Iapetus’) would have opened between Laurentia and eastern Amazonia (‘NeoIapetus’) and between the various displaced terranes and the cratons they are now juxtaposed with of is great importance to understanding the evolution of Iapetus. The absence of evidence for Cambrian rifting in the AA and the WSP (as well as Amazonia) also make it difficult to directly link their separation from Laurentia to the margin-wide early Cambrian rift–drift transition found along eastern Laurentia. Moreover, unless a Clymene Ocean separated Amazonia and West Africa in the Ediacaran (i.e. in the sense of Trindade et al., 2006; Fig. 4i, j), the paleomagnetic data indicate that a substantial ocean separated Amazonia and Rio de la Plata from Laurentia by 575 Ma (Fig. 4h). This provides a further challenge to a Cambrian rifting of WSP-AA from Laurentia, because if the WSP collided with the ESP by ~530 Ma, a Cambrian departure from Laurentia would leave little time for WSP-AA to traverse that wide ocean.

Summarizing the aforementioned paleomagnetic and geologic arguments: the early Cambrian rift-to-drift transition along eastern North America was not likely produced by the rifting of Baltica or Amazonia, both of which likely broke away from Laurentia at least by early to mid-Ediacaran time, if not before. Although WSP-AA may have rifted from Laurentia independently and after the departure of Amazonia, it is not clear when its breakout occurred. However, WSP-AA accreted to western Amazonia and Rio de la Plata in early Cambrian time, so one possibility is that it broke away from Laurentia already in the early to mid-Ediacaran. In this case the spalling of some other block(s) must have been responsible for the early Cambrian rift-to-drift transition observed along the eastern margin of Laurentia: the Taconic ribbon terrane. As reviewed in section 4.1.5 (and Appendix B), this comprises blocks of Laurentian basement that appear to have rifted from the margin in late Ediacaran to early Cambrian time, before being re-accreted in the Cambro-Ordovician. Although a simple solution, this scenario presents a significant problem in that it suggests that rifting of the Taconic ribbon terrane commenced alongside an already open ocean basin between Laurentia and Amazonia and/or WSP-AA.

One elegant solution to this conundrum was presented by van Staal et al. (2013), who proposed that the Taconic seaway between the Taconic ribbon continent and the Laurentian autochthon may have been a hyperextended zone rather than a true oceanic seaway generated by normal seafloor spreading. In their model, the early Cambrian rift-to-drift transition marked the true opening of Iapetus, by the final rifting of WSP-AA from the outer, hyperextended margin of Laurentia. This model thus avoids invocation of a second spreading ridge beside an active one. Furthermore, the presence of such a hyperextended zone would imply that WSP-AA laid at some distance from the Laurentian autochthon in the late Ediacaran, which could explain how it arrived to the Western margin of Amazonia only a short time after rifting from eastern Laurentia. But, given the available paleomagnetic and geologic constraints, is this conceptual model feasible in the context of uniformitarian tectonics? This presents an excellent opportunity to apply quantitative plate tectonic modelling, and we will return to this question in section 5.2.

5.1.3. Peering further with detrital zircons

Thus far, our integrated analysis has revealed that while paleomagnetic and geologic data point to an early to mid-Ediacaran breakup between northeastern Laurentia and western Baltica, the time at which Amazonia rifted from eastern Laurentia remains a critical uncertainty. This is unsurprisingly due to the scarcity of late Neoproterozoic records from western Amazonia, and the near-complete lack of paleomagnetic data (one ambiguous Ediacaran pole). In lieu of stronger paleomagnetic-geologic constraints, we now turn to the detrital zircon records from Amazonia and Laurentia to see if we can glean any further inferences concerning their breakup.

Prior to continental breakup, one might expect rifted margin sediments from conjugate margins to exhibit similar detrital zircon age distributions (Cawood et al., 2013; Olierook et al., 2019), whereas after breakup, the age distributions of the marginal sedimentary packages would be expected to diverge (assuming sedimentary recycling to be imperfect). This presents a useful, if qualitative test to further investigate the breakup timing of Laurentia and Amazonia. Ediacaran and early Cambrian rift-related sediments deposited along the eastern margin of Laurentia are characteristically dominated by ~1.2–1.0 Ga detrital zircon age fractions, interpreted to be shed from the Grenville orogen (e.g. Thomas et al., 2017). Unfortunately, comparably-aged sediments are not preserved along western Amazonia (section 4.3 and Appendix B).
However, the western margin of Amazonia was active since at least ~500 Ma (the start of the Famatinian cycle), so if late Neoproterozoic and/or early Cambrian sediments were indeed deposited there, we could expect a portion of their detritus to have been recycled into younger deposits along the margin during Famatinian orogenesis. Ordovician and younger sedimentary rocks from along that margin exhibit detrital zircon age distributions with large ~1.2–1.0 Ga and ~650–500 Ma age fractions (Fig. 11) (Robert et al., 2020). The older of these two age fractions is similar to that identified in the Ediacaran-Cambrian sediments in Laurentia, and could reflect sediment transport from Laurentia, although they could alternatively have been shed from the Sunnás orogen in Amazonia. In contrast, the younger age fraction has no significant counterpart in the Laurentian sediments of Ediacaran–Cambrian age, which suggests that their original transportation to western Amazonia may post-date the rifting of Amazonia from Laurentia.

Three distinct sources of those ~650–500 Ma detrital zircons have been proposed: (1) igneous rocks generated by Iapetan rifting, (2) the pan-African belts related to the formation of Gondwana and (3) a hidden magmatic arc that formed along the western margin of Amazonia (Chew et al., 2008a, 2008b; Robert et al., 2020). Critically, these different proposed sources present different implications for the timing of Amazonia-Laurentia breakup, but it is unfortunately not yet clear which is correct, as each interpretation presents some difficulties. Proposal (1), which would imply that Amazonia rifted from Laurentia during the Ediacaran or Cambrian, is problematic in that magmatic products generated during rifting are usually dominantly mafic in composition, and thus zircon poor (Cawood et al., 2012). Additionally, there are virtually no potential host rocks (i.e. 650–500 Ma rift-related rocks) exposed along most of western Amazonia; the minor intrusive bodies in Ecuador and Colombia (section 4.3.1) being perhaps the only possible examples. This proposal also fails to explain why such a peak is not found in the rift-related Ediacaran-Cambrian sediments of Laurentia, where rift-related magmatism is widely exposed. Interestingly, there is no clear zircon age peak of that age in synrift sediments of west Baltica either, despite that it too contains a large volume of rift-related magmatic products (Bingen et al., 2011; Bingen et al., 2005; Lamminnen et al., 2015).

Concerning proposal (2), the pan-African belts closest to the western margin of Amazonia are the Neoproterozoic Tucavaca and Paraguay belts along the southern margin of Amazonia, and the Araguaia and Brasilia belts that divide eastern Amazonia from the São Francisco craton (Fig. 10). The Tucavaca and Paraguay belts consist predominantly of folded sedimentary successions with only minor magmatic rocks (section 4.3.3 and Appendix B), and are therefore unlikely the source of the abundant zircons found all along the western margin of Amazonia (Chew et al., 2008a, 2008b). The Araguaia belt, which is also dominantly made of metasedimentary rocks, contains several Neoproterozoic igneous bodies (Hodel et al., 2019), but they are mafic-ultramafic in composition and so, being zircon poor (Cawood et al., 2012), are unlikely to be the main source of the detrital zircons. By contrast, in the Brasilia belt, the Goiás magmatic arc contains a large volume of 900–630 Ma calc-alkaline metavolcanic and plutonic rocks ranging in composition from gabbro to granite, which are intruded by 630–540 Ma bimodal intrusions (e.g. Pimentel, 2016). This magmatic arc could thus conceivably have been the zircon source, although one difficulty with this interpretation is the general absence of 900–650 Ma zircons in the western Amazonia sediments. Another caveat is that the Goiás arc developed on the margin of the São Francisco craton and so could not have supplied detritus to Amazonia before colliding with it. The timing of Amazonia–São Francisco collision is strongly debated: some have proposed they collided by 620 Ma (Ganade de Araujo et al., 2014), whereas others contend that event did not occur until the early Cambrian (Tolher and Trindade, 2014; Trindade et al., 2006), and that a wide, Clymene Ocean separated Amazonia and São Francisco during the Ediacaran.

In the scenario of an early collision between Amazonia and São Francisco (e.g. Ganade de Araujo et al., 2014), zircons from the Goiás arc could have spread across Amazonia from the early Ediacaran (~620 Ma), whereas in the scenario of a late collision (e.g. Trindade et al., 2006) they could have only appeared in Amazonian sediments from Cambrian time. Sedimentary successions deposited on Amazonia and directly adjacent to the Goiás magmatic arc in the Paraguay belt do not display 650–550 Ma zircons in Precambrian strata, whereas a substantial population of such zircons are found in early Cambrian and younger sediments there (e.g. McGee et al., 2018). This may support the notion that São Francisco and Amazonia were not in contact prior to Cambrian time. This scenario might be further supported by a paleomagnetic pole from the Arabia–Nubia Shield (Nairn et al., 1987) if transferred to São Francisco (as part of Central Gondwana), which supports the separation of São Francisco from Amazonia (in West Gondwana) at ~590 Ma. However, as mentioned in section 3.1.3, that block is separated from the rest of the West and Central Gondwana blocks by several orogenic zones (e.g. Siberides) that could have still been active in Ediacarian time (e.g. Sengör et al., 2020), which may render this pole inappropriate to use for constraining the position of São Francisco. Returning to a detrital zircon perspective: some Ediacaran rocks in the Argentine Puna exhibit a distinct 650–570 Ma detrital zircon age fraction, which could indicate that zircons of this age were already shed across Amazonia in Ediacarian time (Naidoo et al., 2016), but the exact tectonic affinity of these rocks (whether belonging to the WSP or ESP) is unfortunately unclear.

If the Brasilia belt/Goiás arc was indeed the source of the 650–500 Ma zircons in sediments of western Amazonia, the age of the first westward flux of those zircons bears strong implications for Laurentia–Amazonia breakup. As mentioned above, the absence of late Neoproterozoic zircons in Ediacaran–Cambrian sedimentary rocks along eastern Laurentia could imply that Laurentia had already broken away from Amazonia before the zircons reached western Amazonia. Thus, a collision of Amazonia and São Francisco by ~620 Ma would suggest that Amazonia rifted from Laurentia before the early Ediacaran (Robert et al., 2020), whereas in the case of a Cambrian collision, Laurentia and Amazonia could have broken up any time prior to the mid-Cambrian. A further complication is the possibility that the zircons are even further-traveled. For example, arc magmatic rocks of ~640–570 Ma age are known from Avalonia, which may have flanked the northern margin of Amazonia in the late Neoproterozoic (Nance et al., 2008). Therefore, the further scrutiny of these zircon age fractions and specifically a determination of their origins and transport history present key areas for future research.

The final proposal (3) attributes the 650–500 Ma zircons to a local (west Amazonian) magmatic arc, since destroyed or at least hidden by younger deposits (Chew et al., 2008a, 2008b; Robert et al., 2020). This scenario would easily explain the abundance of 650–500 Ma zircons in Paleozoic sediments along this margin, and—in obviously necessitating a separation of eastern Laurentia and western Amazonia by this time—also explains the absence of zircons of this age in Ediacaran–Cambrian sediments along eastern Laurentia. This scenario would furthermore imply that Amazonia must have separated from Laurentia at least some tens of Myrs prior to the start of subduction at ~650 Ma, such that subduction could initiate and remain sustained, and that the intervening oceanic basin would not immediately collapse (Robert et al., 2020). However, while this proposal is appealing in its simplicity, the absence of any direct evidence of the hypothesized arc renders it speculative. Nevertheless, there may still be opportunities to further interrogate this important hypothesis (Robert et al., 2021) by looking at the position of an arc along the western margin of Amazonia could be expected to partially reset zircons derived from the Amazonian basement (and in contrast to the Goiás arc which would have affected the São Francisco basement). This would yield a discordant population of zircons with upper concordia intercepts coincident with Amazonian basement ages and lower concordia intercepts coincident with the age of the arc. Interestingly, Spikings et al. (2021) identified numerous lower intercepts ranging from 500 to 555 Ma that may support the existence of a
pre-Famatinian arc (> 500 Ma) in northwest Amazonia.

5.2. Our preferred scenario

The previous sections have underscored that the existing paleomagnetic and geologic constraints on the opening of Iapetus are sparse, fragmentary and often ambiguous. Even when considering these constraints jointly, we are left with many outstanding unknowns concerning the history of the opening of the Iapetus. Despite these important uncertainties, we consider it useful to attempt to integrate these data into a simple plate model to exploit the often subtle but nevertheless powerful

constraints imposed by plate tectonics (Domeier and Torsvik, 2019). Here we describe our first-order preferred scenario of Iapetus Ocean opening in the form of a plate model of Laurentia, Baltica and Amazonia (West Gondwana) extending from 750 to 520 Ma. Instantaneous snapshots of this model are depicted in Fig. 15 and the model details are provided in Appendix C. For each timestep, the model is displayed in a Laurentia-fixed coordinate system (Fig. 15, left column) and in the paleomagnetic reference frame (Fig. 15, right column). The former provides a convenient means of highlighting geological processes related to relative plate motions, whereas the latter displays the model’s compatibility with the existing paleomagnetic data and the attendant

Fig. 15. Snapshots of our regional plate model illustrating the evolution of the Iapetus system from 750 to 520 Ma (a-h and j-l), and an alternative scenario for the opening of west Iapetus (i; discussed in the text). Reconsructions of our preferred model are displayed in a Laurentia-fixed coordinate system (position of Laurentia at 750 Ma kept fixed) in the left column, and in the paleomagnetic reference frame in the right column. In the Laurentia-fixed reconstructions: a mobile grid (grey) indicates the latitude predicted by paleomagnetic data and the South Pole (SP); mid-ocean ridges are red, transform faults in green and subduction zones in blue; main intracontinental fault systems are shown in grey; the Euler poles (EP) describing the main kinematic rotations with respect to Laurentia are displayed by red filled circles and subscripts indicate which block they refer to (A = Amazonia, B = Baltica, P = Precordillera, W = Western Sierras Pampeanas); The Central Iapetus Magmatic Province (CIMP) is illustrated by a large light red circle along with another presumed plume-related magmatic province at 750–700 Ma. Continental block abbreviations: AA = Arequipa-Antofalla; ESP = Eastern Sierras Pampeanas; H = Hoggar; OM = Oaxaquia and Maya blocks, PC = Precordillera terrane; RP = Rio de la Plata; S = Sahara metacraton; SF = Sao Francisco; TR = Taconic ribbon continent; WSP = Western Sierras Pampeanas. In the paleomagnetic reference frame, reconstructions include the plate contours, velocity field and the paleomagnetic poles used. Continental blocks constrained by paleomagnetic data at a given time are colored, otherwise they appear grey. Pole abbreviations are as in Table 1. Size of velocity arrows is changing among snapshots and is given in the lower right hand side (note that no velocity arrows are shown in panel l because it coincides with the temporal edge of the model). In Fig. 1, the alternative scenario for the opening of west Iapetus is shown at 590 (left) and 550 Ma (right), and both snapshots are in Laurentia-fixed coordinates. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Quantitative plate velocities are also depicted in Fig. 16 for each of the major continents.

5.2.1. Late Tonian (750–720 Ma)

At the beginning of our model, at 750 Ma, only one paleomagnetic constraint is available: the Kwagunt Formation pole of Laurentia (Fig. 3a). This pole suggests that Laurentia was positioned at equatorial latitudes (Fig. 15a). Another paleomagnetic pole is available at 720 Ma, from the Franklin large igneous province, which constrains Laurentia to a similar position but slightly more south. Because this time interval coincides with the first onset of the earliest phase of rifting, we assume that Baltica and Amazonia were still effectively united with Laurentia, and so can reconstruct those continents by restoring them to their positions (relative to Laurentia) in Rodinia. As elaborated above, several pre-rifting fits of Rodinia are permissible according to the available paleomagnetic data (e.g. Li et al., 2008 vs. Dalziel, 1992), but we start our model at 750 Ma with the fit of Li et al. (2008) (Fig. 15a) according to: (1) indications of the first phase of rifting in both western Amazonia and the central Appalachians, (2) the occurrence of ~615 Ma intrusives correlated with the CIMP in the Northern Segment of eastern Laurentia, western Baltica and northwest Amazonia, and (3) the similarity of the geological records of the second rifting phase in Scotland-Ireland and western Baltica. We emphasize that these inferred links do not invalidate...
alternative configurations of Baltica and Amazonia relative to Laurentia (e.g. Dalziel, 1992; Evans, 2009; Slagstad et al., 2019), but we consider them less congruent with the geological observations. In that we are now entertaining kinematic arguments, we would also point out that the model of Evans (2009) necessitates a comparatively more complicated Cryogenian kinematic evolution: the constitutive blocks of West and Central Gondwana, after rifting from the western margin of Laurentia at ~780–720 Ma, have to travel around Laurentia to end up in the proximity of eastern Laurentia at ~615 Ma (Fig. 4g). Given the absence of strong evidence of such large relative motions, we did not favor this fit of Rodinia. As explained in section 3.1.3, we also treat West Africa and Amazonia as united by this time, although we implement minor convergence between the two cratons later during the late Ediacaran to take into account deformation related to the Rokelide and Gurupi orogenies (Klein et al., 2020; Villeneuve, 2008) that we interpret to occur in an intracontinental context.

5.2.2. Cryogenian (720–635 Ma)

On the basis of relics of the first rifting phase in eastern Laurentia and southwest Amazonia, we infer that Amazonia had already started rifting from Laurentia in the late Tonian (from 750 Ma). This rifting phase could have been triggered by a mantle plume impinging the Central
Segment of eastern Laurentia at ~765–700 Ma (Fig. 15a,b) (McClellan and Gazel, 2014). In the model, we interpret this first rifting phase to lead to the separation of Laurentia and Amazonia at ~700 Ma and to the opening of the Puncoviscana Ocean. In addition to the preserved rift relics, this scenario is supported by (1) the absence of 650–500 Ma zircons in Ediacaran–Cambrian sediments of eastern Laurentia and (2) the interpretation that their dominance in early Paleozoic sediments of western Amazonia reflects that an arc of such age was present along that margin (which has since been destroyed or obscured). The latter interpretation obviously necessitates that western Amazonia was flanked by a sizable ocean basin by at least 650 Ma, such that subduction could be sustained for an extended period. There is no paleomagnetic data to evaluate this hypothesis in Cryogenian time, but early Ediacaran paleomagnetic data permit that an ocean up to ~2500 km could have existed between Laurentia and Amazonia by 615 Ma (Figs. 4c, 15d). In our simplified scenario, we assume that by 650 Ma the width of that ocean basin was already approaching its zenith at 615 Ma, implying that the drifting velocity of Amazonia from Laurentia from 700 to 650 Ma was considerably reduced after 650 Ma (i.e. following the onset of subduction between them). We therefore imposed that Amazonia first drifted from Laurentia at a rate of ~4 cm/yr from 700 to 650 Ma, comparable to spreading rates from the Atlantic Ocean (Fig. 16a),
followed by a slowdown to 1.5 cm/yr to reach a paleogeography at 615 Ma that is paleomagnetically constrained (Fig. 15d). Given that subduction started along western Amazonia at 650 Ma and that the Puncoviscana ridge subducted by 615 Ma (see next section), these kinematic assumptions constrain the subduction and ridge velocities inside the oceanic domain between 650 and 615 Ma. In our model, it implies that the subduction zone along western Amazonia had a convergence rate of ∼4.3 cm/yr and that the ridge had a full-spreading rate of ∼5.8 cm/yr, which are reasonable relative plate velocities.

Despite the evidence for an early breakup of Laurentia and Amazonia, the geological record from Scotland-Ireland and the western Baltic autochthon is more consistent with slow, progressive continental extension during this time—notably according to the observation that sedimentation along parts of both margins continued largely uninterrupted during the late Tonian and Cryogenian (Fig. 14). Again, there are no Cryogenian paleomagnetic data to test this interpretation, but 615 Ma data from both blocks are consistent with them having remained juxtaposed in Cryogenian time (Fig. 4a-c). To accommodate both the opening and spreading of the Puncoviscana Ocean and the development of a wide rifted continental zone between northeastern Laurentia and western Baltic since 700 Ma, we model the motion of both Amazonia and Baltic (relative to Laurentia) by rotations about the same Euler pole located close to the eastern margin of Greenland (Fig. 15b and c). For the same angular displacement, this Euler pole results in a comparatively small relative motion of Baltic (being closer to the pole) and a larger motion of Amazonia. However, because we infer that the rifting of Amazonia resulted in true continental breakup and seafloor spreading (whereas the shared margin between Laurentia and Baltic was only a wide rifted zone), the ridge that developed between Laurentia and Amazonia must continue as a transform boundary between Amazonia and Baltic (because they share the same relative pole of rotation). In our model, the continuous and slow extension between Laurentia and Baltic occurs at a very low rate of about ∼1 cm/yr from 700 to 590 Ma (Fig. 16a).

5.2.3. Early Ediacaran (635–590 Ma)
At 615 Ma, paleomagnetic results are available from Laurentia, Baltic and West Gondwana (Fig. 3). The data from Laurentia suggest it was positioned at low latitude, while Baltic and West Gondwana were positioned at mid-latitudes. As discussed in section 3.2.2, these few constraints are permissive of a variety of relative configurations of these three continents (Fig. 4a-c). Because our interpretation of the Cryogenian tectonics implies that a wide ocean separated Laurentia and Amazonia at 650 Ma while Baltic was still attached to both Amazonia and Scotland-Ireland, we consider the 615 Ma reconstruction in Fig. 4c to be the most consistent with our previous 650 Ma reconstruction (Fig. 15c and d).

A key event associated with the inception of the second rifting phase
at 615 Ma was the emplacement of the Central Iapetus Magmatic Province (CIMP). This resulted from the impingement of a mantle plume on the lithosphere near the junction of Laurentia, Baltica and West Gondwana. Associated magmatic expressions include the Long Range dikes in Newfoundland and Labrador, the Egersund dikes in southwest Norway and mafic dikes in Ecuador and the Oaxaqua block of Mexico (Fig. 13). We reiterate that the spatial distribution of these features is broadly supportive of the relative configuration of Laurentia, Baltica and West Gondwana that we present at this time, and thus of the earlier framework that it evolved from (i.e. the Rodinia fit of 11 et al., 2008; in contrast to the other starting frameworks discussed). It is also noteworthy that the occurrence of CIMP reliefs in the autochthonous parts of eastern Laurentia and western Baltica suggest those two continents were proximal to one another and the plume conduit, whereas in Amazonia, CIMP reliefs are only known from allochthonous domains, and so that block may have been further removed (as in our model). The emplacement of CIMP-related magmatic products likely continued until ~590 Ma in Scotland-Ireland and western Baltica, but also in the Northern Segment of Laurentia that was intruded for instance by the 590 Ma Grenville dyke swarm in Canada. Because magmatism in the Central and Southern segments of Laurentia did not start before ~570 Ma, the geographical extent of magmatic products that we relate to CIMP (or at least its ~615–590 Ma expression) are largely restricted to the ‘East Iapetus’ domain, and we estimate its magmatic center to have lain between the Northern Segment of Laurentia, Scotland-Ireland and western Baltica (Fig. 14, 15d and e). Among the rare CIMP reliefs in the ‘West Iapetus’ domain, the dikes of the Oaxaqua block are a particularly interesting case. Oaxaqua is usually reconstructed along the northwestern margin of Amazonia during the late Neoproterozoic and so, in the context of our model, would imply that CIMP extended across the Puncoviscana Ocean to West Gondwana. Alternatively, van Staal et al. (2020) proposed that the Oaxaqua block could have lain along the Northern Segment of Laurentia, in a closer position to the magmatic center of CIMP (Fig. 15d, see upper left insert).

The cessation of CIMP in the East Iapetus domain coincided with the development of a passive margin in Scotland-Ireland and western Baltica that we interpret as marking the breakup between Laurentia and Baltica by ~590 Ma (Fig. 14) and thus the opening of the eastern Iapetus Ocean (Fig. 15e and f). If and how a contemporaneous ocean basin opening occurred in the West Iapetus domain is unclear. If CIMP was indeed a major trigger for the opening of the East Iapetus, a different evolution could be expected in the West Iapetus domain, where CIMP reliefs are few (and unknown to the Central and Southern segments of Laurentia) (Fig. 14). Such a distinct evolution could also be anticipated from the eventual subduction of the Puncoviscana ridge below western Amazonia, which could have strongly re-shaped the West Iapetus domain. According to our simple kinematic model, as spreading of the Puncoviscana Ocean and subduction along the western margin of Amazonia continued (from 650 Ma), the ridge would have migrated progressively closer to the subduction zone until ultimately being consumed by it (Fig. 15b-d). We suggest that this ridge-subduction event occurred sometime prior to ~615 Ma, and that continued subduction thereafter instigated the second rifting event observed along the eastern margin of Laurentia (Fig. 15d-e). Specifically, we contend that once the Puncoviscana ridge was subducted and Laurentian oceanic lithosphere (the remnant half of the Puncoviscana Ocean basin) entered the trench along western Amazonia, the eastern edge of the Laurentian continent would no longer have been subject to being slab-driven, and would have first induced extension along the eastern margin of Laurentia and ultimately the spalling of a marginal ribbon terrane, similar to the scenario proposed for the opening of the Neo-Tethys (Robert et al., 2020; Stampfl and Borel, 2002).

The conjugate effect of slab pull and CIMP emplacement may have triggered the second rifting phase in the West Iapetus domain (Escayola et al., 2011; Robert et al., 2020), but the resulting breakup timing remains unclear. A first possibility is that the rifting led to the detachment of WSP-AA concurrently with the breakup of Laurentia and Baltica around ~590 Ma (Fig. 15i), resulting in a synchronous opening of the entire Iapetus Ocean (Fig. 13). However, this scenario struggles to explain the distinct rifting history in the eastern and western Iapetan domains, and notably the continuation of rifting in the latter domain until the latest Ediacaran-early Cambrian (Fig. 14). Another possibility is that rifting in the West Iapetus domain did not immediately lead to continental breakup, but was rather associated with a protracted interval of marginal extension including the development of hyper-extended and hyper-thinned domains before continental breakup ultimately occurred at ~550–530 Ma (Fig. 15d-h) (van Staal et al., 2013). We favor this second option because it better explains the rifting history along eastern Laurentia, but a further dissection of these two alternative models is elaborated below, in the ‘Late Ediacaran’ subsection.

For the sake of parsimony, we continue to model the motion of Baltica and West Gondwana during this time as rotations of variable magnitude about the same Euler pole fixed relative to Laurentia. This means that the motion between Baltica and West Gondwana remains small, and purely transient in the model (Fig. 15c). Although simple, these kinematics are consistent with the observation of asymmetric magnetic fabrics in the Egersund dikes, which suggest those 615 Ma dikes were emplaced under a tectonic regime of sinistral transtension (Montalbano et al., 2016). Furthermore, the trace of small circles centered on this Euler pole are locally parallel to the transform faults that developed along the Iapetan rifted margin of Laurentia during the late Neoproterozoic (e.g. Thomas et al., 2017), and so while simple, our kinematic model is nevertheless consistent with these large-scale marginal structures.

### 5.2.4. Mid-Ediacaran (590–565 Ma)

Relatively numerous paleomagnetic data are available from Laurentia, Baltica and West Gondwana in the mid-Ediacaran, but between 615 and ~575 Ma, and then again between ~575 and ~565 Ma, those data reveal shared, rapid directional changes of ~90° (Fig. 3). As described in section 3.2.2, the fitting of these excursions to a common track results in fully-defined relative continental reconstructions (i.e. both relative latitude and longitude) at 575 and 565 Ma (Fig. 4h), which are broadly similar to the framework of our preferred model at 615 Ma (Fig. 15d). However, a conventional interpretation of those paleomagnetic data would indicate that Laurentia was positioned at the pole at 590–575 Ma (Fig. 15e and f), and then at the equator by 565 Ma (Fig. 15g). The data likewise imply that the positions of Baltica and West Gondwana abruptly shifted (by a ~90° rotation) across that short interval, but they both started and ended at mid-latitudes between 615 Ma and 565 Ma, having quickly traversed higher latitudes in between. In Fig. 15d-g (right panels), we depict our model according to this ‘literal’ interpretation of the paleomagnetic data involving two successive phases of rapid plate motion at 615–590 Ma and 575–565 Ma, occurring at rates of ~45 cm/yr and ~90 cm/yr, respectively (Fig. 16b). As detailed in section 3.2.1, we consider the paleomagnetic data at these times to be dominated by some global, non-uniformitarian process (e.g. rapid TPW or anomalous magnetic field behavior). Because we are herein primarily interested in the relative configuration of the continents through time, we have reconstructed the continents in absolute latitude according to the conventional interpretation of the data for the sake of simplicity (and for lack of a better defined reference system). We stress that we do not expect these particular absolute reconstructions to be accurate, but it is illuminating to see that the data themselves would imply if interpreted in this conventional way.

On the basis of the detrital zircon record, we infer that subduction continued beneath western Gondwana during the mid-Ediacaran. Following subduction of the Puncoviscana ridge by ~615 Ma, the lithosphere that next entered the subduction zone would have belonged to the Laurentian plate and so it could be anticipated that Laurentia and Amazonia would approach one another from 615 Ma. However, the paleomagnetic data rather suggest that those continents grew farther...
apart between 615 and 575 Ma. In our model, this is explained by sustained continental extension and thinning of the eastern margin of Laurentia (Fig. 15 d-g). We specifically model WSP-AA as the leading edge of this extending margin, which presents a diffuse plate boundary. From 615 Ma, WSP-AA is treated as semi-independent of the Laurentian interior, moving in concert with the subducting (remnant) Puncoviscana lithosphere. Assuming a very low rate of subduction beneath western Gondwana (~1.2 cm/yr), this scenario requires that the eastern margin of Laurentia was extended by ~2 cm/yr to reach ~800 km extension in total by 575 Ma (Fig. 16a). In the geological record, wide rift systems are common, but are generally narrower (<800 km) and extend more slowly (<1 cm/yr) than our modelled example (e.g. Svarnani Dias et al., 2015), but these differences are insignificant given the uncertainties on the paleomagnetic data. A corollary conclusion is thus that the available paleomagnetic constraints are consistent with the second rifting phase being associated with the development of a wide rifted margin along the West Iapetus domain (van Staal et al., 2013). To describe the motion of WSP-AA, we again adopt the simplistic approach of employing the same Euler pole used to describe the motion of Baltica and Amazonia (relative to Laurentia). As before, this ensures that the motion between WSP-AA (or rather the remnant Puncoviscana plate) and Baltica is purely transient, and that its motion with respect to Amazonia is purely convergent. As aforementioned, such kinematics are also consistent with the reconstructed orientation of the major transform faults that were then developing along the eastern margin of Laurentia (e.g. Thomas et al., 2017). The specific angular rotations of Baltica, Amazonia and WSP-AA, although individually different, are kept constant throughout this interval (up to 565 Ma).

5.2.5. Late Ediacaran (565–541 Ma)

In late Ediacaran, paleomagnetic data are again available from Laurentia, Baltica and West Gondwana (Fig. 3b). Remarkably, while the paleomagnetic data from Laurentia and West Gondwana appear to indicate that those blocks remained in approximately the same positions from ~565 Ma to ~540 Ma (i.e. reflecting only moderate continental drift), the late Ediacaran data from Baltica exhibit another dramatic directional excursion. Outwardly, that excursion strongly resembles the other ones observed in the Baltica dataset from 615 to ~565 Ma (Fig. 17), and the stage pole of the simplest kinematic solution from ~565 to ~550 Ma plots less than 30° away from the 615–590 Ma and 575–565 Ma stage poles. This similarity might indicate that this later excursion results from the same mechanism responsible for the two previous ones. A critical distinction though is that the excursions from 615 to 565 Ma are also observed in the datasets of Laurentia and West Gondwana and so seem to reflect global motions, whereas the 565–550 Ma one is solely observed in the dataset of Baltica. However, the dataset that defines this excursion in Baltica is considerably more numerous (six poles) than the data from Laurentia (one coeval pole), and the data from West Gondwana are slightly younger (early Cambrian) (Fig. 3b). It is therefore reasonable to hypothesize that these late Ediacaran data from Baltica capture another signal of some non-uniformitarian process (e.g. rapid TPW or anomalous magnetic field behavior) that has not yet been recognized in the datasets from Laurentia and West Gondwana. Alternatively, if the late Ediacaran paleomagnetic data from Baltica are taken at face value and used to reconstruct Baltica at 550 Ma, it is clear that Baltica must have undergone a dramatic rotation involving a strong component of toroidal motion between 565 and 550 Ma (e.g. compare Fig. 4h and k). Again, taking a simplistic approach, we can kinematically accommodate that motion of Baltica by allowing it to rotate about an Euler pole close to the one used at earlier times, such that its relative motion with respect to Laurentia remains dominantly divergent, and its relative motion with respect to West Gondwana remains mainly transient (Fig. 15h). Such a configuration imposes a very fast motion of Baltica away from Laurentia, at a velocity of about ~25 cm/yr (Fig. 16a). Such a velocity is atypical with respect to plate speeds of the last 200 Myrs (Zahirovic et al., 2015); a similar velocity only being observed for India, which reached a speed of ~18 cm/yr during the Cretaceous-Paleogene. Interestingly though, the southern margin of Baltica, which in our model is juxtaposed with Amazonia at ~565 Ma (Fig. 15g), experienced widespread igneous activity between 580 and 550 Ma that has been ascribed to rifting (Poprawa et al., 2020). That magmatism, along with slightly older carbonatites emplaced at ~580 Ma (Meert et al., 2007) in southern Norway and Sweden, could potentially relate to a change in the relative motion between Baltica and Amazonia and the opening of the Tornquist Ocean by ~565 Ma, as in our preferred model. More speculatively, the ~580–565 Ma rift-related mafic rocks in Ecuador described by Spikings et al. (2021), in being younger than other relics of CIMP, might alternatively relate to such strong relative motion between Amazonia and Baltica at this time. Thus, in spite of the very fast motion of Baltica implied by this simple kinematic solution, we adopt it in our model as it fits the presently available paleomagnetic data and is, at least, not in violation of the geological observations. However, we again underscore that this positioning of Baltica at 550 Ma appears remarkably similar to the pattern that emerges from the previous global excursions, and would be less geodynamically problematic if interpreted as such (Fig. 17).

Along the Central and Northern segments of eastern Laurentia, the second rifting phase reached its climax in the late Ediacaran-early Cambrian (Fig. 14). We entertain two alternative interpretations of this event, which we call models A and B. In model A, which is the variant that we have adopted in our preferred model from 615 to 550 Ma, the second rifting phase resulted in the development of a wide rifting zone associated with the development of hyperthinned and possibly hyperextended domains along the margin of eastern Laurentia (Fig. 15d-g). This hyperthinning/hyperextension could have resulted in the detachment of several independent crustal blocks (hanging-blocks or ‘H-blocks’: Lavier and Manatschal, 2006), possibly separated by embryonic oceanic crust (van Staal et al., 2013), with WSP-AA being the most outboard. Ultimately, continued slab-pull from the subducting Puncoviscana lithosphere led to marginal rupture and the initiation of
sustained spreading between WSP-AA and Laurentia, and thus the opening of the West Iapetus. Any crustal H-blocks further inboard than the West Iapetan ridge would have been left behind as the remnant leading edge of Laurentia. Van Staal et al. (2013) suggested that the Taconic ribbon terrane comprised such inboard H-blocks, separated from the Laurentian craton by 300–1000 km of highly attenuated continental crust and embryonic oceanic crust (the Taconic Seaway). Thus, according to this model, an embryonic West Iapetus and Taconic Seaway developed in parallel during prolonged crustal extension, although only the former ultimately evolved into a true oceanic spreading system by the end of the second rifting phase at ~550–530 Ma (Fig. 15h). By contrast, in model B, the second rifting phase along eastern Laurentia (Fig. 14) represents two temporally distinct phases of breakup, leading to the opening of the West Iapetus and the Taconic Seaway at two different times (Cawood et al., 2001). This interpretation implies that the West Iapetus ocean would have formed during the early Ediacaran (~615–590 Ma) at about the same time as the opening of the East Iapetus branch (Fig. 15i left), whereas the Taconic seaway would have opened at ~550–530 Ma (Fig. 15i right).

How do these alternative models compare with the geological record? The second rifting phase in the West Iapetan domain was accompanied by widespread rift-related magnetism (Fig. 14), which poses something of a challenge to the interpretation that the eastern margin of Laurentia was hyperthinned/hyperextended (i.e. Model A) because such margins are generally magma-poor. Nevertheless, van Staal et al. (2013) argue that the use of end-member models is generally inappropriate because magmatic activity within rifted margins usually varies in space and time as they evolve. It is therefore useful to consider the margin segments individually. Geologically, the Central Segment of Laurentia is characterized by a thick pile of Ediacarian synrift sediments, and magmatic activity was restricted to the interval ~570–560 Ma (Fig. 7). This record appears consistent with the continuous, protracted Ediacaran rifting suggested by Model A. In the more magma-rich Northern Segment, synrift sedimentation was apparently more episodic, occurring in the early Ediacaran and the late Ediacaran (Fig. 7 and 14). Such a two-stage synrift sedimentation record is better explained by Model B, whereas Ediacaran magmatic activity in the Northern Segment, which was more or less continuous from ~615 to 550 Ma, is less readily explained by the two successive breakup events of that model.

Beyond those geological considerations, the main problem posed by Model B is of geodynamic nature. The two successive continental breakup events in that model necessitate the development of two parallel mid-ocean ridges (Fig. 15i, right), which—according to the view that ridges mostly evolve as a passive response to regional tectonic forces induced by slab pull—is difficult to explain geodynamically. One way to solve this problem is to assume that the initial (older) ridge, the West Iapetus ridge, was subducted before the subsequent (younger) ridge, the Taconic ridge, had formed. In such a case, there would have only been one active ridge in the West Iapetan domain at any given time. However, the arrival of WSP-AA to the western margin of Amazonia occurred around ~530 Ma, which implies that the West Iapetus ridge must still have been active when the Taconic seaway opened in the latest Ediacaran/early Cambrian. The coexistence of two parallel ridges seems then unavoidable in Model B, and we therefore deem it geodynamically unlikely. In contrast, Model A does not face this problem as only one mid-ocean ridge would have subsided after the terminal rupture of the hyperthinned/hyperextended margin of Laurentia (Fig. 15h). In this model, the opening of the Taconic seaway can be attributed to some combination of delocalized extensional deformation, mantle exhumation along major lithospheric-scale faults and/or embryonic spreading that occurred within the very wide rifted margin of Laurentia. These processes would have occurred synrifting but growth of the Taconic Seaway would have ceased at least by the time of terminal rupture of the margin, leaving the Taconic seaway as the remnant leading edge of the extended Laurentian margin.

To sum up, Model A model provides several advantages over Model B: (1) it readily explains the different rift-drift transitions in eastern North America vs. Scotland-Ireland, (2) it associates the rift history of the Taconic ribbon terrane across all of eastern Laurentia with a clear geodynamic mechanism (treating the ribbon as the residual leading edge after the breakout of Baltica and then WSP-AA) and (3) it avoids the geodynamic quandary of adjacent ridges. We therefore favor this model and implemented it in our preferred plate reconstruction.

5.2.6. Early Cambrian (541–520 Ma)

Relative to the late Ediacaran, paleomagnetic data in the early Cambrian are sparse, and at 530 Ma the positions of Laurentia and West Gondwana are constrained by singular poles (Fig. 3; Table 1). At 520 Ma, the position of Gondwana is constrained by a grand mean pole (mean of 11 poles; Torsvik et al., 2012), but there are no constraints on either Baltica or Laurentia, although the position of the latter continent can be extrapolated from a 510 Ma grand mean pole (mean of 4 poles; Torsvik et al., 2012). The relative positions of Laurentia and West Gondwana indicated by the two poles at ~530 Ma—the Mont Rigaud and Chatham-Grenville intrusion pole (MR) and the Djebel Boho volcano-poles (DJB)—differ starkly from the relative positions of those two continents at ~550–540 Ma (Fig. 15-k). The relative motion implied between those time-steps could be explained by a clockwise rotation of Laurentia with respect to West Gondwana, but would imply strong differential motion in the West Iapetus (convergence in the east between northeast Laurentia and northwest Amazonia, and divergence in the west). That differential motion, in turn, would require some significant change to the network of plate boundaries separating Laurentia and West Gondwana at 550 Ma, and in particular the nucleation of additional subduction zones somewhere between them. While possible, we consider such solutions geodynamically unlikely because they necessitate that the West Iapetan system suddenly experienced strong compression just a few million years after the opening of Iapetus. An alternative and arguably more parsimonious explanation is that another pair of ‘global motions’ (e.g. TPW) occurred from ~540 to ~530 Ma and from ~530 Ma to ~520 Ma (Fig. 15-k), and that this oscillation has thus far only been recovered from the Laurentia dataset (pole MR). This explanation assumes that the DJB pole from West Gondwana, dated at 536 ± 11 Ma, is too old to have recorded the oscillation and so represents the position of West Gondwana at ~540 Ma. Given the low number of paleomagnetic data, we underscore that this interpretation is speculative, but point out that recent paleomagnetic data from the Borborema province of Brazil (Antonio et al., 2021) are compatible with such a global paleomagnetic oscillation between ~540 and 520 Ma (note that this pole is not included in Section 3 because the study was published after the completion of our model herein). We prefer this solution owing to its geodynamic simplicity, which implies that only minor relative motions occurred between Laurentia and West Gondwana during the early Cambrian, and given the lack of contradictory evidence, we adopt it in our preferred model.

During the earliest Cambrian, WSP-AA evolved as an independent block, migrating across a wide oceanic realm. After its breakup from the hyperthinned/hyper-thinned margin of eastern Laurentia in the latest Ediacaran, WSP-AA drifted rapidly away from Laurentia at a velocity of ~8 cm/yr (Fig. 16a), moving progressively toward the subduction zone along western Amazonia (Fig. 15h-j). In the mid-early Cambrian, WSP-AA accreted to the western margin of Amazonia (Fig. 15k), halting that subduction and causing the Pampean orogeny. By ~515 Ma, subduction had re-initiated along the western margin of Amazonia, and WSP-AA was stitched to the margin of Amazonia by the Famatinian arc.

A final puzzling aspect of the Iapetan realm in early Cambrian time is the rift and drift history of the Precordillera terrane. Paleomagnetic, paleobiologic and sedimentologic data suggest that the Precordillera terrane rifted from the Ouachita embayment of the Southern Segment of Laurentia, shortly (but distinctly) after the rifting of the Taconic ribbon terrane (section 4.5.3 and Appendix B). Equally curiously, while the
Taconic ribbon terrane re-collapsed against the Laurentian margin in Cambrian to Ordovician time, the Precordillera terrane continued to drift away from Laurentia and ultimately amalgamated with the southwest margin of west Gondwana in Ordovician time. Kinematically, a simple way to explain those opposed relative motions is to assume that they were separated by a transform boundary linking the main Iapetus ridge to a rift system inboard of the Precordillera terrane (Fig. 15j-I). Interestingly, in the context of the broader Iapetan kinematics of our preferred model, this transform fault would approximately correspond to the Alabama-Oklahoma transform that Thomas (1991) has proposed for dislodging the Precordillera terrane from Laurentia. Geodynamically, however, the development of this long transform boundary presents an unresolved problem: how did it form? One possibility is that the transform developed suddenly by a ridge jump of one segment of the Iapetus ridge (outboard the Southern Segment) toward the Laurentian margin in the early Cambrian (Thomas and Astini, 1996). However, according to our simple plate model this would imply a jump of some 2500 kms, and a significant thickening of the total plate boundary, both of which seem unlikely (see also Dalziel, 1997; Dalziel and Dewey, 2019). Moreover, widespread magmatism along the Alabama-Oklahoma transform is recognized from 580 to 530 Ma, suggesting that strong displacements along that structural corridor were occurring well before the early Cambrian. On the other hand, the initiation and progressive growth of that transform from some earlier (Ediacaran?) time would necessitate that the Southern Segment of Laurentia was somehow decoupled from the subduction system on the opposite side of the Puncoviscana Ocean (along western Amazonia), before suddenly becoming coupled to it in the early Cambrian. Resolution of this problem will necessitate consideration of the tectonic history further ‘west’ of the West Iapetan domain. Being beyond the realm of our focus herein, we have intentionally left that region (and its boundary with the West Iapetan domain) undefined in our preferred model, but for simplicity we model the transform as suddenly appearing at 540 Ma. We subsequently model a rifting of the Precordillera terrane from southeastern Laurentia until its breakout at 530–520 Ma (Fig. 15k).

5.2.7. Post-script

It is beyond the scope of our focus herein to integrate our preferred model with existing full-plate models of younger (post-early Cambrian) time, but it is nevertheless instructive to draw some simple comparisons with other recent work, namely the models of Domeier (2016) (‘D16’) and Meredith et al. (2020) (‘M21’). D16 focuses on the Iapetan realm and runs from 500 to 410 Ma, whereas M21 is a global full-plate model that spans the Neoproterozoic and Phanerozoic. With respect to a Laurentian-fixed frame of reference, our reconstruction of West Gondwana and Baltica is significantly different from the continental framework of D16 at 500 Ma. Specifically, in D16, the eastern margin of Laurentia faces the northwest margin of Gondwana at 500 Ma, whereas in our model the eastern margin of Laurentia faces the western margin of Gondwana at 520 Ma. Among other distinctions, this gives rise to different orientations of the Iapetus ridge between these models. The position of Baltica is even more distinct: at the end of our model (520 Ma), the northern margin of Baltica faces the northeastern margin of Laurentia, and it is far from northern Gondwana. By contrast, in D16, Baltica is both adjacent to northern Gondwana and oriented ‘upside down’ (relatively rotated by 180°) at 500 Ma, whereas in M21 the orientation of Baltica at 520 Ma is similar to that in our model. These differences largely arise because of the absence of good paleomagnetic data between 550 and 480 Ma in the dataset of Baltica. Consequently, while the position of Baltica at 520 Ma in our model is extrapolated from data at 550 Ma (as also in M21), the ‘upside down’ position at 500 Ma in D16 results from the use of late Cambrian and Early Ordovician paleomagnetic poles that are beyond the timeframe of our model here. A robust paleomagnetic dataset is therefore clearly necessary to understand how Baltica moved during this time interval, which reinforces the need for improved constraints at this junction of time and space. A final notable difference between our preferred model and those of D16 and M21 concerns the manner by which the Precordillera rifts and drifts from southeast Laurentia in the early Cambrian. In both D16 and M21 that terrane occupies the backarc position of a juvenile subduction zone, whereas in our model we show an alternative solution that harks back to that of Thomas and Astini (1996). Both of these solutions are associated with significant problems (see previous paragraph), such that neither can be considered more satisfactory than the other at present. We hope that the key discrepancies between these various models will serve to motivate further work to reconcile them.

6. Conclusions

Here we have interrogated the opening of the Iapetus Ocean by way of an extensive review and synthesis of the paleomagnetic and geologic data that provide constraints on that pivotal tectonic event. Despite significant shortcomings in both of those datasets, we show that a careful analysis of them jointly allows for the formulation of self-consistent and geodynamically plausible tectonic scenarios for Iapetus opening. In our preferred tectonic model, breakup in the Iapetus system was polyphase involving the spalling of multiple marginal terranes from Laurentia and the successive opening of several oceans, including a composite ‘Iapetus Ocean’. The first phase of continental breakup was possibly linked to the emplacement of a plume between eastern Laurentia and western Amazonia at 750–700 Ma, leading to the opening of the Puncoviscana Ocean between those two continents. That was followed by the opening of the eastern branch of the Iapetus Ocean, between Laurentia and Baltica, at ~590 Ma, which may have been instigated by emplacement of the Central Iapetus Magmatic Province. Because rifting along the western Iapetan domain continued until the late Ediacaran-early Cambrian, we invoke another mechanism to explain the opening of the western branch of Iapetus. We propose that rifting/breakup along eastern Laurentia was related to slab pull exerted by a subduction zone dipping beneath western Amazonia. That led to the detachment of marginal terranes from eastern Laurentia and the opening of the western branch of Iapetus at ~550 Ma, following a protracted phase of rifting. We contend that our preferred scenario is the simplest solution given the presently available evidence but throughout this review we underlined key outstanding questions and the attendant uncertainties in our preferred model. We make no pretense that this review is the end of this opening story: we hope our testable scenario and extensive database inspire new insights and focused efforts to resolve the critical questions that remain on the road ahead toward a fuller understanding of Iapetus.

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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Appendix. Supplementary data

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